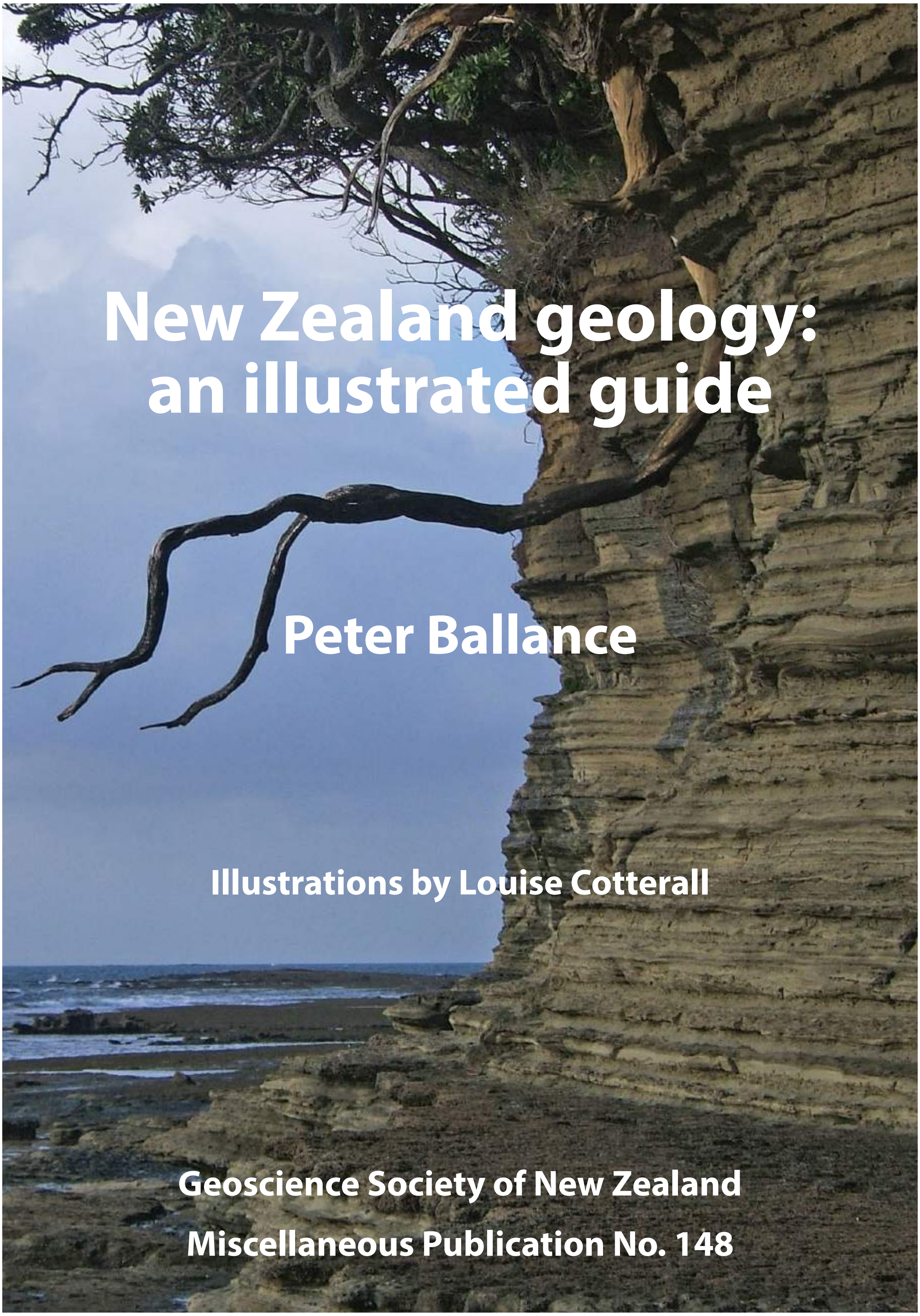

New Zealand geology: an illustrated guide

Peter Ballance

Illustrations by Louise Cotterall

Geoscience Society of New Zealand
Miscellaneous Publication No. 148

Dedication

To Alison, Julian, Joyce and David

who have given us many reasons
to be proud parents

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Cover photo: Waitemata Group strata at Coal Mine Bay, Whangaparaoa Peninsula, Auckland
Photographer Bruce Hayward

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Preface

New Zealand is renowned for its mountains, its volcanoes, its earthquakes and its dramatic scenery. It has these qualities because geologically it is one of the most dynamic places on earth. It is positioned astride a major tectonic plate boundary, where the Pacific and Australian plates collide and grind together. It is the energy of that collision, taking place at speeds of a few centimetres a year, which drives it all.

New Zealand has been in this position for the past 45 million years, and there have been dramatic changes throughout the whole country during that time. However, before that there was another 450 million years of varied geological history. The physical New Zealand we see today is the combined result of all of this history.

The purpose of this eBook is to guide the interested traveller – armchair and actual – through that history, through what we see as we move around the country, or explore from the town we live in. Geological history goes hand in hand with the many different geological processes that have operated at different times. This eBook gives you the opportunity to delve into those processes, which are described in topic boxes.

Senior secondary school students and first-year tertiary students should find the eBook useful for earth science courses. No previous knowledge of geology is assumed, and it is not necessary to travel in order to use the eBook, but I guess that many users will also be travellers, and they can print out pertinent chapters before their journeys. The best way to appreciate New Zealand's superb geology and scenery is, of course, from the road, the walking track and the coast, but if you are unable to travel, the photographs, maps and illustrations may compensate somewhat.

I have tried to present the geological material as far as possible in a "generic" way, so that people can use it to make sense of what they see, wherever they see it. There are some descriptions of specific places, where they are easily accessible and particularly important. On the whole, though, I have concentrated on opening the whole landscape to the traveller's mind. There are other books that describe specific localities in various parts of the country, and these are listed in the Further Reading recommendations at the end of the eBook.

How this book is organised – Topic Boxes and Chapters

To avoid lots of repetition in chapters, general topics such as volcanoes, erosion and limestones are covered in a series of "Topic Boxes". Part 1 introduces the basic facts of Geology that you will need to use the book. Parts 2 and 3 contain chapters each of which covers a part of the country that has a unifying geological theme, and is accompanied by a road map/simple geological map. A simplified geological map of New Zealand and a geological time scale are given in Chapter 1 (Figs 1.0 and 1.2 respectively).

Editors’ Note: This eBook is ideally suited to any desktop monitor or laptop. It can also be used with tablets and kindles (and a magnifying glass), but maps and diagrams are too detailed for the eBook to be viewed on smart phones.

Hyperlinks have been added (coloured blue in the text and also in figures and boxes) to simplify finding figures, boxes or blocks of text referred to in other chapters. Hyperlinks are not used for destinations within the same chapter, and only one hyperlink is used where consecutive destinations are referred to. To activate a hyperlink, click on the coloured part and you will be sent to the section referred to.

When viewing of the hyperlink destination is completed, there will be a method for returning to the previous (source) page. The method depends on the particular version of pdf viewer and might require some experimentation.

Windows operating system:
Using the standard Adobe Reader, returning to the previous page can be achieved either by using a right click drop-down menu and choosing the ‘previous view’ *option, or by using the shortcut* Alt + left arrow *(only the left side Alt key should be used).*

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The standard pdf viewer is Evince Document Reader, in which arrow buttons (< >) on the upper toolbar allow the reader to go back and forth in page history.
For versions 3.4 and older, the arrow buttons are not on the toolbar by default, but can be dragged there (Edit -> Toolbar -> Back).

For other pdf viewers, there are similar functions as buttons or on a right click pop-up menu, always associated with instructions about going back in history.

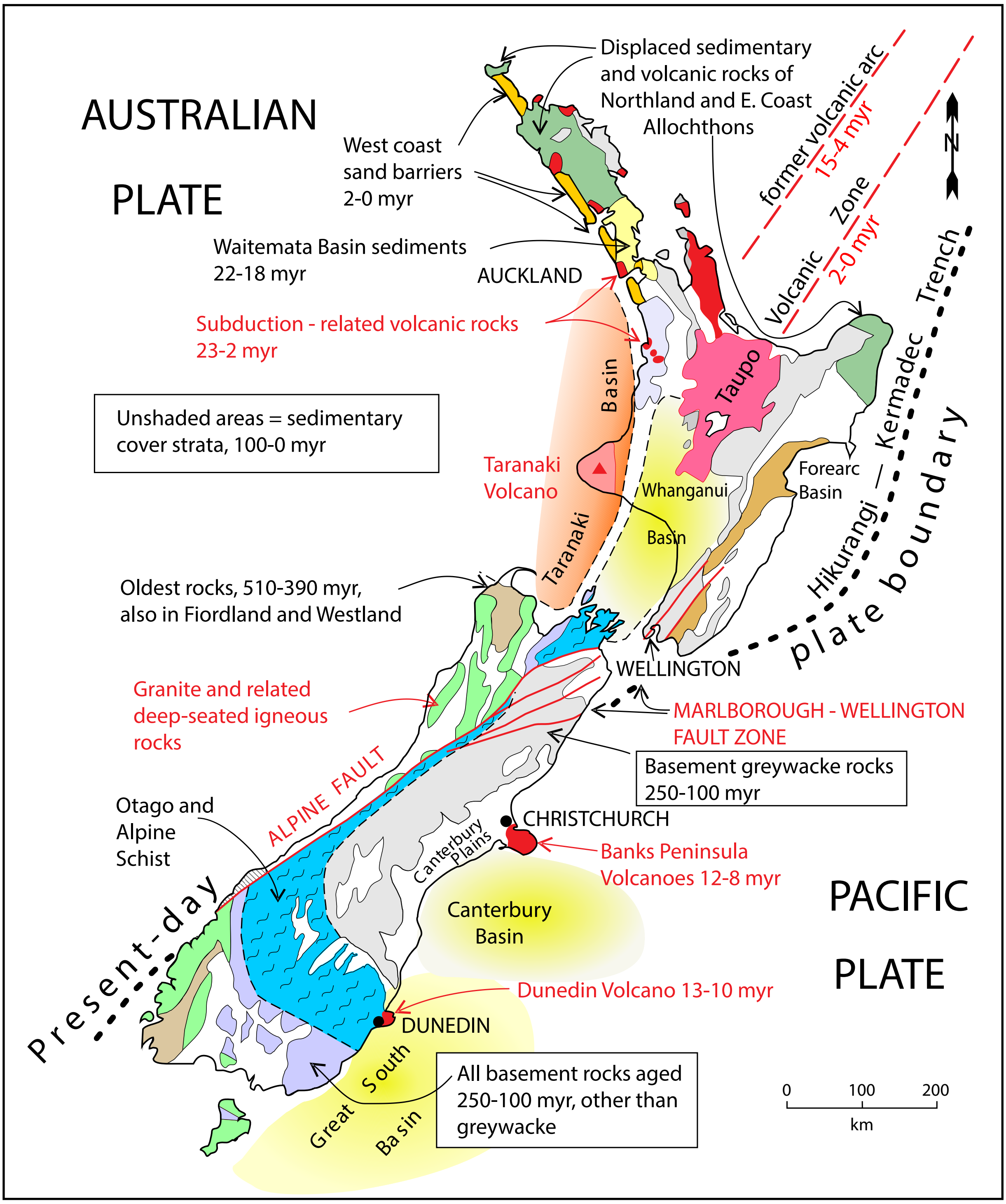


Fig. 1.0. Geological map of New Zealand.

Chapter 1
What is Geology?

Geology is the science of the Earth, and in their study of it geologists gain their information from rocks and landforms. In many ways geology is a science of time and motion – long time, and continuous but generally slow motion. In the past few decades we have learned many ways of measuring geological time accurately, and at the same time we have learned that the Earth's surface is astonishingly mobile. Global positioning systems (GPS) enable us to measure how places are moving, with a precision of plus or minus one or two millimetres per year (mm/yr). As you read this, virtually every place on the surface of the earth is moving. The movement is either upwards, downwards or sideways – mostly the last of these. The only things that are not moving are a handful of volcanic hotspots, such as Tahiti, Hawai'i and Yellowstone, which are rooted very deep down, below the moving tectonic plates.

Motion of the Earth's surface is also estimated from the long-term average effects of tectonic processes. For example, dating of ocean crust at varying distances from a seafloor spreading ridge (Chapter 2) gives a long-term (millions of years) average rate for the production of new crust by the seafloor spreading process, which works out at typically 50–100 mm/yr. It is encouraging that the instantaneous rates now being obtained from GPS usually agree with the long-term estimates.

The rates of motion of the Earth's surface are not exciting. They range from less than 1 mm/yr to about 250 mm/yr. But remember that a rate of 1 mm/yr over a period of a million years adds up to a movement of 1 kilometre (km), and we have many millions of years to play with.

To put time in perspective, the oldest things so far found on Earth are meteorites, which are typically 4.56 billion (4560 million, or 4,560,000,000, or 4.56×10^9, or 4.56 Ga) years old. It is assumed that the Earth is also 4.56 billion years (byr) old, because that seems to be the age of our solar system. The oldest Earth things found so far are individual grains of the mineral zircon, the size of sand grains, dated at 4.4 billion years. The oldest known rocks (a rock is a collection of mineral grains) are just over 4 byr old. The record of the first 500 million years (myr) of the Earth's history seems to have been destroyed, as a result of the high level of tectonic activity that has always characterised our planet. Life forms capable of being fossilised have been in existence for more than 3.5 byr, while abundant life forms with shells and carapaces have been around for about 540 myr.

The oldest-known rocks on the New Zealand mainland are about 510 myr old, while the New Zealand continent (Zealandia is the world's smallest continent, though most of it is submerged) has existed separately for about 80 myr. It has changed mightily in those 80 myr.

Ways of Measuring Time

Most ways of measuring time directly make use of natural clocks. These are radioactive isotopes (called parent isotopes) that decay at constant rates, changing into different isotopes (called daughter isotopes). The amount of radioactive decay can be measured directly, by comparing the quantities of the parent isotope and the daughter isotope in a mineral or rock,

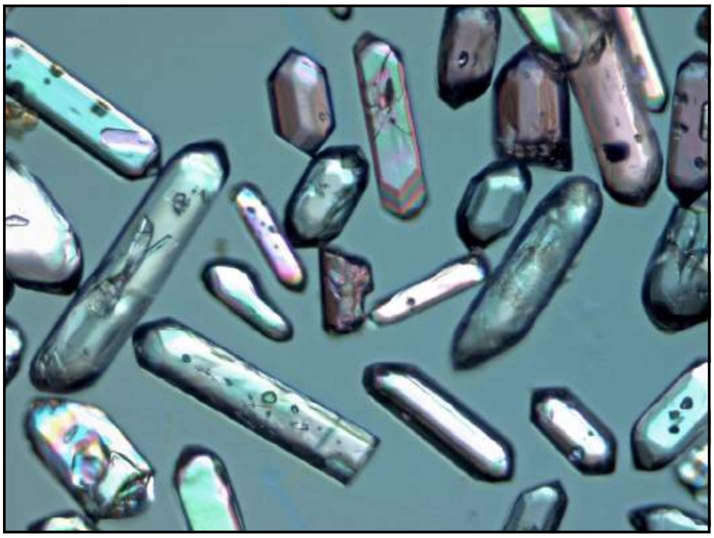


Fig. 1.1. Crystals of zircon, like these, are the oldest things known that were formed on Earth (4.4 billion yrs ago). Dating of zircon crystals is also one of the most used methods for dating many of New Zealand's older rocks. Width of photo 1 mm. Photographer Andy Tulloch.

and converting this into an age in years. Or the amount of decay can be calculated indirectly, by measuring the damage that the radioactive decay does to the surrounding material. Many different radioactive isotopes are used, and the range is increasing all the time. The most commonly employed are uranium, which decays to lead; potassium, which decays to argon; and rubidium, which decays to strontium. For short time periods of 50,000 years or less, radiocarbon can be used. Other dating systems make use of the accumulation of damage by radiation from space. Ages obtained using any of these procedures are called radiometric ages.

Since Ernest Rutherford calculated the first radiometric age early in the twentieth century, dating procedures have become ever more sophisticated, and the analytical instrumentation required correspondingly more expensive. Ages are quoted in years before present, plus or minus the error factor that is inherent in the chemical analysis. Nowadays, this error is generally less than 1%.

In order to reconstruct geological history, we also need to be able to correlate rocks of the same age from place to place. It is not always possible to obtain radiometric 'absolute ages', because they are expensive, and anyway the necessary radioactive elements are not always present. However, sedimentary rocks commonly contain fossils, which enable time correlations to be made around the world – with varying degrees of precision – as life forms have changed systematically over time as a result of evolution. For example, on the geological time scale (Fig. 1.2), all the named divisions of time are based on fossils.

If we are to make sense of radiometric and fossil-based dates, we also need to be able to order geological events – which came first, second, etc. Thus a rock that rests on top of another one, or intrudes into another one, or is made by the alteration of another one, is younger. However, be aware that rocks can be turned upside down by the Earth's movements.

New Zealand as a Continent called Zealandia

New Zealand is not normally thought of as a continent, but as an island in a large ocean. However, when we look below the sea we see that it is indeed a continent – a small one, but a genuine continent. It just happens to be mostly underwater (Fig. 1.3).

ERA	PERIOD	TIME million years	NZ SERIES	NZ STAGE	SUB STAGE	SYMBOL	EPOCH
CENOZOIC	QUATERNARY	0.00117	HAWERA	Haweran	Aranuaian	Quor	HOLOCENE
		0.128			Otiran	Quti	PLEISTOCENE
		0.25			Oturian	Qutu	
		0.34	WHANGANUI	Castlecliffian	Putikian	Wu Wc	
		0.78			Okehuan	Wk	
		1.63		Nukumaruan	Marahauan	Wa Wn	
					Hautawan	Wh	
	TERTIARY NEOGENE	2.4 2.58		Mangapanian		Wm	PLIOCENE
		3.00		Waipipian		Wp	
		3.70		Opoitian		Wo	
		5.33	TARANAKI	Kapitean		Tk	MIOCENE
		7.2		Tongaporutuan		Tt	
		11.04	SOUTHLAND	Waiauan		Sw	
				Lillburnian		Sl	
				Clifdenian		Sc	
		15.9	PAREORA	Altonian		Pl	
		21.7		Otaian		Po	
	PALEOGENE	23.03	LANDON	Waitakian		Lw	OLIGOCENE
				Duntroonian		Ld	
				Whaingaroan		Lwh	
		34.6	ARNOLD	Runangan		Ar	EOCENE
				Kaiatan		Ak	
				Bortonian		Ab	
		42.6	DANNEVIRKE	Porangan		Dp	
				Heretaungan		Dh	
				Mangaorapan		Dm	
				Waipawan		Dw	PALEOCENE
				Teurian		Dt	
MESOZOIC	CRETACEOUS	66.0	MATA	Haumurian		Mh	LATE
				Piripauan		Mp	
		86.5	RAUKUMARA	Teratan		Rt	
				Mangaotanean		Rm	
				Arowhanan		Ra	
		95.2	CLARENCE	Ngaterian		Cn	EARLY
				Motuan		Cm	
				Urutawan		Cu	
		108.4	TAITAI	Korangan		Uk	
	JURASSIC	145.0	OTEKE	Puaroan	Waikatoan Mangaoran	Op	LATE
		150.0	KAWHIA	Ohauan		Ko	MIDDLE
				Heterian		Kh	
				Temaikan		Kt	
		176.0	HERANGI	Ururoan		Hu	EARLY
				Aratauran		Ha	
	TRIASSIC	201.3	BALFOUR	Otapirian		Bo	LATE
				Warepan		Bw	
				Otamitan		Bm	
				Oretian		Br	
		227.5	GORE	Kaihikuan		Gk	MIDDLE
				Etalian		Ge	
				Malakovian		Gm	EARLY
				Nelsonian		Gn	
PALEOZOIC	PERMIAN	251.5	D'URVILLE	Makarewan		YDm	LATE
				Waiitian		YDw	MIDDLE
				Puruhauan		YDp	
		259.8	APARIMA	Flettian		YAf	EARLY
				Barrettian		YAr	
				Mangapirian		YAm	
				Telfordian		YAt	
		288.5		pre-Telfordian			
	CARBONIFEROUS	298.9					
	DEVONIAN	358.9					
	SILURIAN	419.2					
	ORDOVICIAN	443.8					
	CAMBRIAN	485.4					
		541	PRECAMBRIAN				

Fig. 1.2. Simplified New Zealand geological time scale.

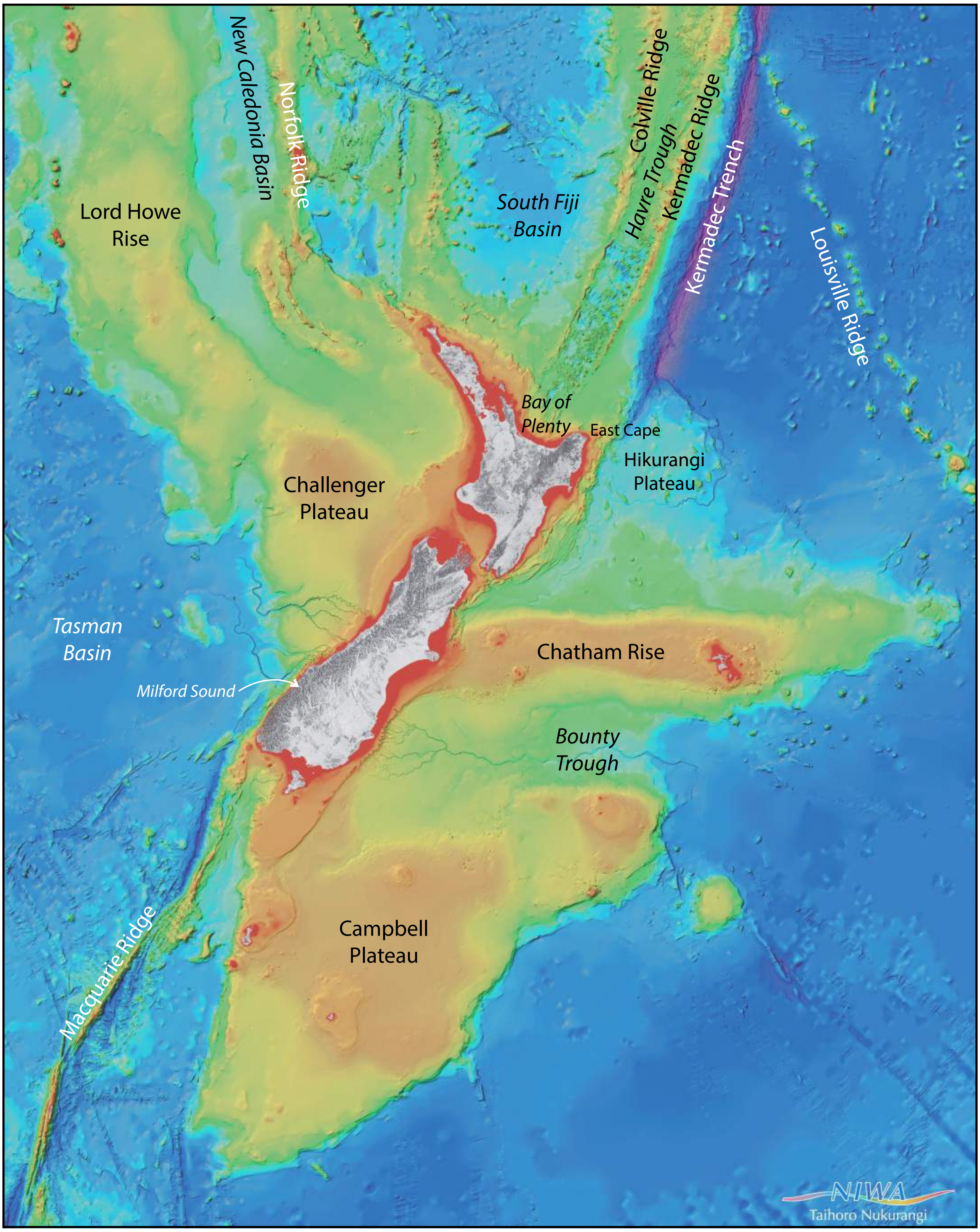


Fig. 1.3. Undersea topography of the New Zealand region. Image courtesy of and copyright to NIWA. Bibliographic reference:
Mitchell, J.S.; Mackay, K.A.; Neil, H.L.; Mackay, E.J.; Notman, P. 2012. Undersea New Zealand, 1:5,000,000, NIWA Chart, Miscellaneous Series No. 92. Published by the National Institute of Water and Atmospheric Research Ltd.

Continents are rafts of lighter material floating on heavier rocks underneath. They are typically about 30 km thick and have sufficient freeboard, relative to the rocks they are floating in, to keep their upper layers above sea level (which is, of course, a completely independent phenomenon). Zealandia is mostly below sea level because it is thinner than average: 20–25 km instead of 30 km-plus. There is a geological explanation for this, as we shall see in Chapter 4.

The lighter rocks of continents contain more silica than the heavier rocks underneath. They are about 2.7 times as heavy as water, and represent the slag or dross that has been sweated out of the heavier rocks through continuing long-term melting and volcanic action. The heavier rocks contain more iron and magnesium, and are about three times as heavy as water. This process of separation of lighter and heavier rocks continues, and as a result continents are getting bigger. Exactly how it works is determined by plate tectonics, which is the subject of the next chapter. Just be aware at this stage that, like everything else on the surface of the earth, continents are moving around, being broken up and being added to.

Generally speaking, once created, continental crust cannot be destroyed. Unlike the heavier rocks underneath, which can be – and are – taken down into the earth's interior (subducted), the lighter continental rocks are usually too buoyant to be subducted. A useful analogy is trying to flush a cork down the toilet. They can, however, be taken apart and redistributed, as is happening today in the South Island of New Zealand. Thus, the continents sail majestically and serenely around the globe, pushed by the 'winds' of moving tectonic plates, while most of the geological action takes place around their margins, such as along the Pacific Ring of Fire. That does not mean that each continent is an immutable object; as noted above, it can be pulled apart, or have pieces added to it.

Chapter 2

Plate Tectonics

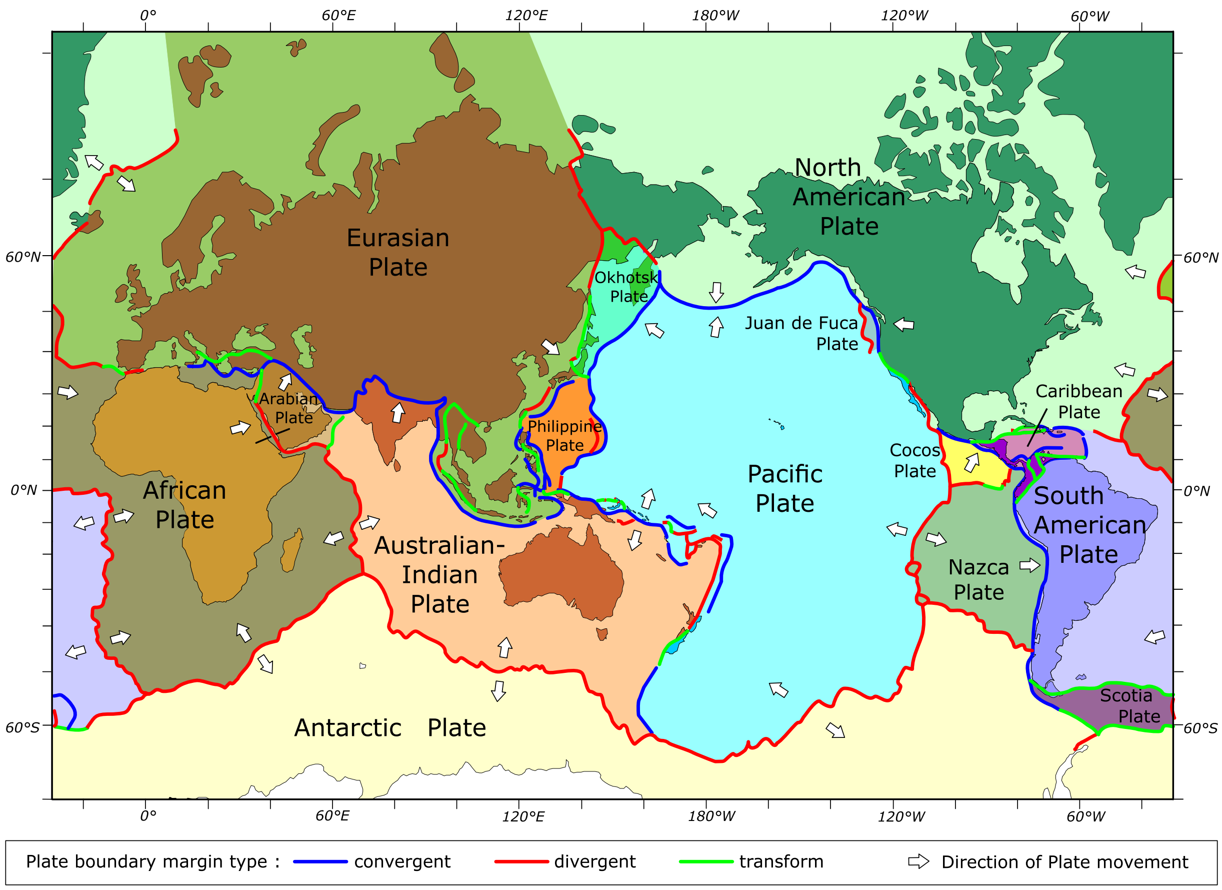
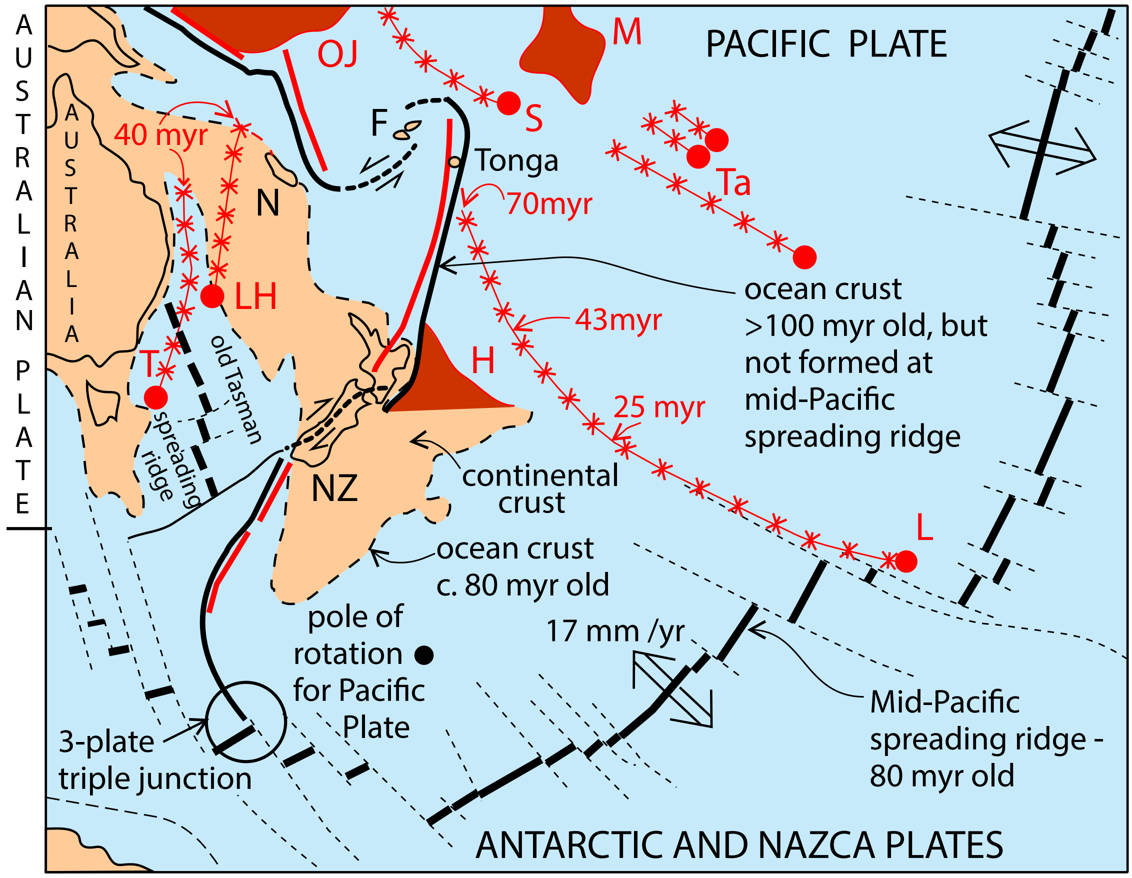


Fig. 2.1. Global tectonic plates.

Tectonics refers to processes that move and deform rocks. A plate is a rigid object that is much wider than it is thick. By coincidence, the Earth's tectonic plates have similar ratios of thickness to width as do dinner plates. The main difference is that a tectonic plate is part of the surface of a sphere, so it isn't flat. A tectonic plate is typically 100 km thick and thousands of kilometres wide. It is rigid and is also elastic, which means that it can be bent and will recover from bending, like an aluminium oven tray, but unlike a dinner plate.

The name for the top 100 km of rigid rock making up the tectonic plates is the lithosphere (the prefix litho- comes from the Greek word *lithos*, meaning 'rock'). If 100 km sounds like a very thick plate, remember that the earth has a diameter of 12,750 km and a circumference of 40,000 km at the Equator. The essence of the plate tectonic process is that the lithosphere is broken up into a dozen or so major plates (and a number of smaller ones) that are being moved around, a few centimetres a year, by the slow overturn of convection currents in the more plastic rock underneath.

Box 2.1. Plate tectonics of the Southwest Pacific.



= deep volcanic hotspot, and chain of volcanic seamounts formed as the plate moves across the hotspot. Ages of seamounts in millions of years. Note that the Australian Plate is moving northwards, away from the Australia — Antarctica spreading ridge.

T = Tasmania, LH = Lord Howe, S = Samoa, Ta = Tahiti,
L = Louisville hotspots. Not all hotspots are shown.

N = New Caledonia, F = Fiji, NZ = New Zealand

= Oceanic plateau — initial outpouring of basaltic lava from a hotspot.
OJ = Ontong Java, M = Manihiki, H = Hikurangi

Volcanic arc
Subduction trench } = convergent plate boundary

transform (sideways-moving) plate boundary

mid-ocean spreading ridge (divergent) plate boundary. Age of ocean crust increases away from spreading ridges. Ridges are often offset across oceanic fracture zones.

The Pacific Plate is presently rotating anticlockwise around its pole of rotation, relative to the Australian Plate, at 1 degree per million years.

Cause of the Convection Currents

Locally concentrated heat causes convection currents – whether in the air, in water in a saucepan, or in the Earth – because it has to be dissipated. The Earth is a giant heat engine. The source of this heat is radioactive decay of some elements in the rocks from which it is composed. The commonest of these radioactive element by far is potassium; it is not highly radioactive, but there is a lot of it. It also gives us the potassium–argon radiometric clock, as noted in Chapter 1. If that heat were not dissipated – i.e. transferred to the Earth's surface and radiated into space – the Earth would melt.

The way the Earth's heat is dissipated is through the formation of convection currents in the mantle (Box 2.2). The mantle lies between the liquid nickel-iron core and the rigid lithosphere. It is about 2800 km thick, and is hot enough to be plastic but not molten. The convection cells that form here vary in size. In cross section they typically measure between 1000 km and 9000 km from the ascending to the descending limb. Lengthwise, it is more difficult to give a figure, because the ascending limbs all link together around the world, as we will see next. There is still debate as to whether the convection cells involve the entire mantle, or only the upper part of it.

The speed of convective overturn is a few centimetres per year. If that seems to be a trifling amount, remember that 10 cm (100 mm) per year equates to 100 km in 1 myr, and that the lifetime of convection cells is measured in hundreds of millions of years (although they do reorganise themselves from time to time). When you factor that in to the speeds noted, you begin to understand how the earth's geography can change radically in a couple of hundred million years.

Effects of Convection – Plates and Plate Boundaries

If the lithosphere is broken into plates that are moving around at a few centimetres a year, driven by the underlying convection cells, it follows that the plates must interact with each other. This interaction gives rise to the observation that much of the world's geological activity is concentrated around the plate boundaries.

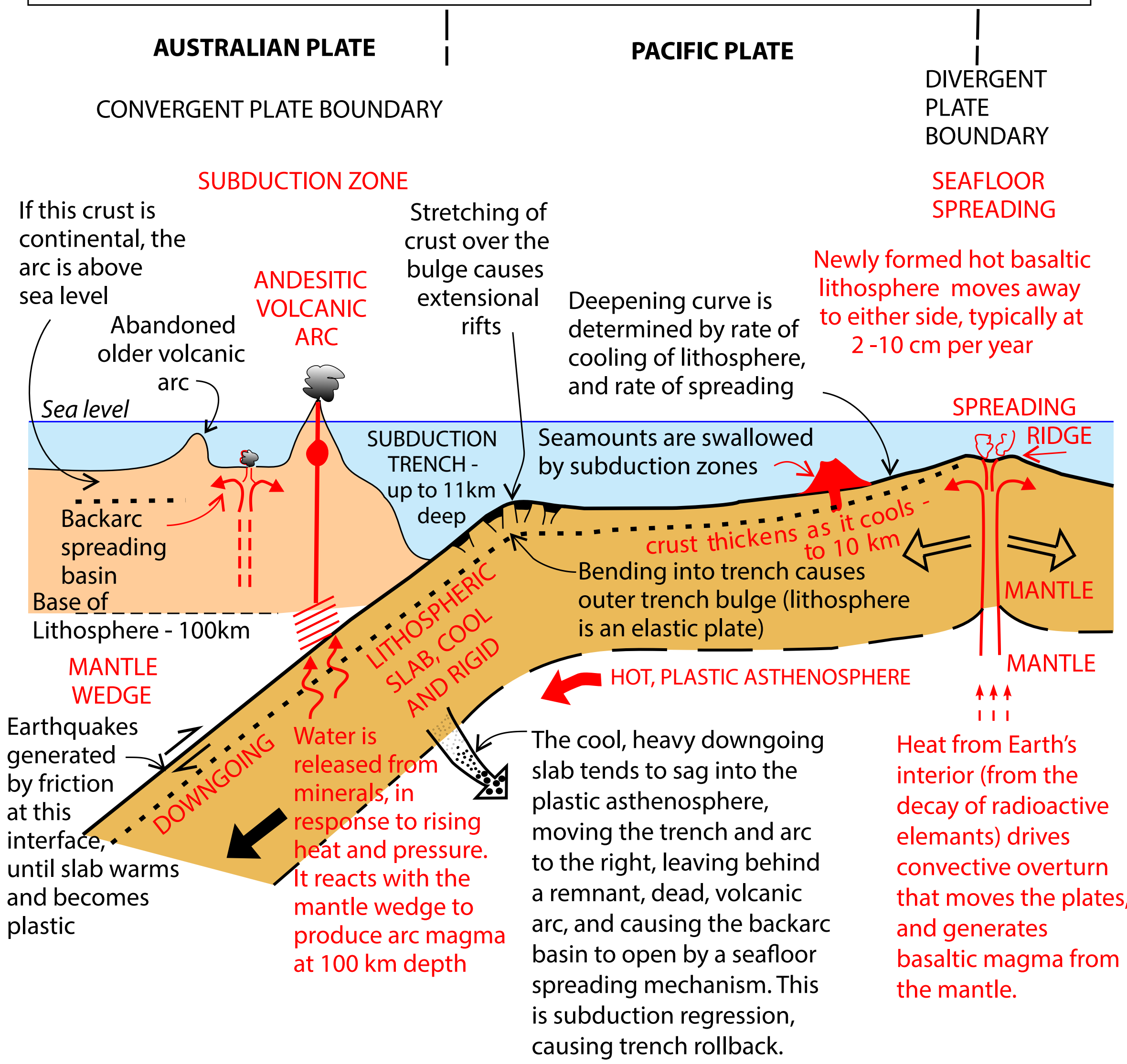
A rising convection limb of hot mantle rock carries surplus heat to the surface. Hot rock is more expanded and therefore less dense than cold rock, and since all rock is floating on denser rocks underneath, hot rock floats higher than cold. If the rising limb is located underneath continental crust, the continent stands high. If it is located underneath oceanic crust, this also stands high but generally remains below sea-level, forming mid-ocean ridges or mountain ranges.

At the surface, the rising convection limb splits into two, one half going to the right, the other to the left (Box 2.2). In the case of the Earth's mantle, the gradual reduction in confining pressure as the hot rock rises allows it to begin to melt, forming magma. Magma is molten rock that contains dissolved gases, while lava is magma that has been extruded at the Earth's surface and has lost much of its dissolved gas.

Magma is even less dense than hot rock, and much more mobile, so it makes its way upwards rapidly. Some reaches the surface and is extruded as lava. Some does not reach the surface, but intrudes into surrounding rocks as molten rock bodies of various shapes. The chemical composition of this magma is basaltic (Boxes 6.9 A, B), which is the same composition as the Earth's mantle.

Box 2.2. Diagrammatic cross-section through the Pacific Plate and part of the Australian Plate (not to scale).

REMEMBER - both plates are moving. Any combination of movement directions and rates is possible.



Rate of subduction = rate of spreading, plus or minus the movement of the overriding plate.
Maximum measured rate = 23 cm/yr, near Tonga.

Divergent (Spreading) Plate Boundaries

The facts above explain the typical form of a mid-ocean spreading ridge. It is symmetrical in profile, and its highest point is located over the central part of the rising convection limb. It normally has a central rift valley, where the division of the rising limb into two convection cells is pulling the ridge apart, and the rift is the site of prolific basaltic volcanic activity. There is high heat flow, as manifested in the geothermal smoker systems that are so spectacular in underwater films. There is also a lot of earthquake activity.

The symmetrical profile of a mid-ocean ridge occurs because the rocks are moving away from the central rift valley at the same speed on either side – a few centimetres per year. The rocks moving away are replaced by new lava, both extruded and intruded, and the slope of the flank is determined by how rapidly the hot rocks cool and sink. Rock is a very poor conductor of heat, so cooling is a slow process, and it takes place at the same rate regardless of how fast the convection cell is moving. Thus, a fast-moving double cell will form a ridge with gentle slopes, while a slower-moving cell will have steeper slopes. The actual depth of ocean floor moving away from the ridge-top, relative to sea-level, is directly related to its age, i.e. the cooling time that has elapsed since it was at the ridge-top.

As shown in Fig. 2.1, the mid-ocean spreading ridges form a connected system about 40,000 km in length. A more detailed view (Box 2.1) shows that they are not continuous, but are broken into segments by transverse fractures called transform fracture zones or faults. Because of the sideways motion of the newly formed oceanic lithosphere and crust, these are sideways-moving faults. They form a subclass of the third kind of plate boundary, the sideways-moving boundary (see below). However, they have a unique feature. If you think of the segments of moving lithosphere between the transform faults as conveyor belts, then in the reach between the offset ends of the spreading ridge one side of the fault is moving sideways compared to the other. The 'conveyor belt' also juxtaposes lithosphere on one side of the spreading ridge (and therefore on one plate) against lithosphere on the other side (on another plate), so it is a genuine sideways-moving (transform) plate boundary. But, as soon as the transform leaves that central reach and is outside the spreading area, both sides of it are moving in the same direction and at the same speed. So, although we can still see it in the seafloor morphology, it is no longer an active fault and lies within one plate, no longer forming a plate boundary. Refer again to Box 2.1.

The story doesn't end, even there. The lithosphere on opposite sides of the inactive transform fault, while part of the same plate, is of different ages and has therefore cooled to different extents, which affects the depth of the sea floor. In part, that is how we can still see the fault. This difference persists right until the lithosphere is eventually subducted, at which point the change in temperature and density across the transform fault can affect the subduction process (see below).

The spreading, or divergent, boundary is thus one kind of plate boundary, where new lithosphere is continually being created. The lithosphere is initially thin, while the rocks are still very hot, but it thickens to around 100 km during cooling. A cold oceanic crust, about 10 km thick, is differentiated from the underlying lithosphere. As we will see next, the long cooling time of the lithosphere has a bearing on the processes of subduction at the other common kind of plate boundary, the convergent one.

Convergent (Subducting) Plate Boundaries

If new lithosphere is continually being created along 40,000 km of divergent plate boundaries, it follows that there has to be an opposite process, destruction of lithosphere – unless the Earth is expanding. There is no evidence for an expanding Earth, but there is plenty of evidence for the destructive process, whereby lithosphere bends at a convergent plate boundary and passes down into the Earth's interior along a so-called subduction zone, the descending limb of a mantle convection cell. Probably the most impressive piece of evidence for the destruction of lithosphere is the fact that there is no oceanic crust (and therefore lithosphere) at the Earth's surface older than about 150 myr, despite the fact that, as far as we can tell, the plate tectonic process has been taking place for most of the earth's 4.5 byr history.

In light of the comments in Chapter 1 about the 4 byr age of some rocks in the continents, we need to qualify the 150 myr age of the oldest oceanic lithosphere. As noted in Chapter 1, continents – which occupy about 30% of the earth's surface – cannot be subducted. This fact has a big influence on the pattern of the tectonic plates, and accounts for the fact that many subduction zones are located along the edges of continents, so that heavy oceanic crust and lithosphere are subducted beneath the continent. Thus oceanic lithosphere older than 150 myr may in fact be hiding underneath continents, which are protecting it from subduction, but we have no way of accessing or dating such lithosphere.

Like divergent plate boundaries, convergent plate boundaries have a characteristic set of features (Box 2.2). Working from the uphill side, so to speak, the bending of the elastic lithosphere into the subduction zone first causes an upwards bulge of the sea floor. Returning to the analogy of the aluminium oven tray at the beginning of this chapter, if you hold one end and bend the other, it bulges in the middle. As rigid oceanic crust rises to pass over the bulge, it is stretched and breaks into rift valleys separated by ridges. The main bend of the plate creates a long, narrow, deep-sea trench, and the bottom of the trench is where oceanic lithosphere of the down-going plate passes underneath lithosphere of the overriding plate, i.e. this is the actual plate boundary. The rift valleys play an important role in carrying material from the inner trench slope down the subduction zone.

If one of the plates is carrying a continent, which cannot be subducted, the plate that subducts will always be the other one, comprising heavier oceanic lithosphere. Thus the trench runs alongside the coast of the continent. This is the situation along most of the west coast of Central and South America, and along the east side of the North Island of New Zealand. If both converging plates are of oceanic crust, either one can subduct. In this case, the age, temperature and density of the lithosphere would be determining factors. If subduction brings two continents together, and neither can subduct, then a major readjustment of subduction systems has to take place.

One of the beauties of plate tectonics is its variety and versatility. For example, looking at two-dimensional diagrams such as Box 2.2, it is easy to assume that two converging plates always approach each other head-on, at 180°, along the same line. That is actually seldom the case. It is more common for the two plates to be moving along quite different lines, and all that is required for subduction to take place is that there is an element of convergence when all the motion factors are summed together. Indeed, you could have a subducting plate boundary where the two plates were moving in exactly the same direction, provided that the following plate was moving faster than the leading plate.

In New Zealand's case, we straddle the Pacific–Australian plate boundary, which runs from north-northeast to south-southwest through the New Zealand continent (Boxes 2.3, 2.4). However, neither plate is moving along a WNW–ESE line at right angles to the boundary. The Australian Plate is moving to the north, while the Pacific Plate is rotating anticlockwise around a point situated to the south of New Zealand, such that it is moving due west in the vicinity of Tonga but is moving southwestward, almost exactly parallel to the plate boundary, through central and southern New Zealand.

This versatility in plate movement scenarios leads to all kinds of interesting variations in subduction zones. There isn't room to explore all of them here, but we will look at those that affect, or have affected, New Zealand.

Beyond the deep-sea trench, the down-going plate moves into the Earth's interior. The angle of subduction varies depending on how hot or cold (i.e. how young or old) the down-going lithosphere is, and how old the subduction zone is. With time, the relatively cold and heavy lithosphere tends to sag into the plastic rocks beneath and to get steeper. At the same time, the sagging causes the position of the main bend in the plate – the trench – to move backwards into the down-going plate, a phenomenon called trench rollback.

Ultimately, the subducting slab – which is the descending limb of the mantle convection cell – steepens to vertical. As rock is such a poor conductor of heat, it takes an age for the descending limb to warm up to the temperature of the surrounding rocks. Using a technique called seismic tomography, we can track the vertical limbs of cooler rock down to depths of 1000 km or more.

One factor here is constant, however, which is that the deeper the lithosphere penetrates, the more the confining pressure increases. At around 100 km depth, regardless of the angle of subduction, the higher pressure causes some changes in the mineral composition of the lithosphere – the chemistry of the rocks stays the same, but some of the minerals change into new ones that are stable at the higher temperatures and pressures. In the process, water is released, and this water reacts with the rocks in the subducting lithosphere and in the base of the overriding lithosphere that is in contact with the down-going slab, forming molten magma.

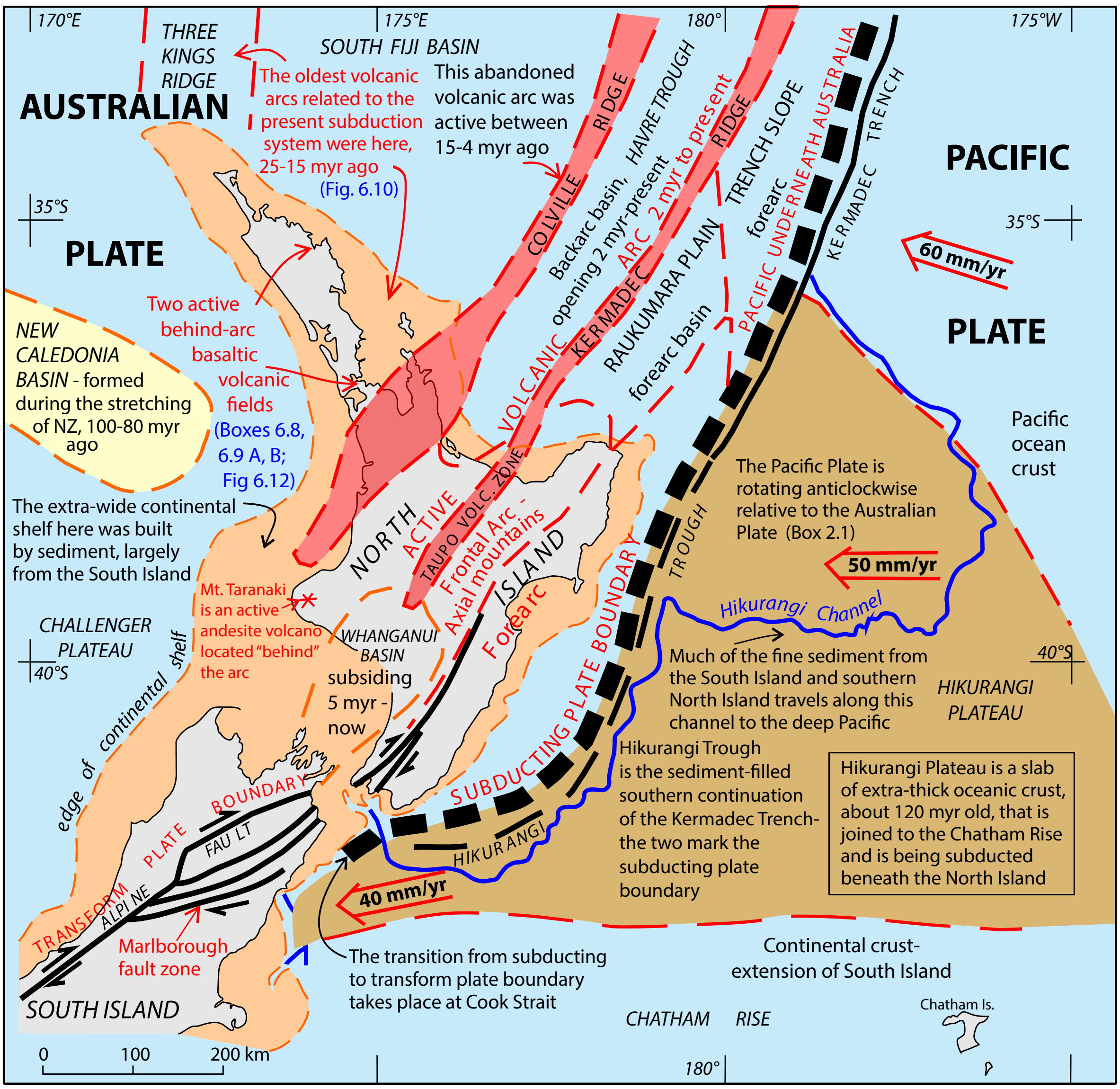
As we discussed above with mid-ocean spreading ridges, magma is buoyant and makes its way to the surface. In the case of the subduction zone, magma forms along a line marking the 100 km depth contour of the surface of the subducting slab of lithosphere. Thus it forms a line of volcanoes known as a volcanic arc or, more correctly, a magmatic arc, because not all of the magma reaches the surface. The arc (so called because on a map it is usually curved rather than straight) is the second major feature of a subduction zone after the trench.

The Pacific Ring of Fire (Fig 2.2) is the best-known collection of magmatic arcs. Most of the volcanoes along these arcs are of the Mt Ruapehu type: large edifices of andesitic chemical composition (containing a higher content of silica than the basalts of the mid-ocean spreading ridges) spaced at intervals of 20–100 km (Box 9.2 A). Andesite seems to be the typical magma generated at 100 km depth. Many of the volcanoes are entirely under water; some reach the surface

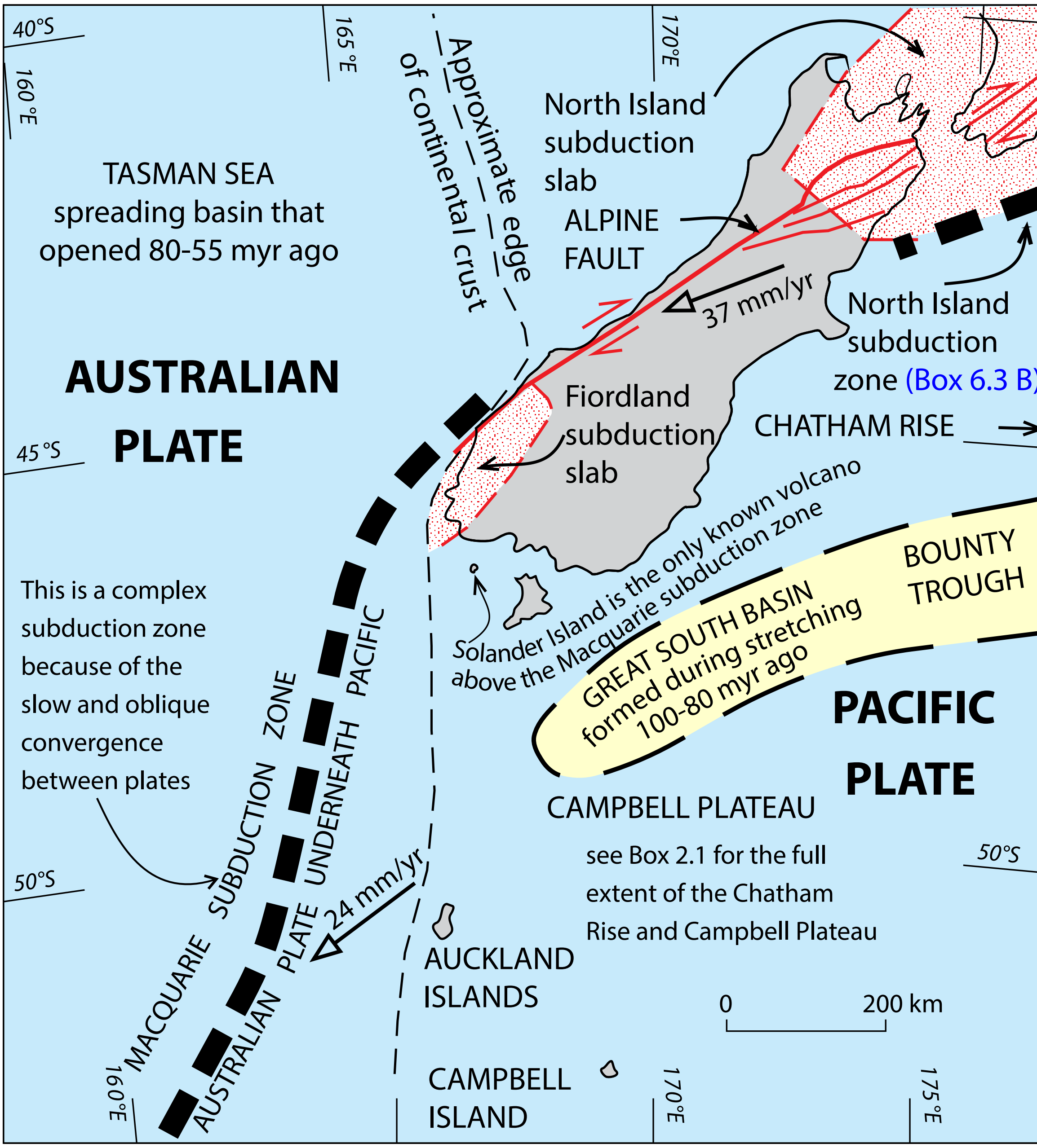
Box 2.3. North Island plate tectonic situation.

The broad picture:

1. Pacific Plate is rotating anticlockwise relative to the Australian Plate, at rates and directions shown
2. Subduction of the Pacific Plate occurs from Cook Strait northwards, causing the active volcanic arc and associated backarc, forearc and trench features (Boxes 2.1, 12.1 A)
3. Older volcanic arcs related to the plate boundary are shown (25-4 myr)
4. The active volcanic arc stops in central North Island. Its continuation southwards is the Whanganui Basin (Box 10.1)
5. Transition to the transform plate boundary of the South Island takes place at Cook Strait. Note the overlap between subduction and transform.



Box 2.4. South Island plate tectonic situation.



The movement of the Pacific Plate, relative to the Australian Plate, is shown by the arrows. It is described geometrically as an anticlockwise rotation, around a point which is located southeast of the map, at 60ºS and longitude 180º. The rate of rotation is 1º per million years. See Box 2.1 for the North Island arrows. Note that this movement is not the absolute motion of the plate (i.e. total movement relative to lines of latitude and longitude) – see Box 2.1 for comment on that. The Alpine Fault is an oblique, compressive, transform plate boundary – the compression at the fault indicated by the 37 mm/yr arrow forces continental crust of the Pacific Plate up and over continental crust of the Australian Plate, forming the Southern Alps (Boxes 19.1 A, B; 19.3).

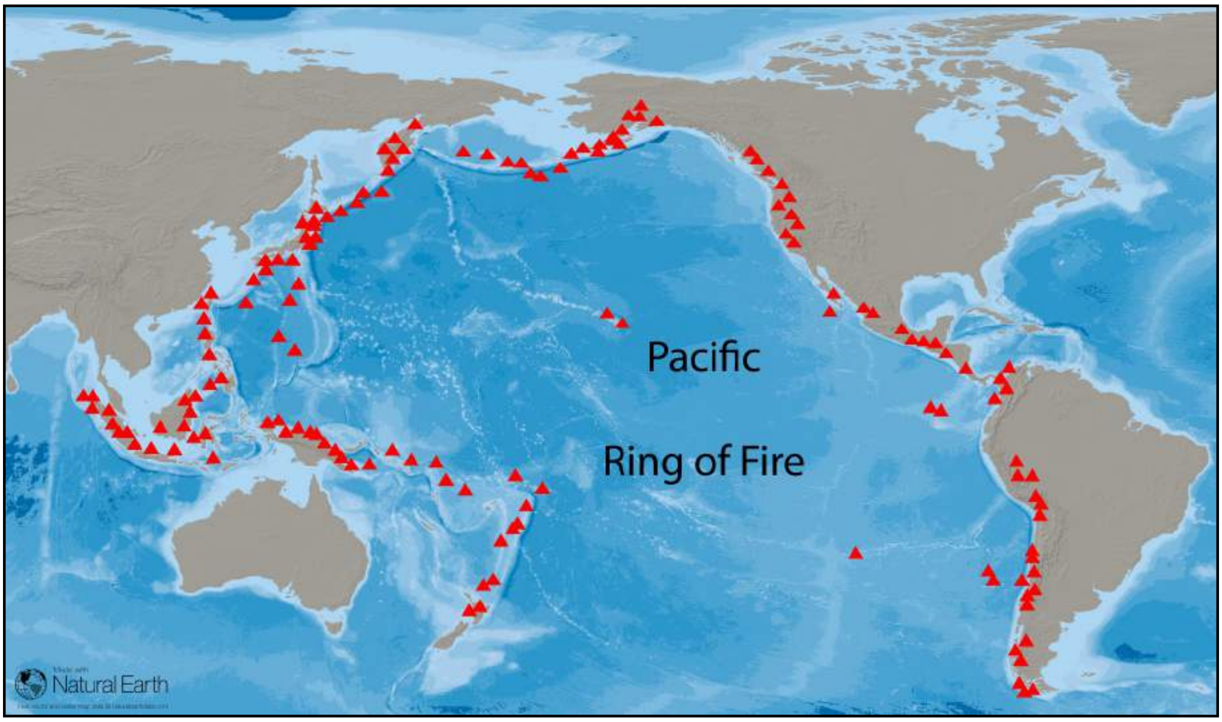


Fig. 2.2. Pacific Ring of Fire. Red triangles are active volcanoes.

to form islands, such as New Zealand's Kermadec Island group; and some perch on continental crust where the subduction zone is beneath a continent, such as Mts Ruapehu, Ngauruhoe and Tongariro in the centre of the North Island. The distance between the trench and the arc varies from 100 km to 300 km, depending on the slope of the subduction zone.

Not all subduction volcanoes are of the Mt Ruapehu type and chemical composition. In the process of making its way upwards, magma interacts with other rocks and can change its chemical composition, which in turn affects the kind of eruption mechanism that takes place. Thus, there are circumstances in which rising andesitic magma melts continental crust, to produce magmas with the same silica-rich composition as the crust. As we will explore in more detail later in Chapter 9, Lake Taupo in the middle of the North Island is the world's largest, most violent and most productive volcano. It erupts every few thousand years, and its lava is rhyolitic in composition (Box 9.3 A), with a silica content of around 75%.

Swallowing seamounts

One remarkable feature of subduction zones is their ability to swallow seamounts. Look at Fig. 1.3 and you will notice that the deep-sea floor is dotted with seamounts. Compare them for size with the big central North Island volcanoes, and you will appreciate that they are very large indeed. They are all volcanoes, mostly of basaltic composition, and come in two kinds. The first kind are scattered randomly across the sea floor, and it is not known what governs their existence. They occur on all the tectonic plates. The second kind is the linear seamount chain, seven of which can be seen in Box 2.1. Two of these chains are near the map's northwest edge, aligned south to north, and record the northward movement of the Australian Plate over two volcanic hotspots. The longest is the Louisville Seamount Chain, which actually extends much further east, to a hotspot nearly at the mid-Pacific spreading ridge.

Seamount chains form as tectonic plates pass across very deep-seated volcanic hotspots. Hotspots were mentioned right at the start of Chapter 1, where it was noted that they are the only fixed entities on earth, being situated below the mantle convection cells. So an active volcano sits above the hotspot for a time, until it is carried away from the hotspot by the movement of the overlying plate and dies, and a new volcano forms (Box 2.1). This process continues, forming a seamount chain with a regular age progression. Hawai'i is the best-known hotspot volcano, and its chain of seamounts stretches far away into the northwest corner of the Pacific Ocean, where it is being subducted at the Kamchatka Trench. The seamounts at that point are about 70 myr old.

Look again at Box 2.1, and you will see that the Louisville Seamount Chain is also being subducted, at the Tonga Trench. The next seamount that will enter this trench is aged 70 myr. Because the chain is oriented northwest to southeast, and because the Pacific Plate here is moving due west into the trench, the point of entry of seamounts into the trench is moving southwards.

It is evident around the world that subduction zones swallow seamounts quite effortlessly. The seamounts do, however, have to bulldoze their way into the inner trench slope. Along the way, they do quite a bit of damage, they leave scars, and they also leave bits of themselves scattered around on the trench slope, ready to be picked up by marine geologists using dredges. In Fig. 1.3 a large seamount subduction scar can be seen on the continental slope of North Island offshore from Ruatoria, just south of East Cape. The seamount has been pinpointed using its magnetic signature, and is located under the continental shelf.

Going back to the Louisville Seamount Chain, a series of large, closely spaced seamounts, each one entering the trench/subduction zone a little south of the previous one, would be expected to leave a cumulative trail of damage. And this has indeed happened. The trench immediately north of the present intersection is particularly deep – it is called Horizon Deep, is nearly 11 km in depth, and is the second-deepest place in the ocean. The inner trench slope is very steep here, and a dredge taken from the lower trench slope contained a variety of basalt lavas and oceanic ooze sediments aged around 80 myr that had been scraped off an earlier seamount. The trench here has been deepened and steepened by 'tectonic erosion'.

Just in passing, long seamount chains like the Hawai'i and Louisville hold useful tectonic information. Any change in direction of plate movement causes a bend in the chain, which can be dated from the ages of the seamounts. The Hawai'ian and Louisville chains are both on the Pacific Plate, and therefore should both record any changes in direction of movement that have occurred on that plate. And they do. Both have a sharp bend at 45 myr, which coincides with the inception of our current plate boundary; it is possible that the collision of India with Asia (which has given us, among other things, the Himalayas and the Tibetan Plateau) shook up the world's plates enough to cause this change in direction. There is another small bend at 25 myr, which coincides with the kickstarting of New Zealand's Pacific–Australian plate boundary (Chapter 3).

Subduction earthquakes

The third dominant feature of subduction zones, after deep-sea trenches and magmatic arcs, are subduction-zone earthquakes. These are the world's biggest earthquakes and arise because of friction between the two plates in the subduction zone. The energy of plate movement is stored as frictional stress until it is suddenly released, causing an earthquake – a jolt of movement on the subduction interface and/or along a surface fault. At any one area of a subduction zone this happens every few hundred years. The situation continues until the down-going plate warms sufficiently to become plastic, and then earthquakes cease. The vertical depth at which earthquakes cease varies between 100 km and 700 km, and depends on how hot the lithospheric slab was to start with, how obliquely the plate is subducting (an oblique trajectory increases the travel distance to a given depth, giving the slab more time to warm up) and how fast it is subducting. The deepest subduction-zone earthquakes, 700 km beneath Tonga and Fiji, occur where old, cold, dense lithosphere is subducting directly and rapidly (at around 20 cm per year).

New Zealand's earthquake pattern is shown in Box 13.2 A–C. Deep earthquakes reflect the way that the boundary between the Pacific and Australian plates changes character as it passes through the country. As the subducting Pacific Plate encounters the continental crust of North Island, a moderate speed of subduction and a slightly oblique trajectory carry the slab to around 300 km depth before earthquakes cease. Going southwards to Cook Strait and northern South Island, the speed of subduction gradually slows while obliquity increases, and so the deepest earthquakes here are shallower than those further north along the plate boundary. The final manifestation of subduction earthquakes, in northern South Island, is at 100 km depth. At that point, the cross-over from an obliquely subducting boundary to a sideways-moving one (see below) is complete. A new subducting slab appears at the southwestern corner of the South Island. Here, by contrast, Australian Plate lithosphere is subducting beneath Pacific Plate lithosphere, but very obliquely and in a complex fashion.

Sideways-moving (Transcurrent or Transform) Plate Boundaries

Given earlier comments about the huge variability in the configuration of plate interactions, it should come as no surprise that there are places where two plates simply slide past each other. It is, however, rare for the two plates to be moving exactly parallel to one another. An example of where this does happen is in the northern South Island, at the Wairau Fault in Marlborough (the northernmost part of the Alpine Fault; Chapter 15). Parts of the San Andreas Fault in California, at the boundary of the Pacific and North American plates, are also purely sideways-moving, and as noted above, transform faults along the spreading mid-ocean ridges are a special case.

More commonly, in addition to the sideways movement there is a component of either convergence, causing compression (known by geologists as transpression), or divergence, causing tension (also called transtension). The Southern Alps of New Zealand exist because the sideways movement at the Alpine Fault (one of the world's most famous transcurrent plate boundary faults) also involves 11° of convergence. The resulting compression pushes up the Southern Alps at rates of up to 11 mm per year, and has been doing so for the past 5 myr. (If you do the sum, that is 50–55 km of uplift, which in turn gives you an inkling of the rates of erosion that take place in mountains.)

Fig. 2.3. View northeast along the straight section of the Alpine Fault forming the west side of the snow-capped Southern Alps in the South Island.
Photograph courtesy of Google Earth.

Chapter 3
Our Changing Country – the Last 45 Million Years

In this chapter we look at how tectonic processes at the Pacific–Australian plate boundary are changing New Zealand right now, and have changed it over the past 45 myr.

When you look at a map of New Zealand, it doesn't change from year to year. We therefore assume its shape is permanent. In fact, from a geological perspective New Zealand is one of the most mobile places on earth. Our location astride the active Pacific–Australian plate boundary, which cuts right through the New Zealand continent of Zealandia, ensures that. Furthermore, we have occupied this location for 45 myr, with the consequence that New Zealand has changed enormously, even in that geologically short span of time.

Go back further, and New Zealand has been associated with plate boundaries for most of its geological development. Each successive phase of development has been superimposed on, and has messed around, all previous phases. We can only unravel the resulting dog's breakfast by working backwards, removing each successive stage of deformation in order to get some understanding (not yet complete) of what went before.

Plate Movement is Predictable

To begin this unravelling process, we need to understand first just how the Pacific and Australian plates are interacting today. Plate movement is, to a large extent, regular and predictable, so we can project things backwards for 45 myr (Box 5.5 B) . To understand the big picture we need to delve into a bit of spherical geometry – spherical because the plates are parts of the surface of a sphere. Geometry on a sphere is different from that on a flat surface, but even so it is not complicated. The main thing to remember is that when any part of a sphere (i.e. a plate) is moved, the movement can be constructed as a rotation centred on a point that is somewhere on the surface of the sphere. A rotation is a regular and predictable movement – think of a record spinning on a turntable. A plate could be any portion of the record, and it might or might not include the central point of rotation. The parts further away from the centre move faster than those closer in, in a regular and predictable way. The centre is called the pole of rotation.

In the case of New Zealand, the pole of rotation of the Australian Plate is a long way away. The plate is moving northwards, away from the spreading ridge located between Australia and Antarctica, and in our vicinity the movement is just about due north at around 4 cm/yr. On Fig. 1.3, a north–south line of submarine volcanoes can be seen near the northwestern edge of the chart. This is a volcanic seamount chain recording precisely plate movement northwards across a volcanic hotspot (see also Chapter 2).

The Northern Subducting Sector

The Pacific Plate pole of rotation, on the other hand, is presently located within the plate, southeast of New Zealand at latitude 60°S, longitude 180°, just off the bottom edge of Fig. 1.3. The plate is rotating anticlockwise at a rate of 1° /myr. Beginning in the north, oceanic crust of the Pacific Plate is subducting beneath oceanic crust of the Australian Plate all the way from north of Tonga to East Cape (Box 2.1). From East Cape to Cook Strait, oceanic crust of the Pacific Plate is subducting beneath continental crust of the North Island, which is part of the Australian Plate. Because this southward transect is bringing us closer to the pole of rotation of the Pacific Plate, the rate of subduction declines from >10 cm/yr to about 4 cm/yr at Cook Strait. At the same time, the direction of subduction changes gradually from due west to southwest, because of the rotation.

The Central Sideways-moving Sector

At Cook Strait, the plate boundary enters continental crust, and the rotation direction is so close to being parallel to the boundary that the movement changes from strongly oblique subduction to dominantly sideways. The fact that there is now continental crust on both sides of the boundary has a strong influence on forcing the change, because neither side can be subducted. The transform boundary first appears as the Wairau Fault of Marlborough (Box 15.4 A), which forms the northernmost sector of the Alpine Fault. Here, as noted in Chapter 2, the southwestwards sideways movement of the Pacific Plate is exactly parallel to the fault, with the consequence that there is no compression and therefore the Southern Alps do not extend here. Further south, round the only bend in the Alpine Fault, the direction of plate movement is such that there is 11° of convergence/compression across the fault when seen in map view, in addition to the sideways movement of around 4 cm/yr. This compression, as noted in Chapter 2, gives rise to the Southern Alps. The Alpine Fault forms the western margin of the Alps all the way to Milford Sound, Fiordland, where it goes offshore.

Everything east of the Alpine Fault, including most of the South Island, the Chatham Rise and the large Campbell Plateau (all made of continental crust), lies on the Pacific Plate. South Island's West Coast, northwest Marlborough (the Marlborough Sounds) and Nelson, along with all of the North Island and the offshore Challenger Plateau, Lord Howe Rise and Norfolk Ridge, consist of continental crust lying on the Australian Plate.

The Southern Subducting Sector

Moving south from Milford Sound in Fig. 1.3, it can be seen that the plate boundary leaves continental crust, re-enters oceanic crust and becomes a subduction zone again. This time, however, the polarity (direction of subduction) is opposite to that north of New Zealand. Australian Plate lithosphere is now subducting underneath the Pacific Plate, but because of the nearness of the pole of rotation of the Pacific Plate the subduction motion is strongly oblique, the morphology of the trench is complex and variable, and there is only one known subduction volcano – Solander Island, south of Fiordland.

This complicated trench/ridge system extends southwards until it joins the mid-ocean spreading ridge that separates Antarctica from Australia and New Zealand. The three-dimensional geometry around the triple junction of one convergent and two divergent plate boundaries here is interesting to say the least.

Winding Back the Plate Movements Over 45 Million Years

The Australian and Pacific plates have been interacting in this kind of way consistently for 45 myr, since the present plate set-up came into existence, but at the same time there has been one other consistent change that has had profound consequences. Unlike the peg in the middle of a turntable, a pole of rotation is not fixed in position for ever. The pole of rotation of the Pacific Plate began life, 45 myr ago, within the present-day New Zealand landmass, very close to the plate boundary, and has since moved steadily (more or less) to the south-southeast. Remembering that rotation close to the pole is slower than rotation further away, the effect of this drift of the pole of rotation away from New Zealand has been to speed things up progressively at the plate boundary.

Thus, for the first 20 myr it is hard to detect the effects of the plate boundary in our geology, and there were no subduction volcanoes. Then, 25 myr ago, subduction at the plate boundary was abruptly kick-started. Sideways movement of the Alpine Fault began, but there were no Southern Alps yet. Subduction volcanoes at this time were all in the far north of the country and were not organised into a simple arc. Immediately prior to the appearance of volcanoes, there was an extraordinary event. The sudden jolt of plate convergence compressed a large volume of sediment that had been accumulating as a wedge along the continental margin since 110 myr ago (as outlined in Chapter 4). These sediments, along with some volcanic rocks whose ancestry is still somewhat mysterious, were buckled and lifted up high enough for them to slide southwestwards onto what is now Northland and East Cape (the Bay of Plenty had not yet opened). The name for such a body of displaced rock is an allochthon (pronounced 'allokthon'). In these two regions today, the Northland and East Coast allochthons play a prominent role in the local geology (Box 6.3 A, B).

At this stage, 23–20 myr ago, the country was being uplifted, after having been almost completely submerged 25 myr ago, and a lot of sediment was being eroded into new sedimentary basins.

It was not until 15 myr ago that the complex subduction system resolved itself into the single trench/arc system we have now, although at this point the new Coromandel–Colville arc was located about 120 km northwest of its present position (Box 5.5 D). This arc was active in that position for 10 myr, but most of the volcanoes are now offshore and hidden or buried. Today, we see them only in the Coromandel Peninsula and Kaimai Range of the North Island (Chapter 8).

About 5 myr ago, the final shift in the pole of rotation of the Pacific Plate took place, taking it to its present position at 180° longitude, 60°S. The importance of that last shift is that it was sideways, and caused a slight change in the direction of rotation of the plate. It set up the small amount of compression at the Alpine Fault, in addition to the dominant sideways movement. This is when the Southern Alps finally began to rise. The cumulative sideways displacement on the Alpine Fault over 25 myr is around 480 km, as shown by the offset of the Red Hills of Nelson (Fig. 3.1) and the Red Hill Range of southern West Coast (Fig. 3.2), which used to be joined (Box 19.3).

Another consequence of the shift in the pole of rotation is that the material pushed up into the Alps (and eroded almost as quickly) has been robbed from central South Island. So the northern two-thirds of the South Island is now about 100 km narrower than it was 5 myr ago (Box 19.1 B). This is a good example of the only way in which continental crust can be destroyed – by pushing it up in

Fig. 3.1. Red Hills of east Nelson. Photographer Lloyd Homer, GNS Science.

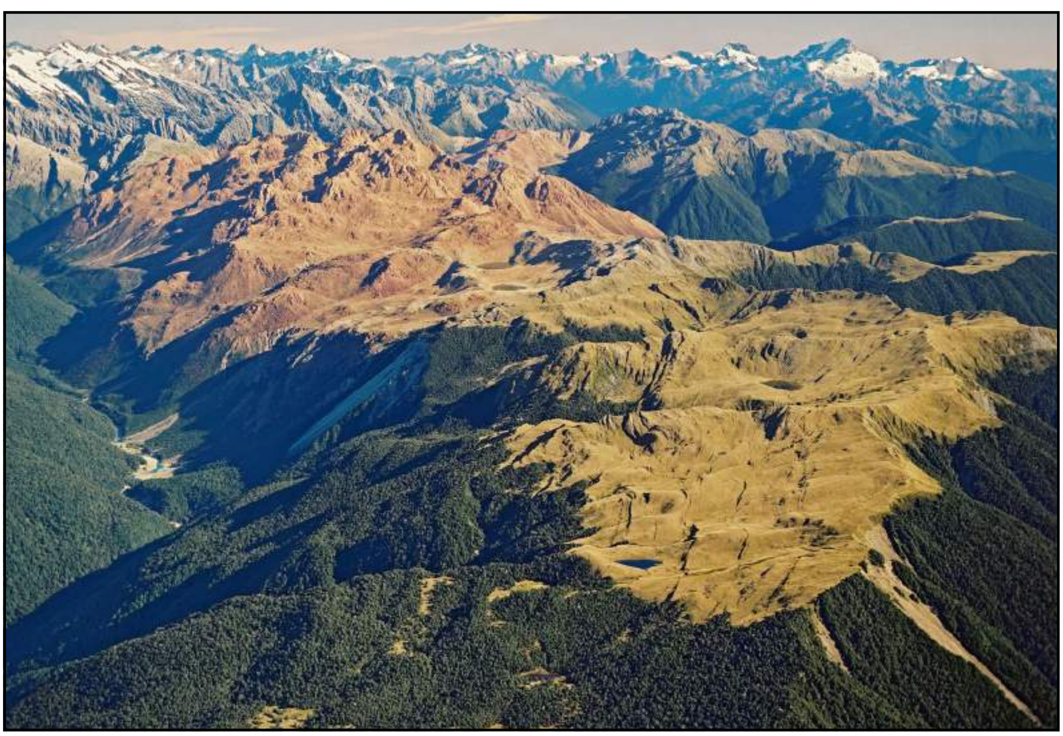


Fig. 3.2. Red Mountain and the Red Hills Range that straddles the West Coast–Southland provincial boundary. Photographer Lloyd Homer, GNS Science.

the air and eroding it away. Actually, of course, 'destroyed' is the wrong word – 'reconstituted' would be better. All that gravel, sand and mud has not disappeared, but has been redistributed around the country – mostly, as it happens with the present configuration of two islands separated by Cook Strait, to the east coast of the North Island (Box 14.2). This means that continental material from the Pacific Plate is being transferred to the Australian Plate, a fact politicians and economists might like to make something of!

While all this was going on over the last 5 myr in the South Island, the volcanic arc in the North Island shifted fairly quickly to its present location, probably largely by trench rollback (Chapter 2), and has been chugging away nicely in this position for about 2 myr. It is still retreating slowly to the southeast, and is also extending further southwards into the centre of the North Island, as the Australian Plate moves northwards over the Pacific Plate subduction slab. The shift left behind a 'remnant arc', the Colville Ridge, as shown on Fig. 1.3 and Box 5.5 D. The gap between the active arc (Kermadec Ridge) and the Colville Ridge remnant arc is known as a back-arc basin, and as a result of the southeastward rollback retreat of the active arc, the basin is getting wider. It extends into the North Island through the Bay of Plenty, which is also getting wider. For example, during the 1987 Edgecumbe earthquake the Bay of Plenty widened by 1 m. In effect, the North Island is unzipping down the middle – to date, it has unzipped as far as the central volcanoes.

When we wind back 45 myr of anticlockwise rotation of the Pacific Plate, at 1°/myr, eastern South Island and the Chatham Rise – which at present are oriented due east–west – end up oriented northwest to southeast, as shown in Box 5.5 C.

The Pace of Change

From reading this chapter you will have got the picture that in a mere – geologically speaking – 45 myr there have been profound changes in New Zealand, and that the pace of change is still accelerating. Geological change is not always as breathtaking as this, but life astride a plate boundary is constantly exciting. Because New Zealand has been associated with plate boundaries, both convergent and divergent, for most of its geological existence, we can expect to have an interesting time as we wind the clock back 500 myr in the next chapter.

Chapter 4

New Zealand between 110 Million and 45 Million Years Ago

In the period between 110 myr and 45 myr ago, before the inauguration of the modern Pacific–Australian plate boundary, New Zealand parted company with Gondwana. It was also during this time that the Tasman Sea opened up.

Gondwana was an ancient super-continent that comprised today's five southern continents (South America, Africa, Australia, Antarctica and Zealandia) plus India. For a time it was attached to the northern continents, forming a super-super-continent called Pangea. The series of events that had accompanied the subduction of the old Phoenix Plate beneath Gondwana – events that built up what is now the bulk of New Zealand – ended when that long-lived subduction episode finally ended, 110 myr ago (Chapter 5).

From 110 myr onwards, the mountains that had been formed by those subduction events gradually wore down, while a series of rift valleys was created by the development of a brand-new spreading plate boundary underneath our sector of Gondwana. Our situation then was exactly analogous to the East African Rift Zone today; the only difference was that we were at high latitude. The rifts existed for 30 myr. They included the present Taranaki Basin, which holds most of our recoverable hydrocarbons, and the Bass Strait Basin, which holds some of Australia's hydrocarbons.

Successful and Failed Arms

When a new spreading centre forms underneath continental crust there is a typical pattern of rifts: a three-armed star like the Mercedes-Benz logo. The continental crust is being stretched from inside, and the pattern that forms is similar to the splits that develop in a tomato skin as the fruit expands. To see a modern example, look at a map of northeast Africa and the Arabian Peninsula. In this case, the East African Rift Zone, the Red Sea and the Gulf of Aden are the three arms. The Red Sea and Gulf of Aden arms have proceeded (just) to production of oceanic crust. The typical pattern from here on is for one arm to develop no further (the 'failed arm'), while the other two (the 'successful' arms') continue and form a new ocean (Box 5.5 A).

Failed arms are important because they typically contain thick accumulations of sediment rich in coal and organic matter, which can be converted to oil and gas. Thus the North Sea and Nigerian oil and gas fields are failed arms from the Atlantic Ocean spreading ridge, while our Taranaki and Australia's Bass Strait oil and gas basins are failed arms from the Tasman Sea rift system.

In our case, oceanic crust appeared in what is now the Tasman Sea around 83 myr ago. New Zealand and New Caledonia were now identifiable as a separate continental entity for the first time, which has been called Zealandia. New ocean lithosphere and crust were formed symmetrically about the spreading ridge for 30 myr, and Zealandia was moved steadily out into the Pacific Ocean, with profound consequences for life on it. We also moved from high latitudes to mid-latitudes, again with large biological consequences.

Some 50 myr ago, the Tasman Sea itself became a failed arm. It had been connected to two other arms – the less advanced spreading ridges between Australia and Antarctica and between New Zealand and Antarctica. For whatever reason, the more advanced arm failed, and all spreading movement was transferred to the other two arms, where it continues to this day.

Why New Zealand is Largely Under Water

This is a suitable place to consider why so much of Zealandia is under water, when no other continent is like this. It is a result of the rifting process between 120 myr and 90 myr ago, which caused heating (from the rising convection cells underneath) and consequent uplift – hot rock is less dense than cold rock, and floats higher in the dense rock underneath. The mountains and high ground that were created by this uplift suffered deep erosion, thus losing material from the top. At the same time, the rifting process stretched and thinned the crust from below, which is now around 20 km thick instead of the usual 30 km. As this thinned crust gradually cooled following the rifting episode and movement away from the mid-Tasman spreading ridge, it subsided, and now most of it does not have enough freeboard to project above present-day sea-level. The stretching process is explored further in Chapter 20.

The progressive subsidence discussed above gave rise to a unique and economically important episode in New Zealand's history. Large coastal swamps formed and thick peat layers accumulated. The swamps were then submerged underneath large, shallow shelf seas in which extensive limestone beds were deposited on top of the peat, the latter in time turning into coal seams (Boxes 7.3 A, B). In places the coal measures and limestone beds were deposited on top of the older rift valley sediments, and collectively they included the source rocks for most of our hydrocarbons. This is when we acquired most of our economic coal, oil, gas and limestone resources.

Fig. 4.1. The Stockton Opencast mine on the West Coast works Eocene Brunner Coal Measures that were deposited in a large coastal swamp about 35 myrs ago before being completely submerged beneath the sea.
Photographer Lloyd Homer, GNS Science.

At the peak of subsidence and marine transgression, around 30–25 myr ago, shell limestone was accumulating over most of what is now New Zealand, and land areas were small and worn down. This was a critical time for all of our land organisms, both plant and animal. In evolving from their Gondwanan roots towards the present biota, all lines of evolution had to pass through this 'Oligocene bottleneck' of limited, low-lying land space. Given that much of our flora and fauna has arrived since the separation of Zealandia from Gondwana, we have to hold open the possibility that Zealandia may even have become totally submerged at this time.

Note that this submergence event, partial or complete, occurred well after the 45 myr inauguration of the present plate boundary. The very slow and small effects of the early boundary allowed the previous trends of subsidence to continue for a further 20 myr. It was only after the increased plate convergence 25 myr ago that Zealandia began to rise again, and to be converted into the New Zealand we now know, as we discussed in Chapter 3.

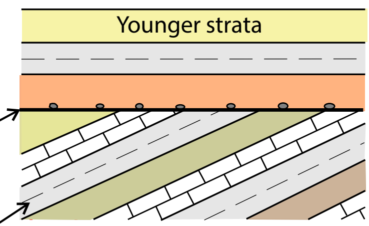
Country-wide erosion surface

The prolonged erosion and subsidence beneath the sea left a near-planar erosion surface covering most of New Zealand. New strata gradually covered this surface, creating a geological unconformity over the whole country, known as the 'Great New Zealand Unconformity' (Fig. 4.2 and Box 4.1). An unconformity is an erosion surface separating older rocks from younger strata that covered the older rocks after an interval of time. This interval may be long.

Fig. 4.2. Exposure of the "Great New Zealand Unconformity" at Waitete Bay, Coromandel Peninsula running through the middle of the photograph. This erosion surface cuts across steeply dipping Cretaceous greywacke overlain by early Oligocene conglomerate. There is a gap of about 70 million years between the older and younger rocks.
Photographer Bruce Hayward.

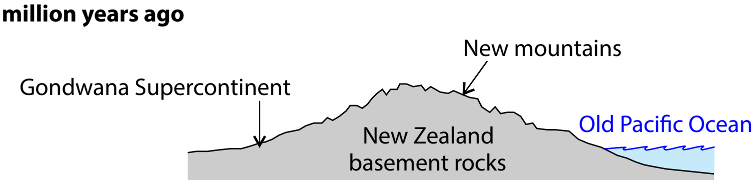
Box 4.1. Unconformities.

Unconformity is the relationship between older and younger rocks that gives information on episodes of tectonic activity and erosion of rocks. In its simplest form -

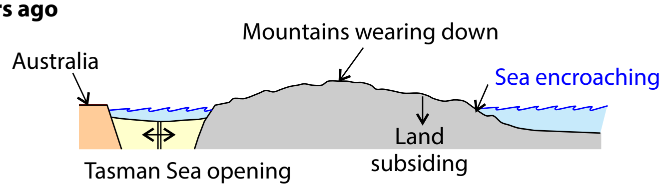


<u>The Great New Zealand Unconformity</u>

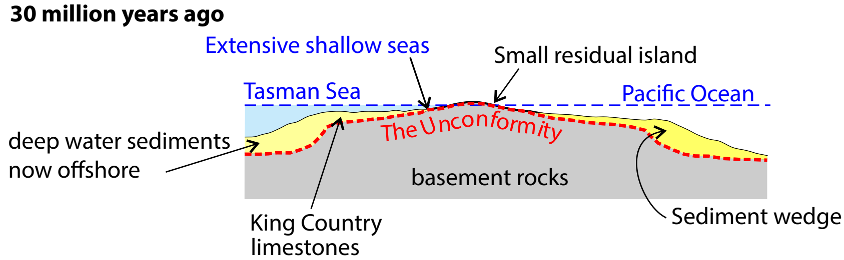
A 100 million years ago



B 70 million years ago

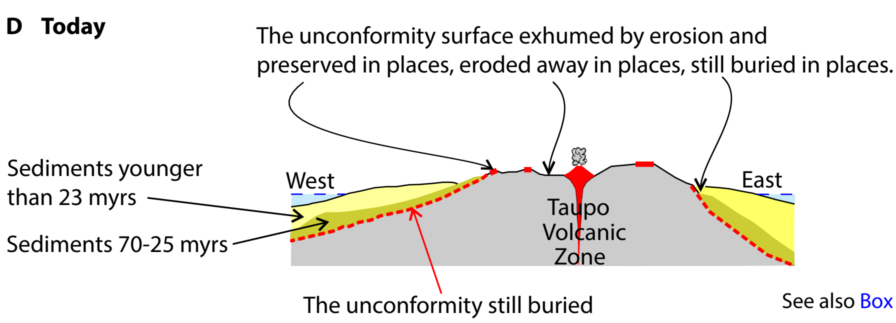


C 30 million years ago



25 million years ago - a major change took place, to the present tectonic regime of earthquakes, volcanoes and rising mountains.

D Today



See also Box 5.5 B

Chapter 5

Our Portion of Gondwana between 500 Million and 100 Million Years Ago

The period between 500 myr and 100 myr ago is characterised by the incremental growth of Gondwana by the addition of successive strips of continental crust, as well as by igneous activity associated with subduction.

We begin with the idea of a continent growing by having bits and pieces of continental crust added to it, one at a time. This is the notion of the 'tectonostratigraphic terrane'. The term is a mouthful, with three parts: 'tectono', as we saw earlier, is to do with the movement and deformation of rocks; 'stratigraphic' relates to time; and 'terrane' is a chunk of country. The spelling 'terrane' was chosen deliberately to signify a solid chunk of country, as opposed to 'terrain', which signifies surface features.

The Gondwana Case

In the case of Gondwana, a long-lived subduction zone existed along the edge of the Antarctica-Australia-New Zealand sector, very much like the situation along the west coast of South America today. The subducting plate was the old Phoenix Plate, and over the course of nearly a billion years it brought a succession of terranes to the continental margin, where they were all incorporated into Gondwana. Thus Gondwana, like all continents, was a patchwork of unrelated bits and pieces of continental and oceanic crust: a mosaic of terranes. Older terranes and parts of Gondwana were left behind in Antarctica and Australia when New Zealand separated, and we consist mostly of the younger terranes – both the most recently accreted terranes and those containing the youngest rocks.

If you think about it, there is no relationship between the age of the rocks forming a terrane and the time of its accretion to a continent. It doesn't necessarily follow that the most recently accreted terranes consist of the youngest rocks – they can be any age. Thus the fact that Zealandia does contain younger terranes than Australia and Antarctica is of some geological interest, and it tells us things about the processes that take place at subduction zones adjacent to continents.

How Continents Grow in Total Area

As we noted above, continents grow with time. The source of new continental material is the volcanic activity associated with subduction. Strictly, this should be termed igneous activity because it includes both the extrusive (volcanic) and intrusive (granite) components. The combined effect of the melting of subducting lithosphere at 100 km depth (Chapter 2) and the fitful rise of the resulting magma towards the surface – with rest periods sitting in magma chambers – is to allow the lighter components of the melt to separate from the heavier ones. The lighter material that eventually reaches the surface, or is intruded as high-level granites, is enriched in silica, to varying degrees.

Different Kinds of Terrane

Many kinds of terrane make new continental crust. They arise in various ways, but usually through the subduction-related production of magma, which takes place both underneath the edges of continents and out in the oceans. In the former case, the new magma is intruded into and extruded onto pre-existing continental crust. However, erosion carries a lot of this new stuff into the sea, along with older stuff, where it forms a new terrane – and thence new continental crust – through the subduction process.

Magmatic Arc Terranes

Out in the ocean, where oceanic crust subducts beneath oceanic crust, the resulting magmatic arcs build substantial submarine ridges made primarily of volcanic products, as in the Pacific Ring of Fire. The typical Ruapehu-type andesitic volcano sends out most of its products explosively, as broken-up lava. This material is easily eroded and moved around, building the submarine ridges. These volcanic ridges are long and voluminous, and they constitute terranes. They represent the first stage of production of new continental crust. At some stage, the plate tectonic process will add them to a continent.

To illustrate these two types of newly forming terrane, consider the New Zealand example as shown in Fig. 1.3. Starting in the top right-hand corner, follow the subduction pair of deep trench and volcanic ridge from the Kermadec Islands (volcanoes) southwestward towards New Zealand. This actively forming intra-ocean ridge extends as far as White Island volcano in the Bay of Plenty. The parallel Colville Ridge, a short distance to the west, is the older equivalent of the Kermadec Ridge. It was active between 25 myr and 5 myr ago. The two ridges could become two terranes in a future scenario; they could be pushed together and onto a continent in the future, but for the time being they will remain separate and distant from a continent. These are volcanic or magmatic arc terranes.

Accretionary Wedge Terranes

As you can see in Fig. 1.3, the Kermadec Trench loses its simple character as soon as East Cape is reached. This is because there is now far more sediment entering it – in addition to the volcanic products, there is all the sediment being eroded from the axial mountains and the eastern seaboard of North Island. And there is even more than this: much sediment eroded from the South Island also enters the trough. It does so by way of the submarine canyons seen on the chart that pass alongside the Marlborough coast as far south as Banks Peninsula; or through Cook Strait from along the West Coast and around Farewell Spit, and from Taranaki around the Manawatu coast.

The sediment is transported down into, and along, the Hikurangi Trough by turbidity currents (Boxes 6.5 A, B). What is important to note is that large quantities of sediment (sand, mud and some gravel) are being shovelled onto a conveyor belt and carried into the subduction zone. What happens next is a matter of great scientific interest. Similar situations pertain in many other subduction zones, and much research has been done utilising detailed contour mapping, very high-resolution seismic profiles from research ships, and dredging and deep-sea drilling. In a nutshell, the overriding continent (North Island in our case) acts like a bulldozer blade. It scrapes the loose sediment off the conveyor belt (subducting plate) and heaps it up. As the pile accumulates, it maintains a characteristic angle of repose and builds further out. Fresh sediment is carried underneath the pile.

The pile ends up being organised into slices that dip into the bulldozer blade, with a characteristic arrangement of oldest slices at the top and youngest slices at the bottom – i.e. the opposite of the normal geological arrangement of oldest at the bottom, youngest at the top. The slices are separated by thrusts (geological faults, in which the top side is pushed up and over the bottom side). Furthermore, the bulldozer blade rides up over some of the sediment heap, which then gets pushed down the subduction zone into regions of higher temperature and pressure. While this is going on, the layers of sediment are being constantly deformed. There is enormous energy in this process.

The end result is a terrane that is a long, wedge-shaped body of sedimentary material. This material is partly volcanic and partly eroded from the continent, is severely deformed internally, and is converted into solid rock by high temperatures and pressures. The final rock assemblage is known as greywacke (pronounce the final 'e' as 'ee'); Boxes 6.2 A–C. Each of these wedges comprises many thousands of cubic kilometres of material, and it is positioned ready to be added to the adjacent continent. This appears to happen when subduction ceases or moves elsewhere, and the wedge can rise according to its relatively light weight. The name for this kind of terrane is an accretionary wedge.

There is a complication that can enter here – one that is giving geologists headaches in their attempts to unravel New Zealand's old terranes. This is that oblique subduction (like we have now under the North Island) can shuffle the terrane wedge sideways, for long distances. This seems to have happened to several of our old terranes, and the problem is to reconnect them with their point of origin.

Forearc Basin Terranes

A third type of terrane is a forearc basin terrane. These form in positions between the magmatic arc and accretionary wedge. They typically contain volcanic-derived sandstones and mudstones (but without lavas). They are not as deformed as accretionary wedge terranes.

New Zealand's Basement Terranes

As a result of the various processes involved, the New Zealand continent consists of about 10 tectonostratigraphic terranes that were added to Gondwana at different times by subduction. There are several accretionary wedge terranes (hence the great preponderance of greywacke rocks throughout the country), two volcanic arcs, and a forearc basin terrane. Some parts of the accretionary wedges were metamorphosed (changed by heat and pressure) to the state of schist, as found in Central Otago (Boxes 15.3 A, B).

It is also important to note here that if the subduction zone/plate boundary keeps on stepping backwards as terrane after terrane is added to the edge of the continent, so the volcanic/magmatic arc can keep on re-forming closer to the new continental edge, on top of and through the newer terranes. So, in our case, in addition to the two accreted volcanic arc terranes, there are two *in situ* magmatic arcs. These are now seen mostly as granite because the volcanic superstructure has been worn away. The granites are intruded into the older terranes, and are visible today in the South Island, in northwest Nelson, the West Coast, Fiordland and Stewart Island/ Rakiura. The younger arc lies to the east of the older one, as would be expected.

As we work our way around New Zealand, we will encounter these tectonostratigraphic terranes in many different places and guises. They form long, narrow strips parallel to the old margin of Gondwana, and were added successively, though not necessarily one at a time, from subduction of the old Phoenix Plate beneath Gondwana. The plate was moving from present-day east to west. Thus, the first terrane that arrived is the westernmost on maps of terranes.

Our oldest rocks are contained in the two westernmost terranes. They are of early Paleozoic age (510–420 myr). Interestingly, those two terranes retain fragments of early cover strata (see below), of Devonian (420–360 myr) and Permian (300–250 myr) age, that have survived all the erosion of the intervening period. The remaining terranes are aged between the Carboniferous–Permian periods (330–250 myr) and Middle Cretaceous (110 myr). Remember that the age of the rocks in the terrane is not the time at which the terrane was added to the continent.

Fig. 5.1. The oldest fossils in New Zealand are Cambrian (about 510 million years old) and occur in Trilobite Rock, Cobb Valley, Northwest Nelson. Photographer Bruce Hayward.

Box 5.1. Basement terranes.

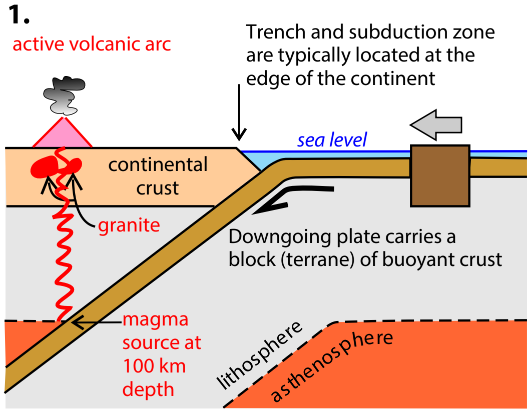
New Zealand's basement rocks underpin the country. They are thick — more than 10 kilometres — and enormously voluminous — millions of cubic kilometres. Think of them as comprising about ten large building blocks — terranes — that were individually added to the margin of the Gondwana supercontinent over a period of time. The period involved was about 450 myrs, which covers both the times during which terranes formed, and the later times when they became attached to Gondwana. All of this happened before the opening of the Southern Ocean and the Tasman Sea, which began 100 myr ago. It was the period when Antarctica, Australia and Zealandia were all still joined together as part of the Gondwana supercontinent - which was dwindling as Africa, India and South America were breaking away, while it was still growing in the New Zealand sector by the addition of terranes.

How subduction zones respond to terrane collisions

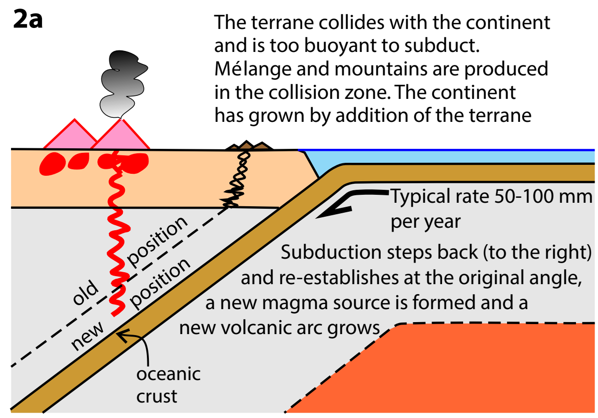
When a buoyant terrane is located on a downgoing plate and arrives at a subduction zone, as it must inevitably do sooner or later, the subduction zone is forced to re-arrange itself — it either steps backwards to the other side of the terrane, or it reverses direction, as on the following diagrams.

SCENARIO 1 — the overriding plate carries continental crust approximately 30 km thick

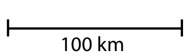
1.



2a

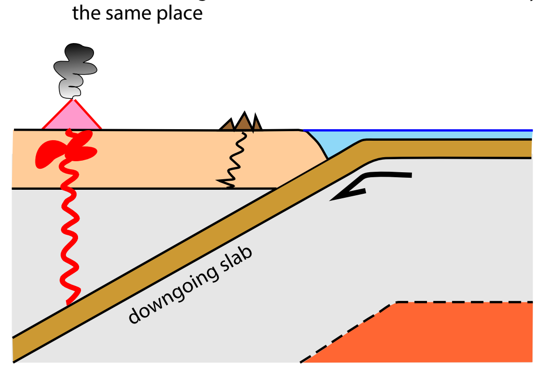


These diagrams are drawn to true scale (vertical = horizontal) apart from the volcanoes which are exaggerated



A terrane arrival is a genuine collision involving damage. It is simply slower than a collision between two moving cars.

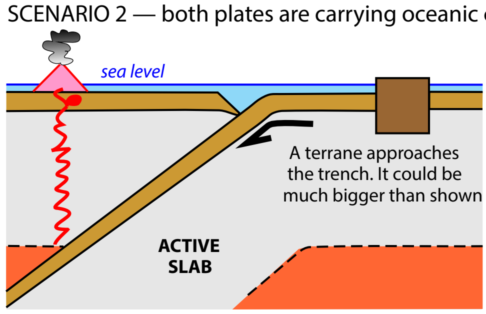
or 2b — the subducting slab flattens, and the volcanic arc stays in the same place

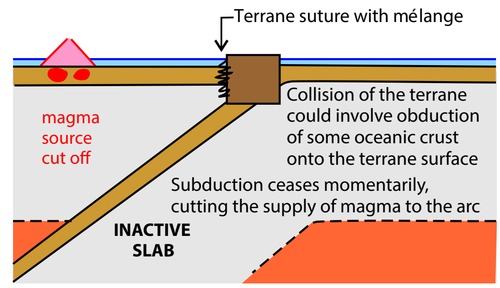


This is what happened during the addition of New Zealand's younger terranes to Gondwana. The arc became the Median Batholith terrane, which we see now mostly as the granite bodies.

Box 5.2. Basement terranes continued.

SCENARIO 2 — both plates are carrying oceanic crust 10 km thick





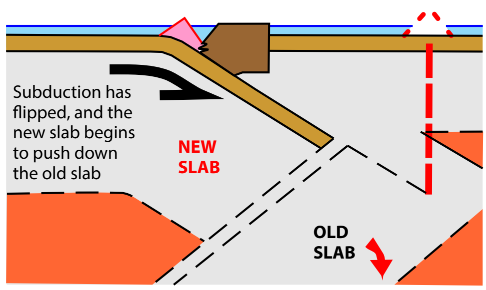


Plate movement continues and subduction must resume, either by stepping to the right, as in Scenario 1, or by reversing direction (the technical term is flipping!). Collision of a really big terrane, like India with Asia, can have global repercussions on plate movements.

Remembering that many downgoing plates make an oblique approach to the subduction zone, it can happen that, following its initial collision, a terrane is then shunted sideways along the continental margin by the sideways component of plate approach, possibly for hundreds of kilometres.

When the new slab reaches a depth of 100 km it will establish a new magma source for a new arc. At around the same time the old arc (a terrane in its own right, made of newly formed continental crust) will itself collide with the first terrane, giving now a composite terrane (the beginnings, perhaps, of a new continent). Following further adjustment of the subduction zone, steady state subduction will resume. World stock of continental crust has increased by addition of the dead arc — this is how continents grow in total size.

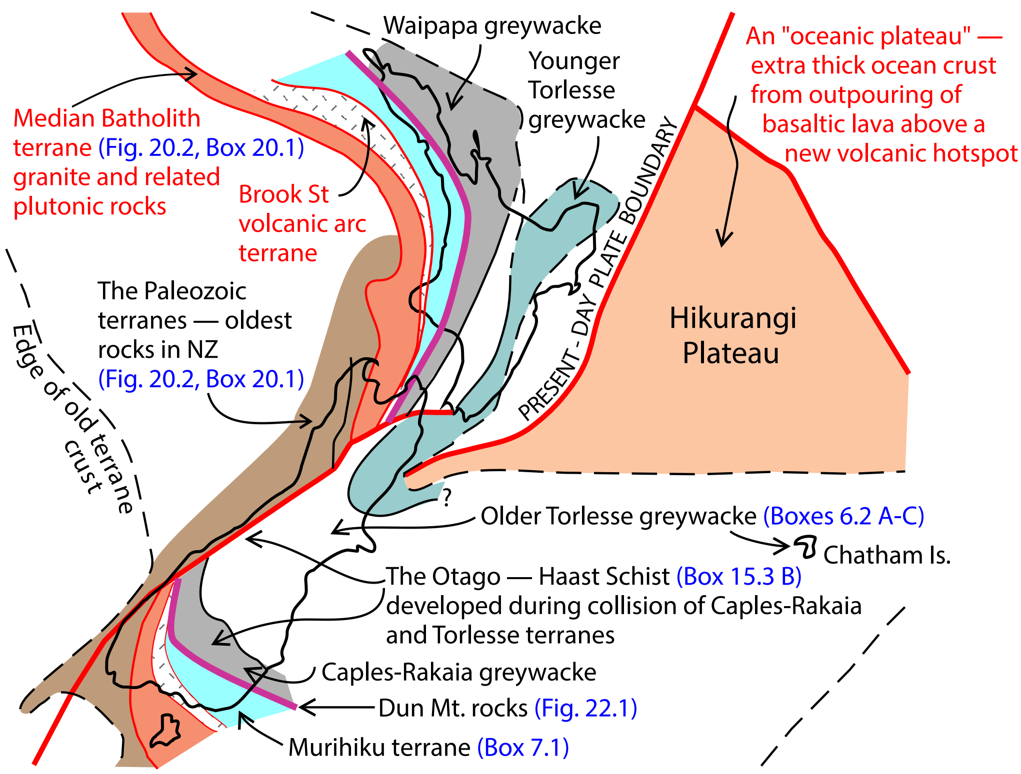
New Zealand basement terranes

The following maps show the main basement terranes (those older than 100 myr). All of them, apart from the granitic batholith, were brought from somewhere else to a long-lived subduction zone at the Gondwana continental margin, at different times, and there transferred to Gondwana by rearrangement of the subduction zone in the way shown above. When the Torlesse greywacke terrane collided with the Caples - Waipapa greywacke terrane, the energy of the collision caused the metamorphism of parts of both terranes to schist. **The Median granitic batholiths** and their associated volcanic rocks were generated in place by subduction, between 250 myr and 120 myr ago. They are the volcanic arc of the subduction zone. As the diagram shows, the subduction slab must have flattened progressively as terranes arrived, in order to allow the arc to remain in the same place. **The Hikurangi Plateau** is an oceanic plateau which seems to have arrived at the margin on the Phoenix Plate, and attached itself to the Chatham Rise, at about the time that subduction ceased at the Gondwana margin.

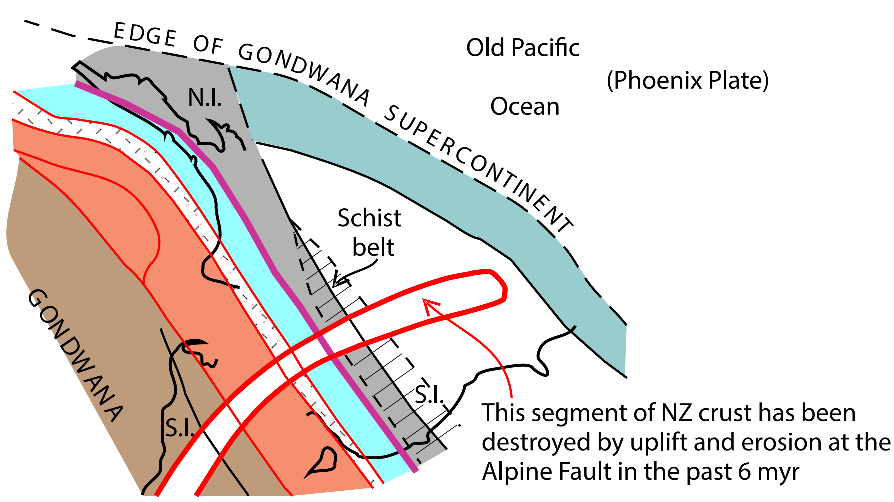
What manner of rocks are the terranes, where did they come from, and when did they arrive (dock with Gondwana)? The first question is the easiest to answer. From present-day west to east, the old Buller Terrane (Fig. 20.2, Box 20.1) comprises continental margin deep-water sediments; the Takaka Terrane (oldest rocks in NZ) comprises tropical-latitude marine sediments with a volcanic arc. These two terranes record an earlier phase of terrane arrivals, and their volcanic arc is probably represented by the Karamea Granite suite (Fig 20.2, Box 20.1). Continuing eastwards, the Median batholith granitic terrane formed where it is now, as noted above; the Brook Street Terrane is an old volcanic arc (the rocks are from 270-220 myr old - that is not the time at which the rocks became a terrane, some time later); Murihiku Terrane (Box 7.1) is an old forearc basin; Dun Mountain is an obducted slice of ocean floor (ophiolite) with an associated set of forearc basin rocks; and finally all the greywacke terranes are accretionary wedge rocks (Box 6.2 A). In other words, all the terranes were formed in association with a convergent (subducting) plate boundary, and clearly the two oldest terranes were docked with Gondwana before the Median batholith started to form, 250 myr ago, because the granites intrude the old rocks. For the rest, however, it is proving very difficult to sort out. The Brook Street volcanic arc terrane, for example, although it is now located adjacent to the Murihiku Terrane forearc basin, is not the source of the sediments that filled the Murihiku basin, even though a volcanic arc is expected to form with an adjacent forearc basin (Box 7.1). Similarly, the times at which the terranes docked with Gondwana are still unclear.

Box 5.3. Basement terranes continued.

This map shows only the basement rocks — everything older than 100 million years.



Below: How the terranes may have been disposed 100 myr ago, before the tectonic squeezing and bending associated with the development of the Alpine Fault.



In geographic order from west to east, the terranes are:

Buller Terrane

Fiordland, Westland and northwest Nelson. Ordovician flysch sequence of quartz-rich sandstones and black mudstones. A continental passive-margin sequence, with no input from active volcanism. Dated by graptolite fossils.

Takaka Terrane

Fiordland (metamorphosed), Westland and northwest Nelson. Cambrian volcanic arc sequence overlain by non-volcanic Ordovician–Silurian–Devonian quartz-rich sandstones and limestones (now marble). Tropical origin. These first two terranes do not fit comfortably into the categories of terrane listed above, being continental passive-margin sequences that were not accretionary wedges, but nevertheless became detached from their parent continents and moved as independent terranes.

Brook Street Terrane

Southland and east Nelson. Permian oceanic magmatic/volcanic arc sequence, with its flanking volcano/sedimentary apron. No continental input.

Murihiku Terrane

Southland, east Nelson and west Auckland–north Taranaki. Latest Permian to Early Cretaceous, very thick (~20 km), well stratified and simply folded sequence of volcanic sandstones and mudstones, with a small continental contribution. The fill of a large and very long-lived forearc basin to a magmatic/volcanic arc located on a continent lying to the west.

Dun Mountain-Maitai Terrane

Southland, east Nelson, and concealed under Cook Strait and the North Island (where it is traceable by a distinctive geophysical signature, the Stokes magnetic anomaly – Chapter 22). Permian ophiolite sequence (Dun Mountain; obducted oceanic crust with mantle rocks overlain by volcanic rocks), overlain in turn by a continental-margin sequence of sandstones and limestones, in part volcanic-derived and in part continent-derived (Maitai Group).

Caples-Waipapa Terrane

Caples Terrane in Southland, east Nelson–Marlborough Sounds. Permian–Triassic volcanic-derived, deep-water accretionary wedge accumulation. The North Island equivalent lies in part of the Waipapa composite terrane.

Rakaia Terrane (older Torlesse greywacke)

Canterbury–Southern Alps, North Island axial ranges. Permian–Triassic continent-derived, deep-water trench flysch sequence; an accretionary wedge accumulation, not volcanic-derived. Our most extensive terrane. The Otago Schist (Chapter 18) is derived from metamorphism of part of the Caples and part of the Rakaia terranes, and was produced by collision of the two terranes around 200 myr ago, long before they collided with Gondwana.

Pahau Terrane

North Canterbury–Marlborough, eastern flank of North Island axial ranges, also part equivalent of the Waipapa composite terrane. Late Jurassic–Early Cretaceous continent-derived, accretionary wedge accumulation, very similar to Rakaia but less strongly metamorphosed.

Waioeka-Omaio Terrane (younger Torlesse greywacke)

Waikato–Coromandel and eastern North Island. The final episode in accumulation on the Gondwana convergent margin. Early Cretaceous volcanic- and continent-derived flysch sequence, in part originating from the Gondwana interior, in part from active volcanism and in part from filling of the old, now inactive subduction trench. This terrane is a transition between the convergent and spreading/passive phases of the Gondwana margin, and was emplaced in its present position by sideways displacement.

Karamea and Median batholiths

Sometimes called terranes, but not displaced from their original positions as the roots of active magmatic/volcanic arcs. The Median Batholith is actually tangled up with various terrane bits and pieces. The two batholiths record periods of apparently straightforward subduction of the Phoenix Plate beneath Gondwana, 395–295 myr and 245–105 myr ago, uncomplicated by terrane arrivals.

Distinguishing Between Basement Rock and Cover Strata

The terrane rocks are very thick – 10 km or more – and they form the 'basement' of New Zealand. This is a good place to introduce the distinction between basement and cover rocks. Old, hard basement rocks of indefinite thickness are covered in many places by younger, softer cover strata (layered sedimentary rocks, not metamorphosed), whose thickness varies. These overlying strata can reach several kilometres in thickness, but are generally less than 1–2 km thick. Most of the cover strata have accumulated since we parted company from Gondwana, so they contain our history for the past 100 myr.

How do we Know When a Terrane Gets Added to the Continent?

We use several lines of evidence to deduce the time when a terrane was added (sutured or accreted) to the continent. Cover strata can help, because if they lap across from one terrane to another they define a younger time limit. Igneous intrusions that pass across terrane boundaries also define a younger limit. The process of accretion is a genuine collision (we tend to think of collisions as high-speed car crashes, but they can happen slowly), and, as in all collisions, damage is sustained. It is sometimes possible to date the collision damage using radiometric techniques. The age of the rocks in the terrane defines the older limit, but commonly we cannot define the age of collision closely.

Order of Terrane Arrival

The terrane chart (Box 5.4) summarises the present state of knowledge regarding terrane collision events prior to the date of final docking with Gondwana. The chart differentiates types of terrane, and shows the period of time during which they accumulated, wherever that was. It also shows the periods during which the two successive magmatic/volcanic arcs, Karamea and Median, driven by subduction of the old Phoenix Plate, were active. The chart indicates clearly that magmatism flourished between terrane collision events, but was switched off during collision

Box 5.4. Terrane chart, summarising terrane collision events.
Chart produced by Nick Mortimer, GNS Science.

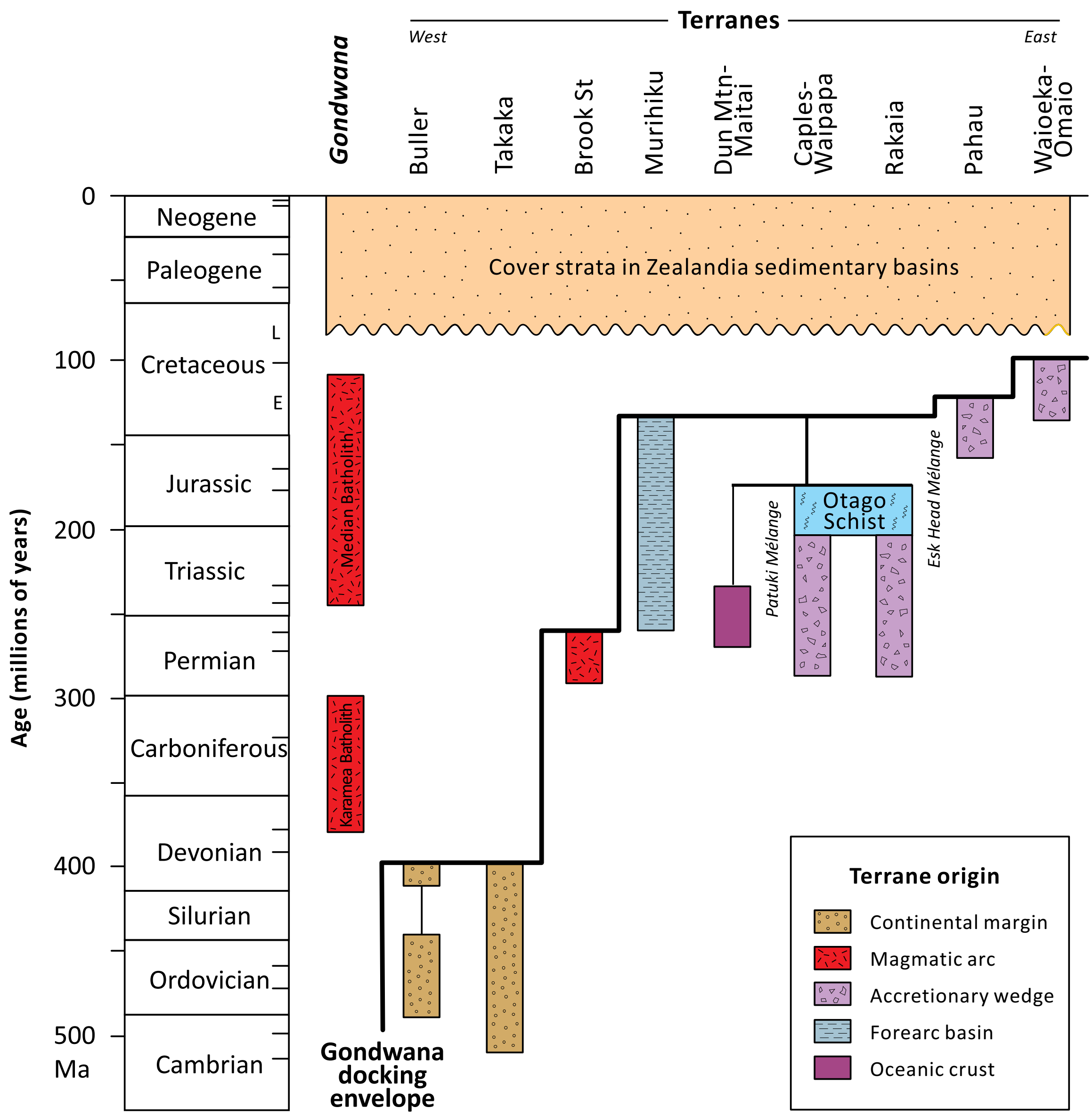


Table 5.1. – to accompany Box 5.4

Rules of the Terrane Game

- A terrane is a sizeable chunk of crust, which is created by plate tectonic activity in some way, and that existed as an independent entity, travelling piggyback on a moving plate, for a limited period of time.
- A composite terrane is made of two or more older terranes that were amalgamated at an earlier time.

Continued on next page.

Table 5.1. – *continued from previous page.*

- A terrane can be added directly to a continent from a subducting plate.
- A terrane can be added indirectly to a continent by being moved sideways along the continental margin from a previous location.
- The subduction that brings terranes to continental margins and other subducting plate boundaries (e.g. between two oceanic plates) usually drives a magmatic/volcanic arc on the overriding plate.
- Terrane sutures (faults or mélange zones) are commonly reactivated as faults during renewed tectonic activity.
- A terrane must be added to the continental margin later than the terrane inboard from it – on the terrane chart, later than the terrane to its left (hence the Gondwana docking envelope).
- Any earlier terrane amalgamations – those below the Gondwana docking envelope – occurred elsewhere.
- Mélange zones can occur both between and within terranes.

events. The young original age of the Murihiku Terrane (up to 130 myr) places a major constraint on the Gondwana 'docking envelope', which requires that most of the terrane amalgamations took place somewhere else. That in itself raises a whole bevy of new questions.

Some conclusions

A number of conclusions can be drawn from the terrane chart:

1. The Murihiku Terrane sets a young Gondwana docking time (around 120 myr ago) for itself and therefore for the terranes that followed it, regardless of the age of those terranes.
2. A number of earlier terrane amalgamations did occur – i.e. they are below the Gondwana docking envelope on the chart. However, where and how these earlier amalgamations occurred is not yet known.
3. Where a terrane comprises an accretionary wedge body that accumulated below the continental trench slope and above a subducting slab (e.g. the Caples, Rakaia and Pahau terranes), the mechanism whereby the accretionary wedge was detached from its parent continent, uplifted, transported as an entity detached from the rest of the subduction zone, and sutured to the edge of Gondwana, is not clear. Sideways shuffling may have taken place.
4. Just how two accretionary wedge terranes can collide with sufficient force to generate a large body of schist (e.g. the Otago Schist from collision of the Caples and Rakaia terranes 200 myr ago), apparently away from the Gondwana margin, is not entirely clear.
5. The relationship between docking with Gondwana and folding and metamorphism of terranes can be a direct cause and effect (e.g., apparently, the Buller Terrane) but this is not always clear. Metamorphism of accretionary wedge terranes is thought to occur largely within the wedge as it is compressed and expands during formation, i.e. before the wedge is separated from its adjacent and overlying continent.

From the terrane chart, it can be seen that the arrival and docking of the Buller and Takaka terranes was not accompanied by any known arc magmatism. They may have been located further west on Gondwana, but there are no volcanic products in either terrane, apart from the earliest rocks in the Takaka Terrane. There is a strong supposition that arc magmatism was switched off during accretion, despite the 40 myr time gap between the two docking events.

Several other statements can be made about subduction-driven arc magmatism in relation to the arrival and docking of terranes:

1. No terranes docked with Gondwana during the period of activity of the Karamea magmatic/volcanic arc.
2. There is no obvious cause for the cessation of the Karamea Arc.
3. Approach and docking of the Brook Street Terrane was not accompanied by arc magmatism within New Zealand, though the terrane is itself an oceanic magmatic arc (i.e. it was not built on Gondwana or any other continent) that accumulated somewhere above a subduction zone during the gap between Karamea and Median arcs.
4. No terranes docked with Gondwana during the duration of the Median magmatic/volcanic arc.
5. The Median Arc was terminated by events related to accretion of the Rakaia-Caples Terrane (see below).
6. As a general statement, we can say that arc magmatism along the New Zealand position of the Gondwana margin occurred only when subduction was unhindered by complications accompanying terrane arrival.

The Buller, Takaka and Brook Street terranes seem to have docked neatly and sequentially in that order, one at a time. The remainder are more complicated.

Most logically, docking of the various terranes to the right of the Brook Street on the terrane chart went in the following order:

1. Rakaia to Caples, generating the Otago Schist (210–200 myr ago).
2. Rakaia-Caples composite terrane to Dun Mountain, generating the Patuki Mélange (?170 myr).
3. Dun Mountain-Caples-Rakaia to Murihiku (?130 myr). Murihiku-Dun Mountain-Caples-Rakaia to Gondwana (?130 myr).
4. Pahau to Gondwana, generating the Esk Head Mélange (120 myr).
5. Waioeka-Omaio to Pahau by sideways shuffle (90–100 myr).

As geological investigations progress, our understanding of the timing of these events will undoubtedly change.

Remaining questions

There are many remaining questions concerning New Zealand's terranes, chiefly involving the whereabouts of their original formation, and the mechanism(s) by which they were detached from their place of origin, transported to, arrived at and docked with the Gondwana margin. It seems that there might have been considerable sideways transport along the margin, following initial contact. That would imply oblique subduction of the old Phoenix Plate beneath Gondwana for at least some of the period in question. Oblique subduction could also account for the absence of arc magmatism at some times, as mentioned above and discussed in Chapter 9.

The Dun Mountain-Maitai Terrane poses a difficult problem. This slice of lower ocean crustal rocks, or ophiolites, is overlain with just slight unconformity by a thick sequence of continent-derived sediments that include shell limestones and plentiful andesitic volcanic detritus (i.e. coming from an active magmatic/volcanic arc that was located on the adjacent continent). This time, however, 260–230 myr ago, was not one of arc activity based on Gondwana rocks within the New Zealand sector, which suggests that the Dun Mountain-Maitai Terrane was located either adjacent to another continental margin or, more probably, further along the Gondwana margin. Another oddity is that the Maitai rocks have not been in an accretionary wedge, but were built out from the continental margin in the manner of a passive-margin wedge of sediment to rest directly on the ocean crust volcanic rocks. Both ocean crust and the overlying pile of strata must have been close to horizontal. If subduction was occurring it has left few clues.

At some time between 240 myr (the age of the youngest Maitai strata) and ?170 myr ago (a likely age for production of the Patuki Mélange marking the Dun Mountain–Caples-Rakaia terrane collision), the Dun Mountain-Maitai rocks were somehow detached from their parent continent – just as the Buller and Takaka terranes had been – and set adrift. Given that the deformation of the Dun Mountain-Maitai Terrane to its present steeply dipping form, with an upturned sliver of ocean crust outboard of the sedimentary rocks, requires a solid backstop for the colliding Caples-Rakaia Terrane to dig and push against, we assume that the Dun Mountain Terrane was located adjacent to the Gondwana continental margin. Plate movement and subduction would have been required to accomplish all the terrane movements required.

With the Dun Mountain–Caples-Rakaia collision accomplished, the next step was for this new composite terrane to collide with the Murihiku Terrane, and then for this to collide with the Brook Street Terrane, bringing us back into 'home territory'. As with the other wandering terranes, the Murihiku was detached from its original place at the Gondwana margin, presumably at the time its youngest rocks were being deposited (130 myr ago), and moved laterally along the margin. Subduction was occurring, which would have held the Murihiku against the Brook Street, though there is no indication of a subduction trench. The final act in this saga was for subduction to bring the composite Dun Mountain-Caples-Rakaia Terrane into collision with the Murihiku Terrane, 130–120 myr ago, compressing the Murihiku rocks against the Brook Street Terrane and forming the giant folds that are characteristic of Murihiku strata.

Another big question concerns the timing of, and mechanism by which, the western terranes (Buller, Takaka, Brook Street, Murihiku and Dun Mountain, along with the Karamea and Median batholiths) were strongly squeezed, and in places tectonically completely excised, through the middle of New Zealand on either side of the Alpine Fault (Box 5.3). There are two obvious possibilities. The first is that it was caused by strong compression and dextral shear associated with having the Alpine Fault cutting obliquely through the terrane belts during the 480 km of sideways displacement that began 25 myr ago. However, that timing does not explain why the Caples and Rakaia terranes, plus their Otago Schist and Marlborough Schist equivalents (which lie between the two), were not also squeezed and narrowed.

The second possibility is that narrowing occurred earlier, during accretion of the Caples-Rakaia composite Terrane. The reconstructed Caples-Rakaia Terrane, pre-Alpine Fault, has a triangular map shape, widest in the central part of New Zealand (Box 5.3). A triangular terrane may dock point-first, and hence the point may act as an 'indenter', locally squeezing the terranes it is docking with, until the full length of the triangle is pressed up against the dock. This would have happened around 170 myr ago, producing the Patuki Mélange as it indented the Dun Mountain rocks. However, the fact that the Brook Street Terrane was squeezed when it was already accreted to the Takaka Terrane, and when it was in part intruded and bolstered by the early Median Batholith, and the fact that the Murihiku Terrane did not dock with the Brook Street until around 130 myr ago, throws doubt on this possibility.

The jury is still out on all of these questions.

Mobility of Gondwana

One final point needs to be made in this chapter. As noted earlier, if we cross the ocean to Antarctica and Australia the terrane story continues backwards in time, as this is where the older terranes of Gondwana reside. Furthermore, we cannot assume that Gondwana was standing still while all this terrane accretion was taking place. In company with the surrounding plates, Gondwana was performing a stately dance around the globe the whole time. Thus we have terranes that originated in all environments, from the tropics to the sub-Antarctic, while Gondwana itself bears glacial cover deposits of a wide range of ages (500–250 myr). These are found in different parts of the super-continent (now Africa, South America and India, in addition to Australia and Antarctica, but not New Zealand), as its dance took it around and over the South Pole. In a geological timeframe, the mobility of the earth is astonishing. The changing map of New Zealand over the past 600 myr is depicted in Boxes 5.5 A–D.

Phase one

Gondwanan phase of terrane accretion, 505–110 million years (myr) ago (Box 5.5 A, upper map).

Phase two

Break-up of Gondwana and separation of the New Zealand continent (Zealandia). Rifting 110–83 myr ago, and Tasman Sea spreading 83–50 myr ago; note the overlap with phase three (Box 5.5 A, lower map).

Phase three

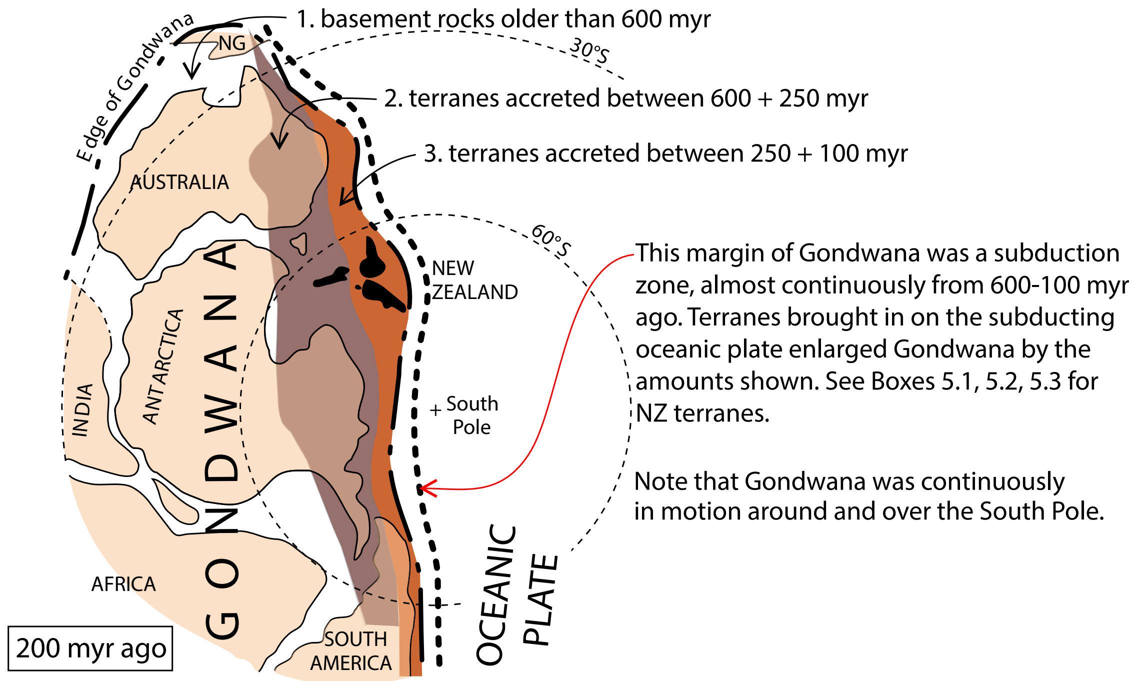
Zealandia as a passive blob in the Pacific Ocean, gently subsiding beneath the waves, 80–25 myr ago; note the overlap with phases two and four (Box 5.5 B, upper map).

Phase four

New Zealand astride the Pacific–Australian plate boundary, leading to accelerating tectonic and volcanic activity, 45 myr–present day; note the overlap with phase three (Box 5.5 B, lower map).

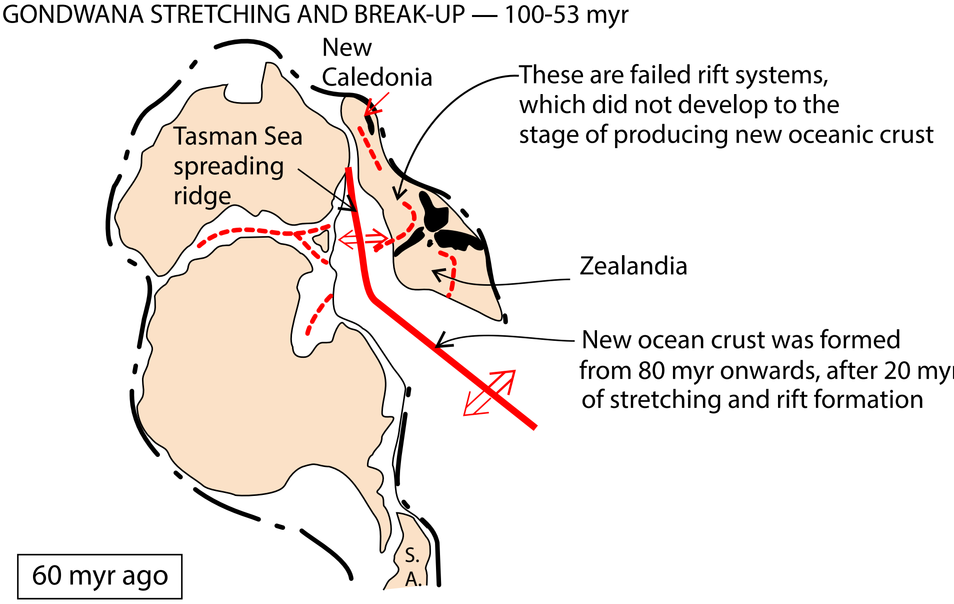
Box 5.5 A. Geological development of New Zealand, 600 to 53 myr ago.

1. GONDWANA ENLARGEMENT — 600-100 myr



By 100 myr ago, India and Africa had separated and moved away. The remaining four continents began to break apart 100 myr ago.

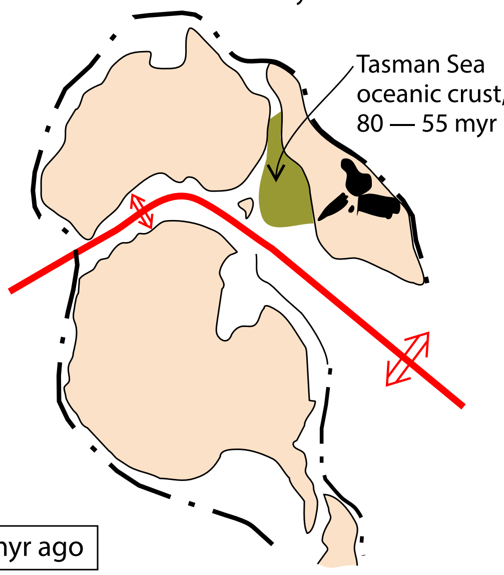
2. GONDWANA STRETCHING AND BREAK-UP — 100-53 myr



Australia and Pacific, separated at the mid-Tasman spreading ridge; Zealandia, wholly on the Pacific Plate until the failure of the Tasman spreading ridge, at which point (50 myr) a plate boundary disappears and the Australian and Pacific plates become one plate.

Box 5.5 B. Geological development of New Zealand, 53 myr ago to present day.

3. RIDGE RE-ARRANGEMENT — 53 myr



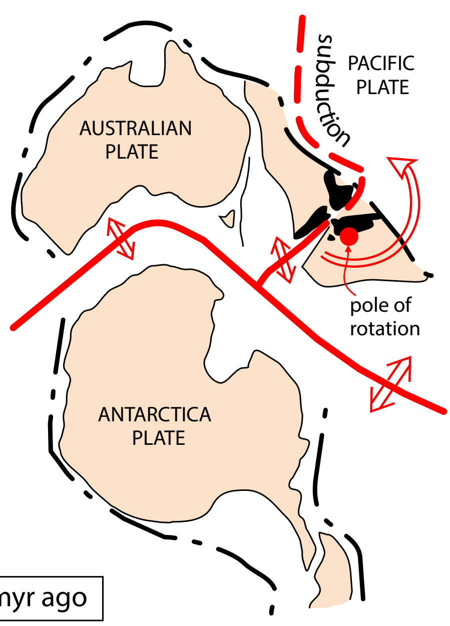
50 myr ago

At 53 myr the Tasman arm of the spreading ridge failed, and the old rift between Australia and Antarctica, south of Tasmania, was re-activated as the continuation of the the mid-Pacific spreading ridge, linking with spreading ridges in the Indian Ocean. Box 2.1 shows the present-day arrangement of ridges.

From 80 myr onwards, the stretched and thinned crust of the Zealandia mini continent gradually cooled and subsided. It had been heated during the development of the Tasman spreading ridge, 100-80 myr ago. By 30-25 myr it was largely submerged, and land areas were small and low lying. Most of it is still submerged, being only 20-25 km thick.

Note that at this time Australia and New Zealand were on the same plate. The Pacific and Australian plates were combined (50-45 myr) then they separated.

4. NEW PLATE BOUNDARY — 45 myr to present

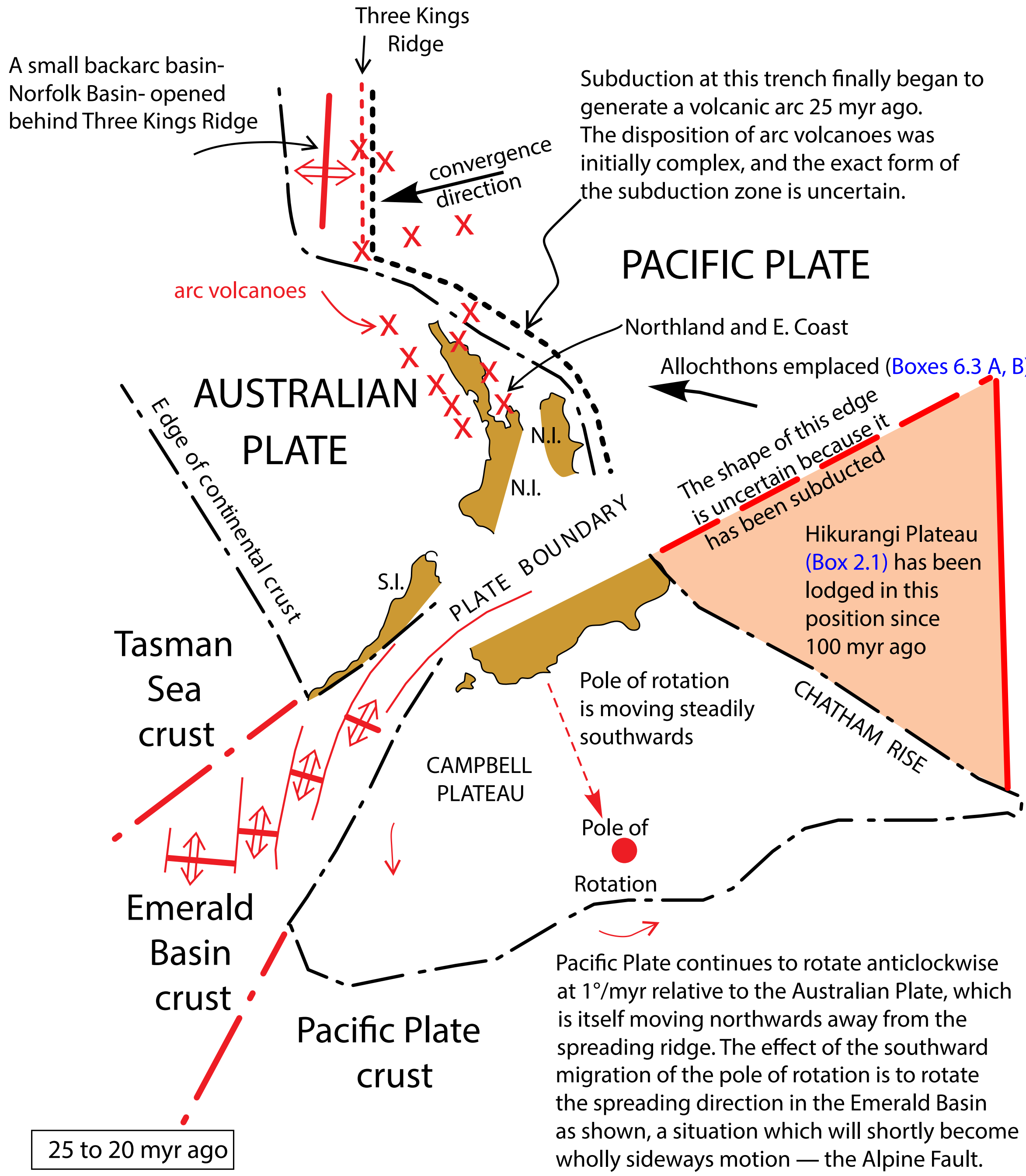


30 myr ago

Spreading continued around Antarctica, and at 45 myr a new plate boundary appeared, separating the new Australian and Pacific Plates. The Pacific Plate began to rotate anti-clockwise around a point in southern South Island at 1°/myr. The effect was to open up a spreading basin between Westland and the Campbell Plateau, while north of the rotation point ("pole of rotation") slow and oblique subduction began, too slowly to produce a volcanic arc or to uplift new land.

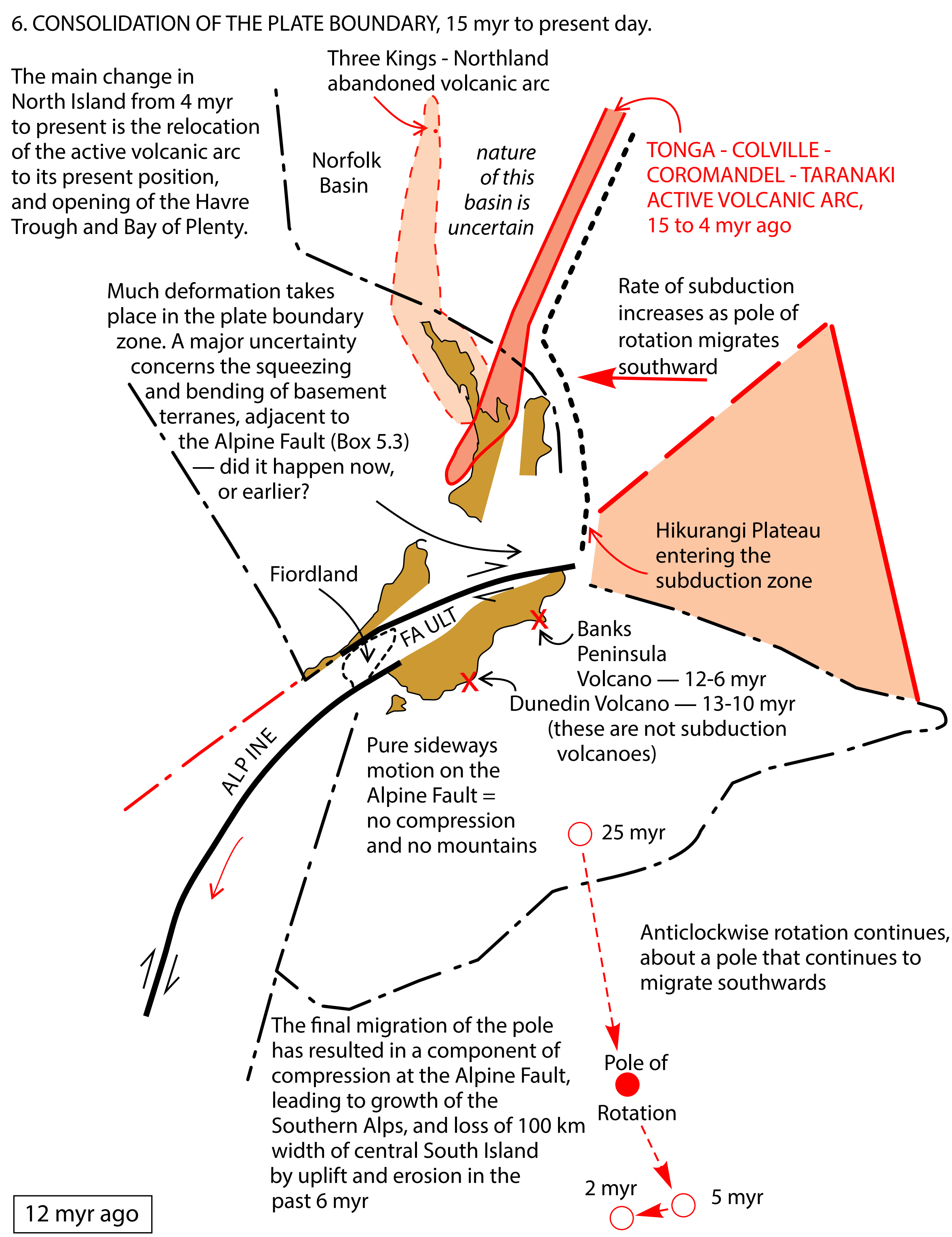
Box 5.5 C. Geological development of New Zealand, from 25 myr ago.

5. VITALISATION OF THE PLATE BOUNDARY, VOLCANOES, BASINS AND UPLIFT — AFTER 25 myr



Box 5.5 D. Geological development of New Zealand, 15 myr ago to present day.

6. CONSOLIDATION OF THE PLATE BOUNDARY, 15 myr to present day.



PART 2: REGIONAL GEOLOGY OF THE NORTH ISLAND

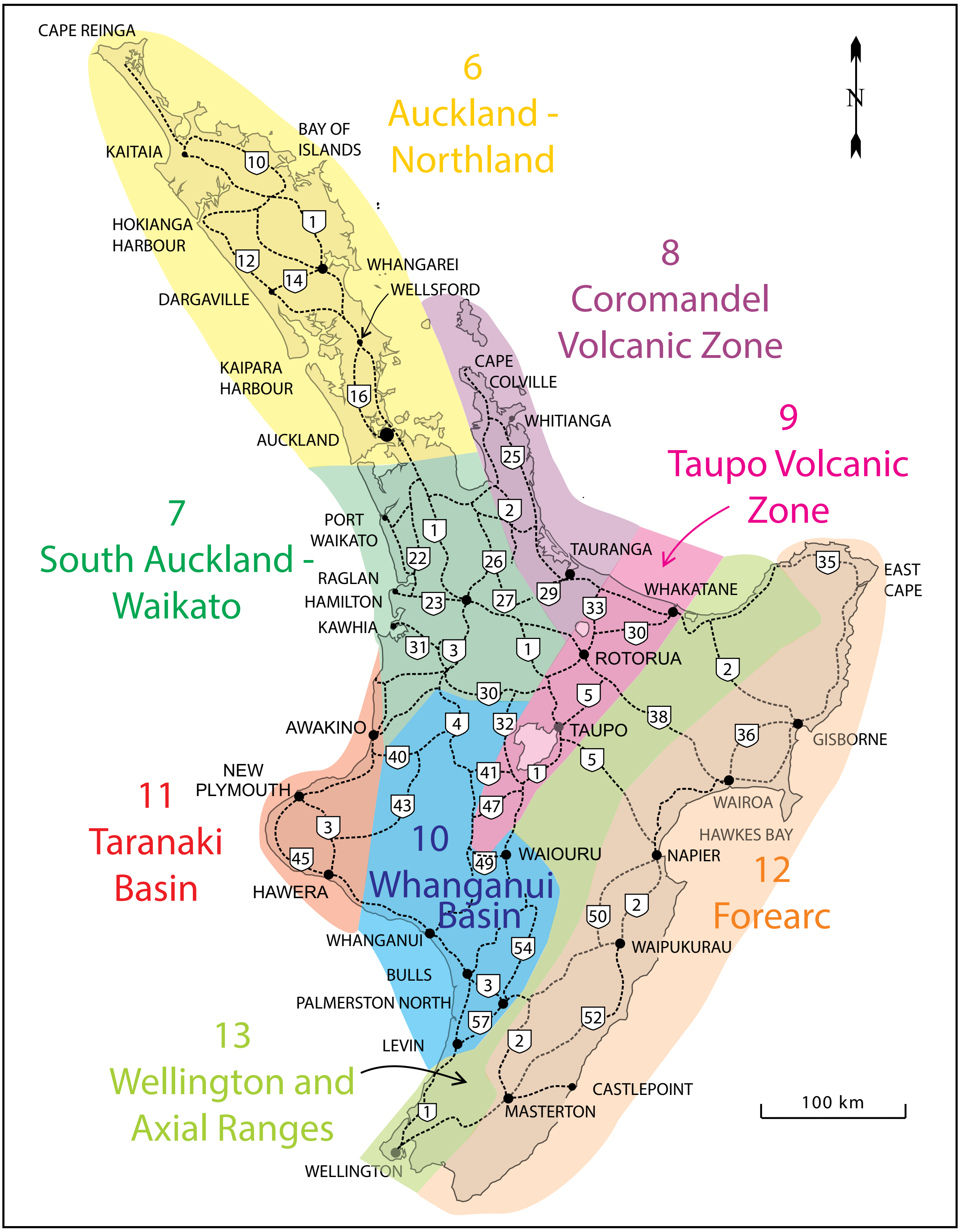


Fig. 6.0. North Island of New Zealand, showing the area covered by each chapter.

Chapter 6

Auckland and Northland

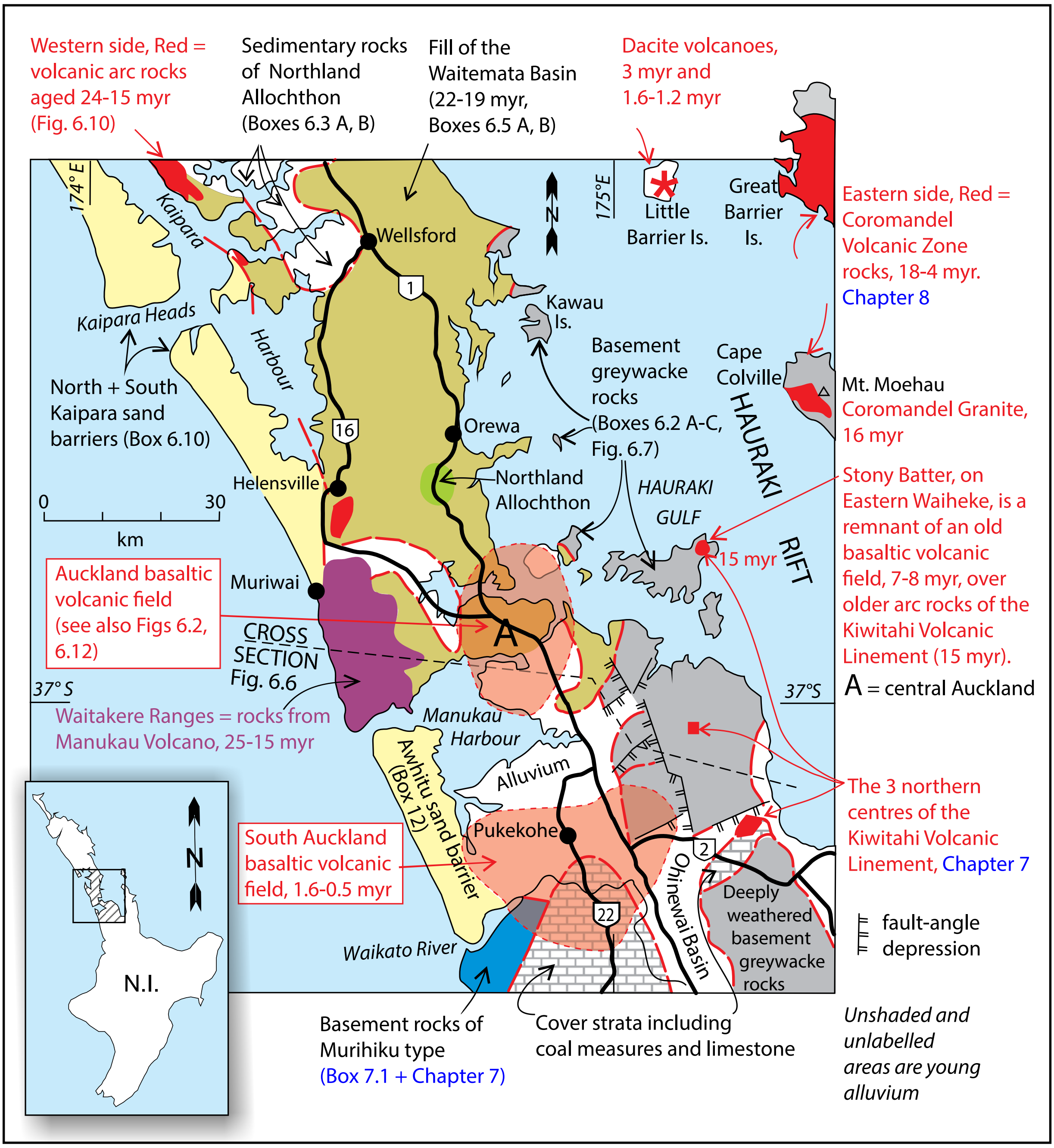


Fig. 6.1. Simplified geology of Auckland.

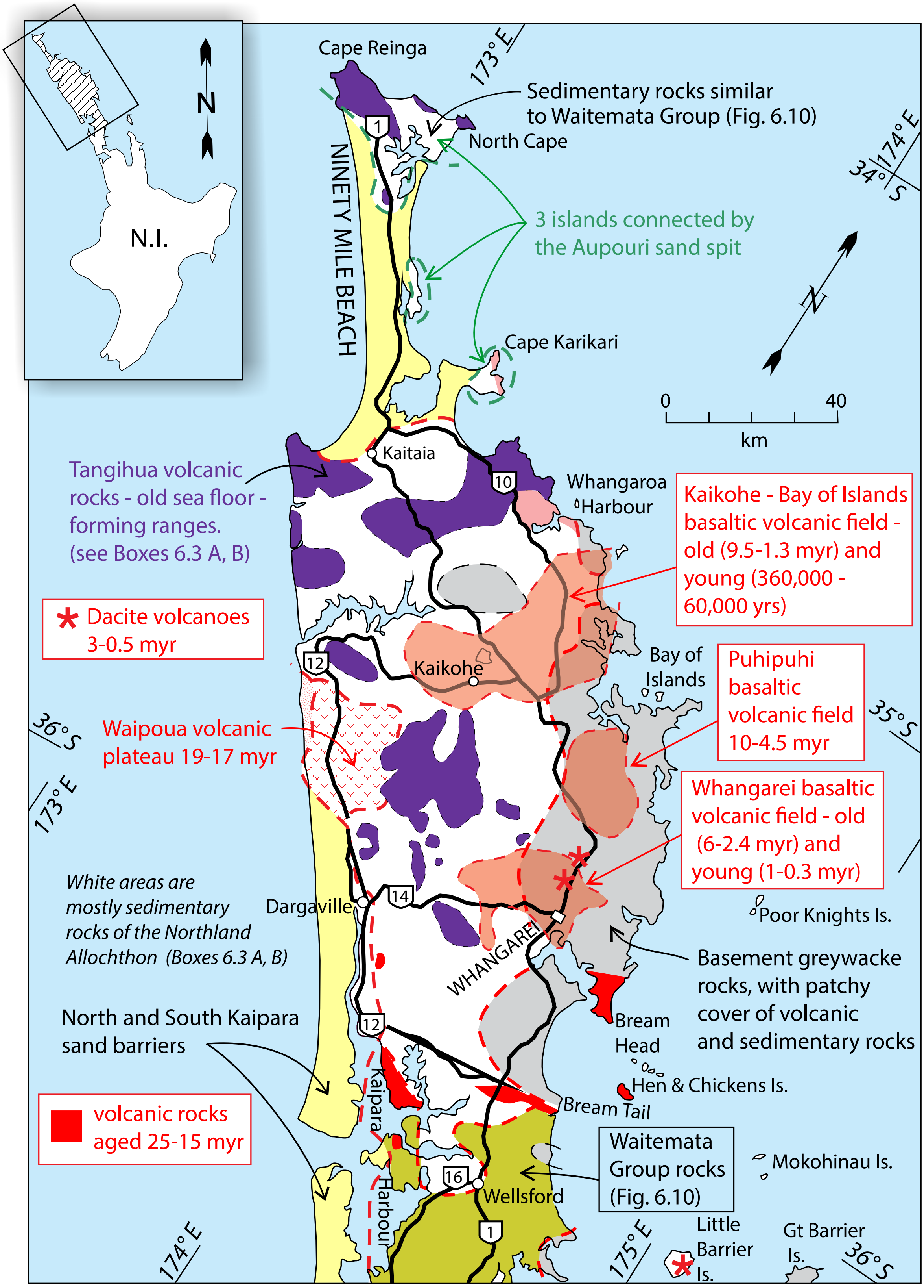


Fig. 6.2. Simplified geology of Northland.

This, the northernmost part of New Zealand, is our warm-temperate to subtropical region. Climate has a strong bearing on how we see the geology – the operative word here is 'weathering'.

Chemical Weathering

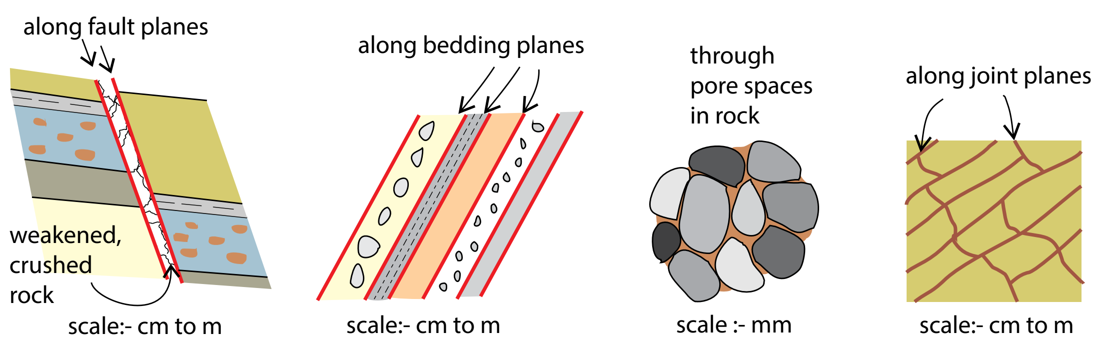
Most rocks have been buried, before being exposed at the surface as a result of erosion of the overlying rocks. In contrast, lava is extruded and crystallised at very high temperatures at the surface. Either way, some or all of the minerals in rocks were in chemical equilibrium with an environment that differed from that at the surface.

When a mineral is placed in a new chemical environment – in this case at the earth's surface – it begins to adjust and change in order to get back into equilibrium with its surroundings. This process is called chemical weathering. Most of the common rock-forming minerals change – slowly – to clay, which is the main outcome of chemical weathering. The only common mineral that is immune to this process is quartz. Boxes 6.1 A–C show this process diagrammatically.

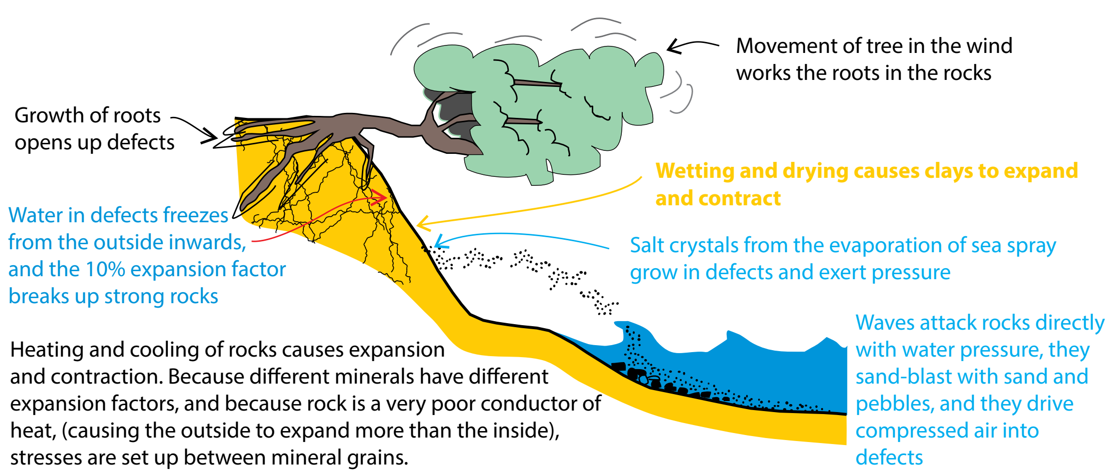
Box 6.1 A. Weathering of rocks.

Rocks break down when they are at or near the earth's surface - that is **weathering**. It is well known for its effects on old stone buildings and monuments. It weakens rocks and makes them susceptible to erosional processes, and it provides the basis for agricultural soil.

Weathering depends on water and air and plant roots gaining access to rocks by way of various **defects:**



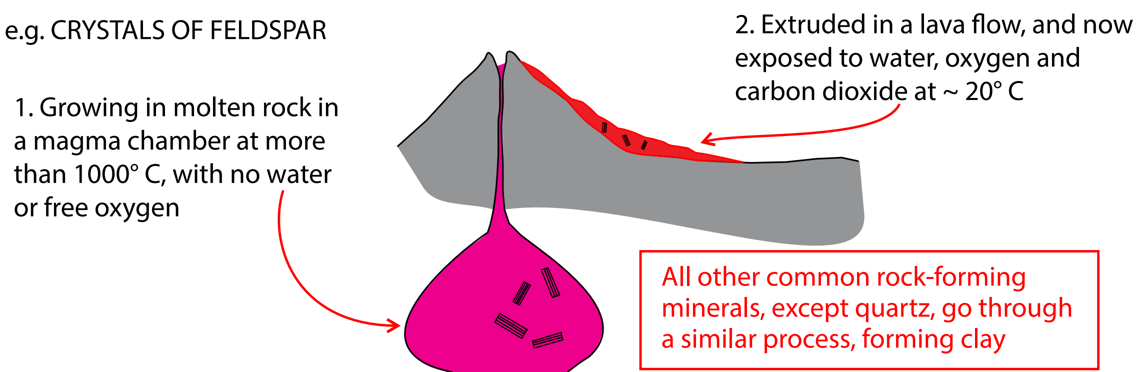
(I) Some weathering processes are purely physical —



Box 6.1 B. Chemical weathering.

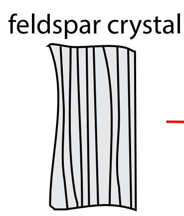
(II) However, the most important weathering process is chemical

Most rocks that are now exposed at the earth's surface either formed at depth in the crust, or have been buried to some depth in the crust.
Their minerals are "at home" with high temperatures and pressures, and a chemical environment quite different from the earth's surface. When exposed at the surface, they begin to make chemical adjustments — that is chemical weathering.



All other common rock-forming minerals, except quartz, go through a similar process, forming clay

WHAT HAPPENS -

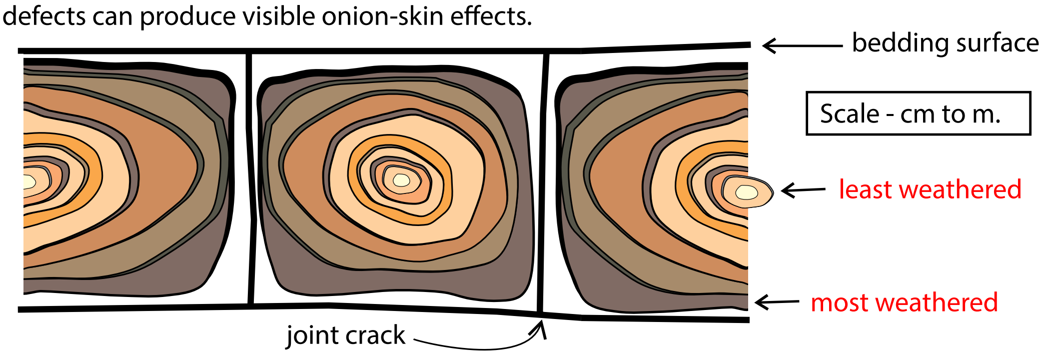


Add air and water → goes to → clay and iron oxide (rusty stains) which stay behind → dissolved silica and lime, which are removed in streams

These changes are slow - they take many centuries to complete. On reaching the sea, the dissolved silica and lime are used by marine organisms to make their shells, and some carbon dioxide has been removed from the atmosphere (in the form of carbonate, CO_3)

ONION-SKIN WEATHERING

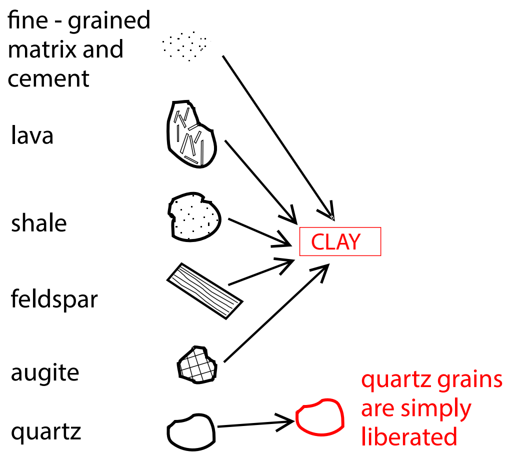
At an intermediate stage in chemical weathering, the different degrees of change working inwards from defects can produce visible onion-skin effects.



If rocks at this stage of chemical weathering are released by erosion, they quickly lose the more weathered outer "onion skins", leaving well-rounded kernels of less weathered, and therefore harder, rock. This rounding has not been produced by abrasion and transportation. In areas of volcanic rocks it is common to see round boulders of lava lying on the surface — these are normally "onion-skin" kernels.

Box 6.1 C. Weathering of greywacke rock.

Greywacke (Box 6.2 B, Fig. 6.7) is a type of sandstone. The sand grains comprise mostly fragments of lava and other fine-grained rocks, along with crystals of quartz, feldspar and "ferromagnesian" minerals like augite and hornblende. As a result of deep burial, the sand grains are welded together so firmly that when the rock is broken the fracture surface passes through the sand grains rather than round them. However, all the component sand grains except quartz are highly susceptible to chemical weathering.



Microscope view of greywacke.
The white grains are quartz.

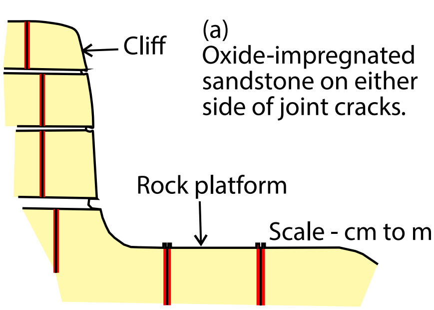
The weathering process is greatly helped by the closely-spaced defects that pervade greywacke, and by the warm-temperate, moist climate of New Zealand, particularly northern North Island. Thus in the greywacke terrains north and south of Auckland there is commonly several metres thickness of residual clay (i.e. fully weathered greywacke) between the surface soil and the rock beneath. Completely 'fresh' greywacke will be several metres further down.

Similarly, most other rocks in the North Island are overlain by a metre or more of clay, given a few tens of thousands of years for the weathering to take place. Weathering extends downwards into rock at a rate that helps to control soil erosion - the more advanced the stage of weathering, the weaker the rock and the more readily soil erosion can work (Boxes 12.5 A-C).

Dissolving of limestone by rainwater and groundwater is a special case of chemical weathering (Box 7.2 B).

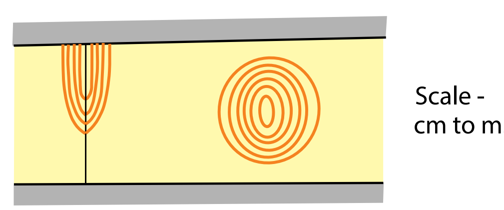
OXIDATION FEATURES

Oxidation of iron (most rocks contain a little iron) to rust (the mineral limonite) occurs commonly in sandstones. The rust forms a dark cement around the sand grains.



(a)
Oxide-impregnated sandstone on either side of joint cracks.

(b)
Leisegang-like rings, alternate high and low concentration.



These oxidation features reflect the way in which oxygen diffuses into the porous sandstone (and occasionally other kinds of porous rock) from defect cracks. The Leisegang Ring features (concentric bands of high and low concentration, in this case of brown limonite) are a poorly understood chemical phenomenon. They form in many instances where a fluid substance (e.g. potassium permanganate in solution) diffuses into a solid substance (e.g. jelly).

Chemical reactions take place faster in warmer surroundings and, in the case of the change in minerals, in a moist environment, which is why in New Zealand the chemical weathering of rocks to form clay is particularly prominent in the warm, moist north. The practical result of this is that wherever you see a potential rock exposure, in a cutting, cliff or quarry, there is normally 2–3 m (and up to 5 m or more) of clay between the topsoil and rock that is recognisably rock – though even then that rock will be partially affected by chemical weathering.

The clue as to whether you are looking at partially weathered rock or unweathered rock is colour. Most rocks contain a few per cent of iron. In unweathered rocks the iron is in a chemically reduced state and the rock is blue-grey in colour. When it comes into contact with air and water, this iron does what all iron does – it oxidises (rusts). Partially and completely weathered rocks are therefore usually coloured in shades of red, orange, yellow and brown. Quarry operators normally seek unweathered rock because that is stronger, and therefore in a quarry you will usually be able to see the downwards transition from topsoil through completely weathered rock (brown clay), partially weathered brown rock, to unweathered blue-grey rock (Fig. 6.3). Some sea cliffs show the same sequence, but natural exposures away from the coast seldom show unweathered rock, except sometimes in river beds and river cliffs.

Fig. 6.3. In Glenbervie greywacke quarry, Whangarei, it is easy to see the yellow-brown clay at the surface passing down through brown weathered rock to fresh blue-grey rock at depth. Photographer Bruce Hayward.

Rock defects

Air and water also gain access to rocks along defects – cracks and discontinuities – which is where most chemical weathering takes place. The commonest of these are bedding planes in layered sedimentary rocks, and joint cracks in all sorts of rocks (Fig. 6.4). Joint cracks generally occur at right angles to sedimentary layering. They are, in part, a result of the expansion of rocks that takes place as they get nearer the surface and the confining weight of the overlying rocks is gradually removed as they are eroded.

Chemical weathering typically starts alongside joint cracks and bedding planes, and works its way inwards into the joint-bounded block of rock – generally 1 m or less in size. Weathering works fastest near the corners of this block, and the mid-point stage in the weathering sequence is sometimes a rock showing multi-layered onion-skin weathering, each 'onion' being a joint-bounded block (Box 6.1 B).

Fig. 6.4. This weathered greywacke rock on Rotoroa Island, Hauraki Gulf, is cut by numerous near-vertical joints (cracks). Chemical weathering has produced the orange iron-oxide bands along each joint. Photographer Bruce Hayward.

Liesegang colour rings

An interesting phenomenon sometimes seen as a result of chemical weathering is Liesegang colour bands. Stains produced when one substance diffuses into another substance commonly show as alternating bands of dark and light colour. These bands are called Liesegang rings after the German photographic chemist who first described them. The diffusion of oxygen and water from defects into porous rock quite often produces Liesegang rings of red-orange-brown iron oxide staining. The systems can be centimetres or metres across, they are commonly concentric circles and swirls, and they cut across (and commonly obscure) any original structures in the rocks (Box 6.1 C, Fig. 6.5).

Physical Weathering

Aside from chemical weathering, rocks are also subject to weathering and break-up by various physical processes. These include the expansion of tree roots and ice in cracks in the rock, wave attack, sand-blasting, landslides, rock falls and so on (Box 6.1 A).

Fig.6.5. Liesegang rings of rust-coloured iron oxide staining in weathered andesite on Pudding Island, Otago Harbour. Photo width 3 m. Photographer Nick Mortimer.

Geological History of Auckland and Northland

Auckland and Northland contain rocks representing all four phases of New Zealand's evolution. From the mouth of the Waikato River northwards, the country is tilted geologically to the southwest. The effect of this is that, following lots of erosion over the years, the lowest, oldest rocks in the pile (corresponding to phase one) are exposed at the surface along the eastern seaboard, particularly on the coast northwards from Whangarei Heads (Fig. 6.6). These are greywacke-suite rocks belonging to the Waipapa Terrane (Boxes 6.2 A–C, Fig. 6.7) and they slope away to the southwest underneath the younger rocks. At depth they are juxtaposed against other basement terranes such as the Murihiku Terrane, as seen at the surface further south (Chapter 7).

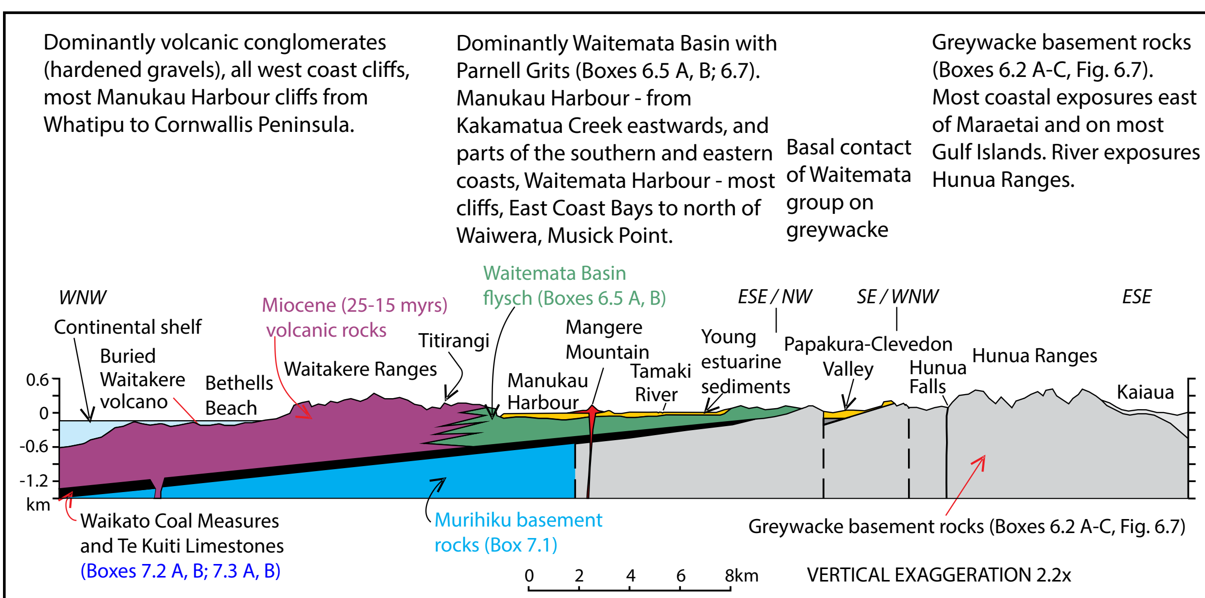
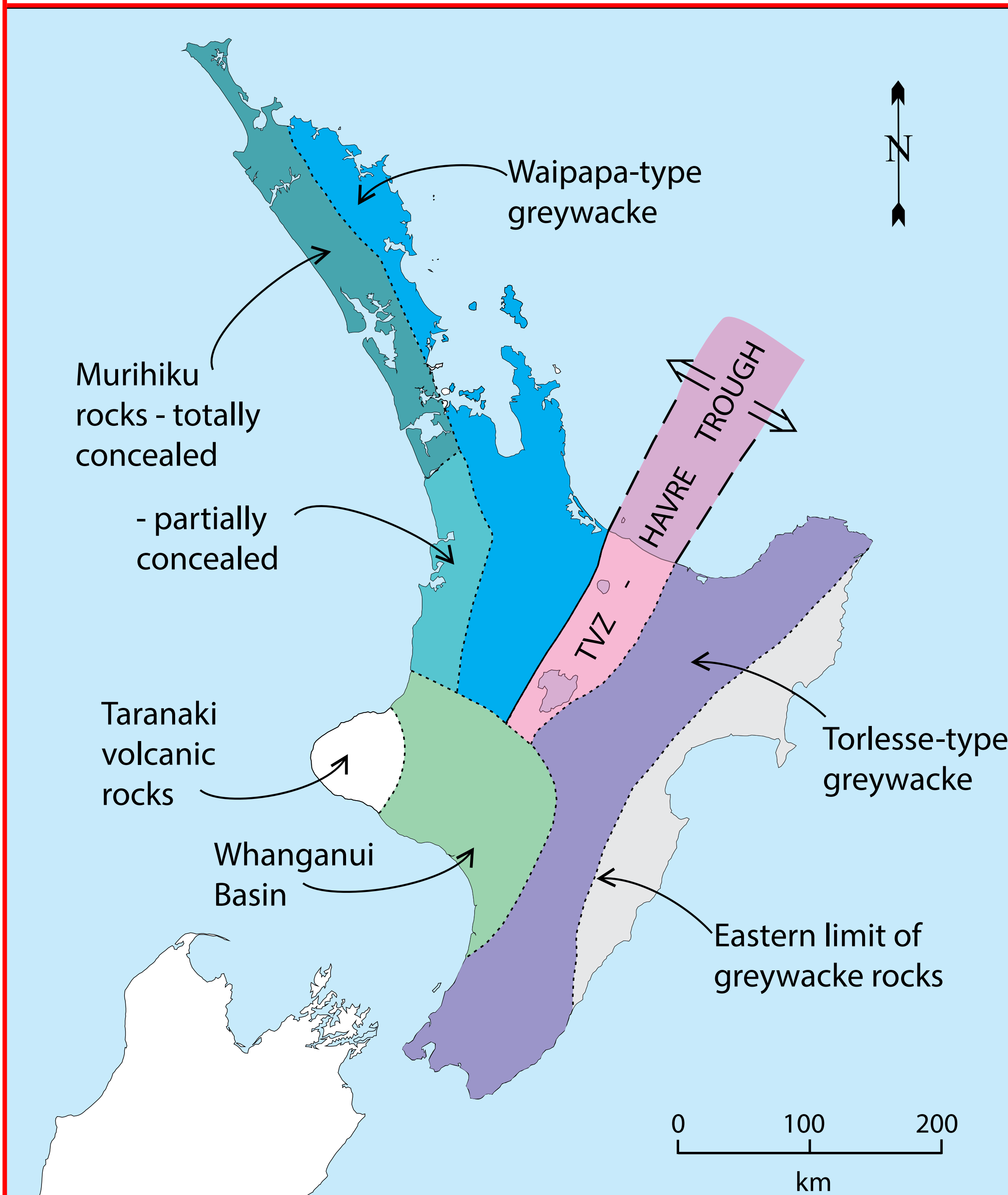


Fig. 6.6. Generalised cross section through Auckland.

Box 6.2 A. Distribution of North Island greywacke rocks.
See [Box 7.1] for the associated Murihiku rocks.
The difference between Waipapa-type and Torlesse-type greywackes lies in details of the mineral types they contain. The areas of greywacke and Murihiku rocks shown here are partially concealed beneath patches of younger rocks.
TVZ = Taupo Volcanic Zone.
See also the information box on the following page (Box 6.2 B).



Box 6.2 B. The greywacke association – or simply "the greywackes"
– to accompany Box 6.2 A.

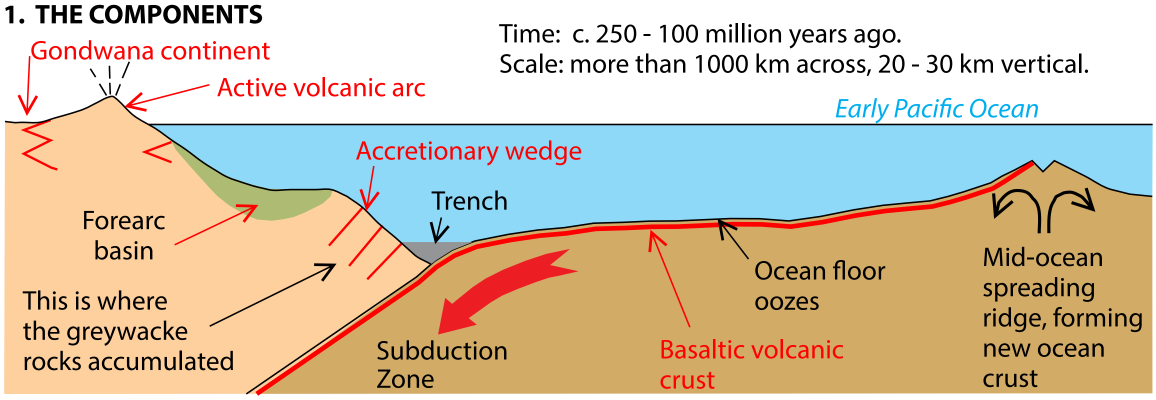
These rocks underlie much of New Zealand, to a depth of at least 10 kilometres. There is a huge volume - many thousands of cubic kilometres. They are between 250 million and 100 million years old, and in many places are covered by younger rocks. The diagram explains how the rocks formed, in an 'accretionary wedge' above a subduction zone, at the margin of the old supercontinent, Gondwana (an accretionary wedge used to be called a 'geosyncline'). Actually, over the long time period there were several accretionary wedges, giving rise to the different 'terranes' that are recognised in the **greywackes** (Box 5.3). It is called a rock association because it includes two quite different sets of rocks, **trench sediments** (greywacke sandstones and **argillites**) and **oceanic crustal rocks** (**basaltic pillow lavas, cherts**, rare limestones).

Greywacke rocks originated as layers of sand and mud in a deep-sea trench, while oceanic rocks originated far away from the trench at a mid-ocean spreading ridge. The following diagram (Box 6.2 C) explains how the two rock groups were brought together and mingled in the accretionary wedge. The colour grey dominates these rocks. It represents sandstones (originally layers of sand), hard argillites (originally layers of mud) or bands of highly sheared and mashed up rock (mélange, a French word, from the cooking vocabulary, meaning mixture). Why not simply call them sandstones and mudstones? Well one can, but the significance of greywacke (from the Harz Mountains in Germany, grauwacken was an old miners' term) relates to the hardness. These are very hard rocks. If you break a piece of unweathered greywacke sandstone, the rock is so tightly bound together that the fracture passes through quartz grains, not around them, and quartz is a very tough mineral. Box 6.1 C has a microscope picture of greywacke. How did an originally loose sand come to be so hard? The answer lies inside the accretionary wedge, where there was high pressure and high temperature in combination with water. The rocks were tightly folded, and new minerals like chlorite grew and bound the rocks, particularly the sandstones, very tightly. These were early stages in a process of change ('metamorphism') which led ultimately to the South Island schists (Box 15.3 B). There is only a small area of schist at the surface in the North Island. Another consequence of hot water circulating in the accretionary wedge is an abundance of white veins in cracks in the rocks, containing quartz, calcite or zeolite minerals which precipitated from the hot water. More than one of these minerals may be present in a vein, along with other minerals like gold. Veins range in width from millimetres to metres, and cut across other structures.

The two most typical components of the ocean crust association are dark basaltic pillow lava (Box 6.4 A) and red chert which overlies it. The chert is typically in prominent layers 5 to 10 cm thick, and brick red in colour. It is made of silica of a high degree of purity, which derives from the shells of microscopic marine organisms called radiolaria and diatoms. It was changed to chert by the same hardening processes that affected the greywacke sandstones, and the red colour is caused by a small quantity of iron oxide. Chert layers are commonly bent into metre-sized folds. Greywacke is usually folded and sheared, often complexly. Most fold structures formed in the accretionary wedge, but some shearing may record more recent tectonic activity. Maps showing the distribution of basement terranes in the South Island can be found in Boxes 5.3, 15.2 and Figs 16.1 A, 16.1 B, 17.1, 17.2, 20.2, 22.1.

Box 6.2 C. Origin of New Zealand greywacke rocks.
Idealised section through an off-scraped portion of ocean crust overlain by oceanic and trench sediment. The trench sediment may be much younger than the ocean rocks. See Box 13.1 A for a field example of this sequence.

1. THE COMPONENTS



This is the standard Plate Tectonic model — compare Boxes 2.2, 2.3

2. THE ACCRETIONARY WEDGE.

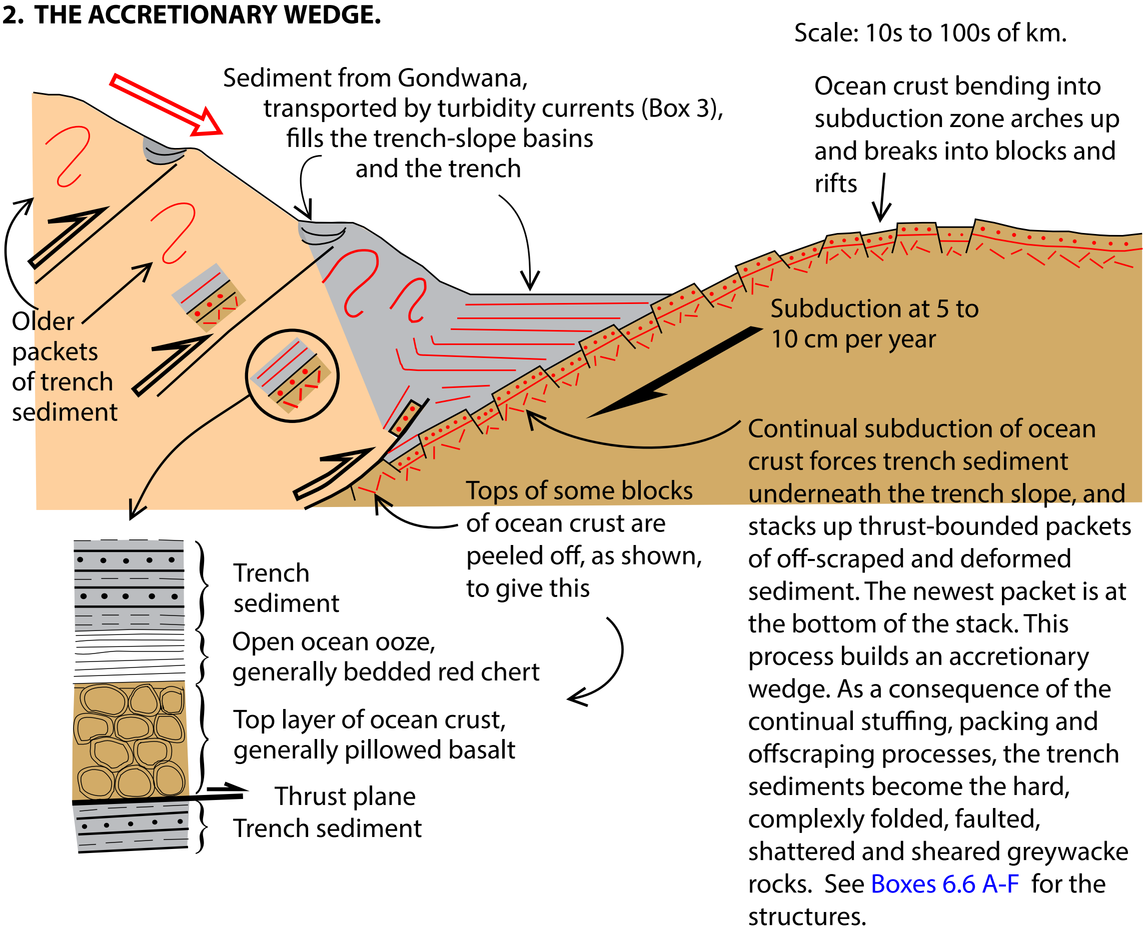


Fig. 6.7. Typical greywacke (indurated sandstone) of the Waipapa Terrane that outcrops along much of the east coast of Northland and Auckland. The rock is cut by numerous quartz and zeolite veinlets. Rakino Island, Hauraki Gulf.
Photographer Bruce Hayward.

Rifting of Gondwana

Phase two of the country's geological history, the rifting phase, is represented by the northern extension of the Taranaki rift basin. This lies entirely offshore, extending almost to Cape Reinga, and furthermore is completely buried underneath younger rocks and sediments of the western continental shelf. However, its oil and gas potential makes it an attractive target for oil exploration, and it has been the subject of intensive geophysical exploration. At the time of writing (2009), exploratory drill holes in the region, both onshore and offshore, have been few and unsuccessful. However, the spin-off from the geophysical work (chiefly comprising seismic reflection profiles) in terms of new geological knowledge has been considerable. In particular, the profiles showed that the Northland Allochthon extends well offshore underneath the continental shelf, that it is subdivided internally into a succession of overlapping thrust slices, and that it is overlain by a thick, seawards-tapering wedge of layered sediments building the continental shelf.

Passive subsidence of New Zealand

Phase three rocks, from the passive subsidence era, are seen at the surface around Whangarei. Elsewhere in the region, they are mostly either buried or have been eroded away. They show the classic subsidence sequence of strata. At the bottom, resting on a former land surface of greywacke, there are sandstones and mudstones of coastal swamps. These contain coal seams (originally peat layers), which were once worked at Kamo and Hikurangi, north of Whangarei. The coastal swamp deposits are overlain by shallow marine deposits. Initially these were well-worn quartz-rich sands, sometimes containing the green mineral glauconite, which actually grows on shallow sea beds. The sands are overlain by widespread shell limestones, around 30 myr old, as the supply of sand from the low-lying, shrinking land areas dwindled. In places where humps in the old land surface precluded the development of swamps, limestone may rest directly on greywacke, as it does between Waipu Cove and Langs Beach near Bream Tail (Figs 6.2, 6.8).

Still in phase three, and talking countrywide now, the sedimentary deposits comprise two distinct packages. Those like the Whangarei rocks were deposited on the top platform surface of the subsiding continent, in swamps and shallow water. However, all continents have edges, connected to deep water by continental slopes. Considerable amounts of sediment accumulate on these slopes and are quite different in character from the shelf/platform deposits.

Like the platform deposits, there is a characteristic sequence of sediment types in the slope deposits. The oldest of these strata formed during the time when there was still plenty of sediment being eroded from the landmass (initially this was still Gondwana, 100 myr ago). These deposits consisted of gravity flow material, alternating sandstones, and mudstones and rare gravels of flysch character (Boxes 6.5 A, B). As the supply of sediment dwindled, the slope sequence remained in gravity flow mode, but around 55 myr ago the sand supply changed in character to quartz-rich sands containing green glauconite – this was continental shelf/platform sand redeposited down

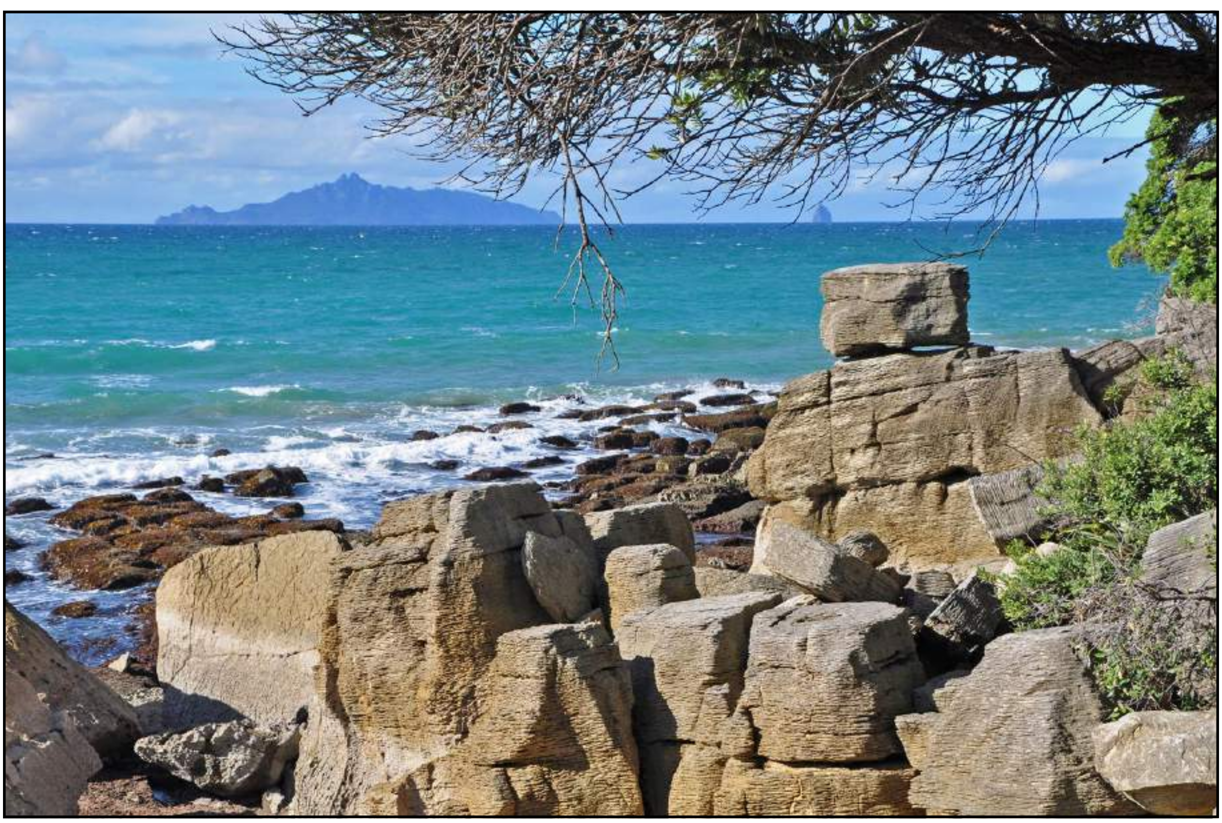


Fig. 6.8. Thirty-million year old, recrystallised shelly limestone occurs along the coast between Waipu Cove and Langs Beach, east Northland. Photographer Bruce Hayward.

the slope into deeper water. Finally, that supply fizzled out and deep-water calcareous ooze was deposited, represented now by white, very fine-grained, slightly muddy limestone. Chalk is an appropriate name here.

We find the upper two-thirds of this sequence all round the New Zealand continent. The lower third was deposited only on the original continental slope of Gondwana, before the Tasman Sea opened, along what is now the eastern seaboard of New Zealand.

All types of continental slope deposits of this era occur in Northland, but the interesting thing is that they have all been displaced from their place of origin. They now comprise a rock formation known as the Northland Allochthon (Chapter 3). This feature looms large in the surface geology of Northland, but it belongs to phase four of New Zealand's development (see below).

The invigoration of the plate boundary

Northland and Auckland contain the country's fullest and richest record of phase four of our geological development. There is quite a lot to cover. As noted in Chapter 3, the first 20 myr of development of the Pacific–Australian plate boundary (45–25 myr ago) are largely cryptic. The only apparent sign of its influence in Northland is that the rather narrow continental platform (which continues northwestward under the sea as the Norfolk Ridge) on which the 30 myr limestones were accumulating began to subside in the west, and limestone of a deeper character was deposited.

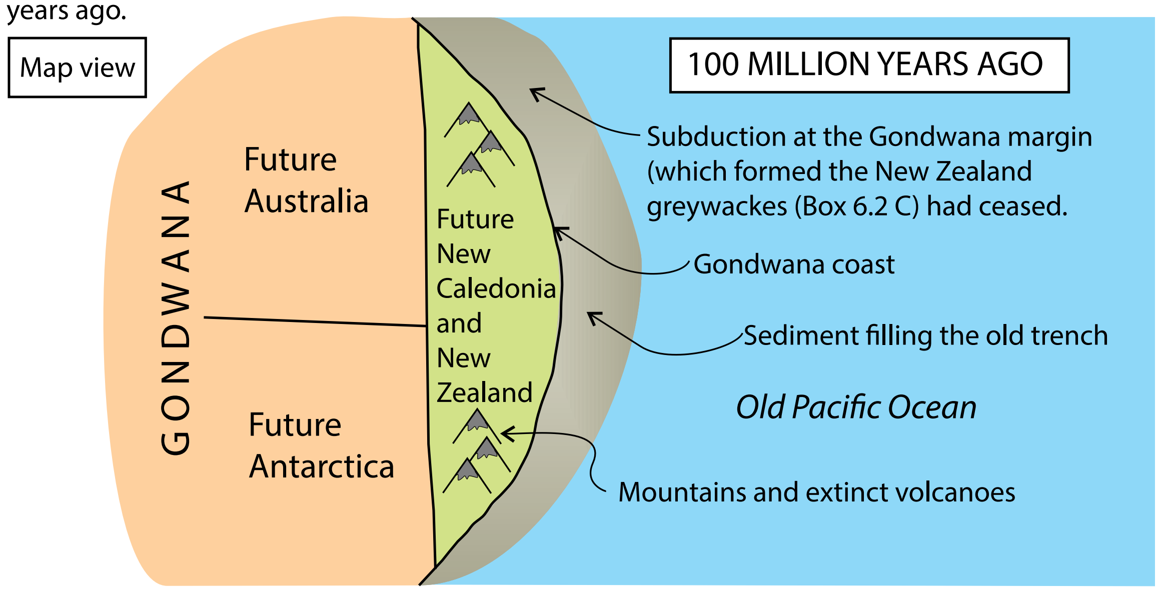
Northland and East Coast allochthons

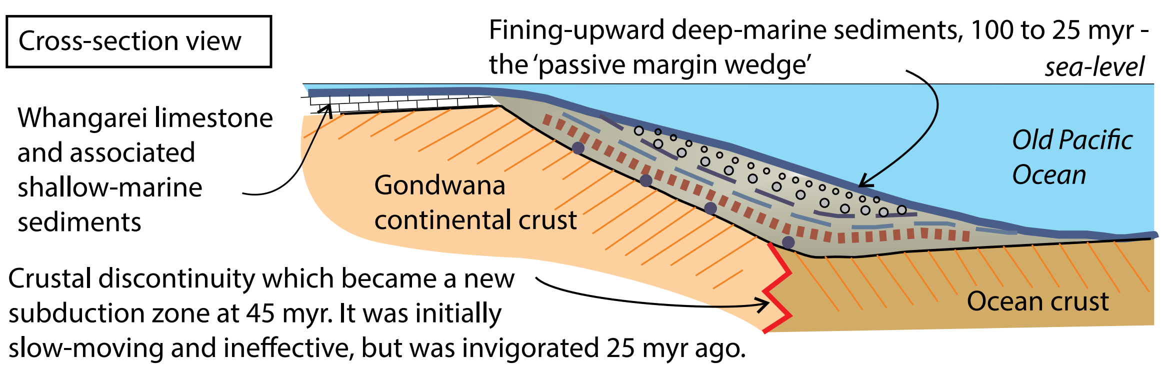
There was a subtle change in motion of the Pacific Plate 25 myr ago. It is recorded in a slight bend in volcanic hotspot chains throughout the Pacific Ocean, and in concert with the southwards shift of the plate's pole of rotation, it caused the plate boundary to roar into life – literally. There was an immediate pulse of subduction underneath the northeastern flank of Northland/Coromandel, including East Cape (the Bay of Plenty had not yet opened). The passive-margin sediment wedge ('passive' because there had been no subduction beneath it before 25 myr) making up the continental slope was strongly squeezed and deformed. If you squeeze deformable material in a long vice with no way out of the bottom, the only way for it to go is upwards. This is what happened to the passive-margin wedge. It was lifted up sufficiently high for it to become gravitationally unstable, and it began to slide southwestwards onto both Northland and East Cape, forming the Northland and East Coast allochthons. At this stage the region had subsided and was under the sea (Box 6.3 A, B).

Emplacement of the allochthons was spread over a couple of million years. The deformed wedge tended to slide off as successive slices. Thus the first slice, which tended to include the topmost, youngest sediments of the wedge, was later overtaken and overridden by the second slice, containing older rocks. The process went on like this until four or so slices had accumulated. The result was an interesting reversal of stratigraphic order, with the slice containing the younger rocks at the bottom of the stack of slices, and the older rocks in the topmost slice. We discussed another mechanism that achieves a similar reversal in the section on accretionary wedge terranes in Chapter 5.

Box 6.3 A. The Northland and East Coast Allochthon.

ALLOCHTHON is a body of rock that has been moved a significant distance from its place of formation by tectonic forces. The opposite term, for rocks still in their place of formation, is autochthon.
Most of the rocks in Northland and the Raukumara Peninsula that are aged between 100 and 25 million years were displaced from their place of formation by a major tectonic event 25 million years ago.

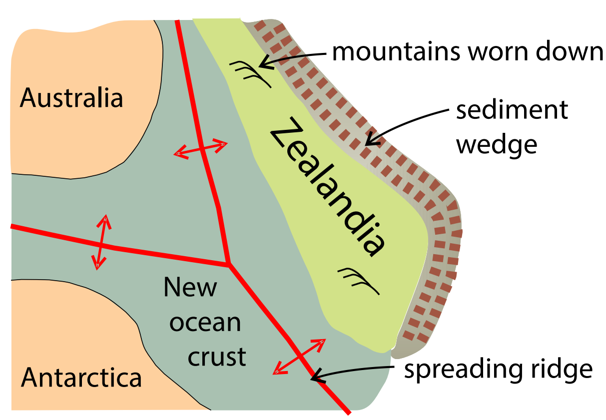




Map view

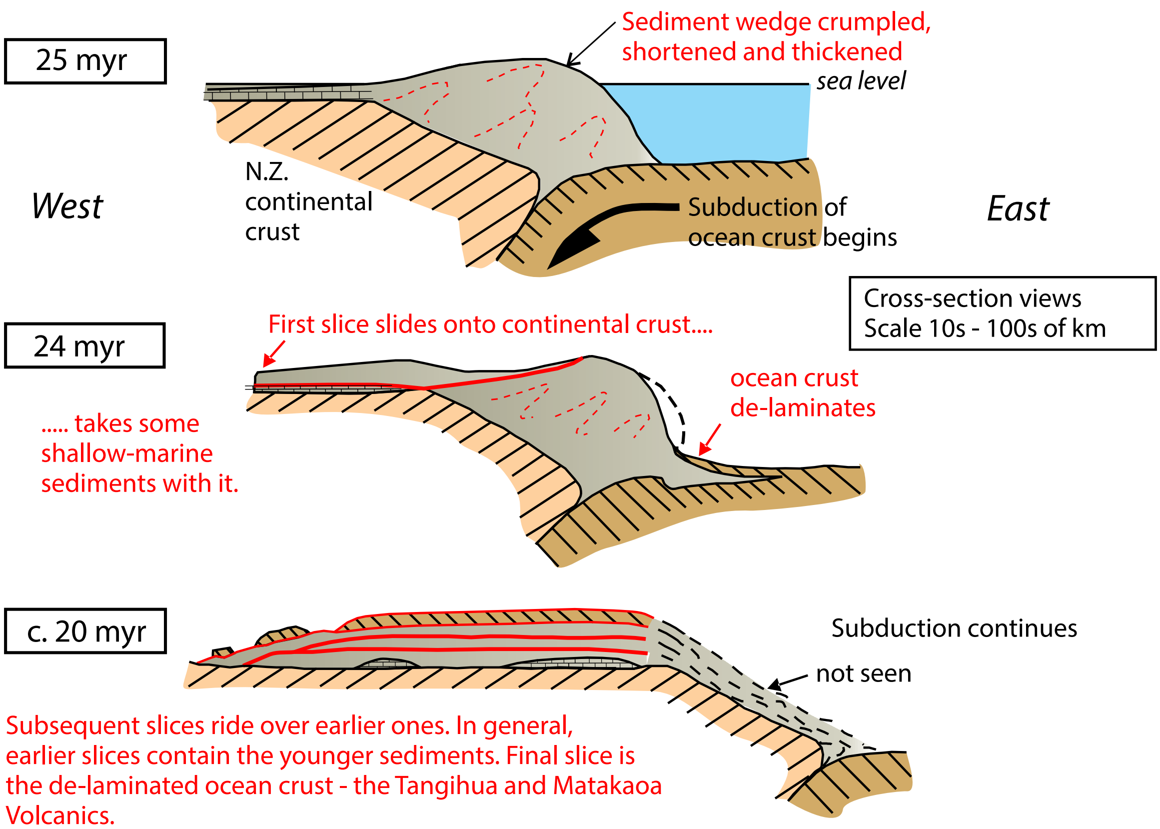
About 70 myr ago

While the passive margin wedge was accumulating, New Zealand/New Caledonia, (= Zealandia), Australia and Antarctica were separating by sea floor spreading (Boxes 5.5 A, B)



(continued below)

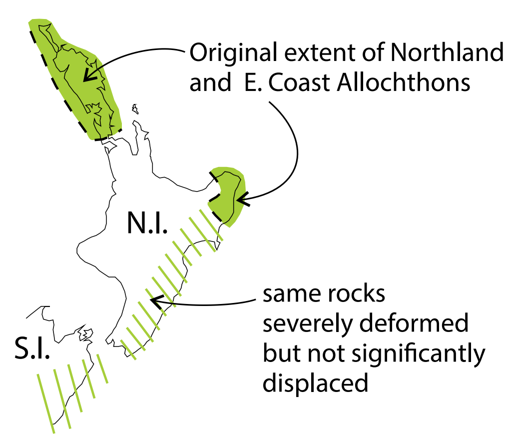
Box 6.3 B. The Northland and East Coast Allochthon *(continued from previous page).*



At 20 myr these rocks now comprise the Northland and East Coast Allochthons. They are severely deformed. Why two allochthons? There was originally only one. See "Origin of Bay of Plenty" (Box 12.3)
The scenario sketched above held from North Cape/Three Kings Islands as far south as Gisborne. South of Gisborne, as far as Marlborough, the old sedimentary wedge has been continually compressed, deformed and thickened for the past 25 myr, but it never got thick enough to slide onto the adjacent North Island.

NOTE:
1. drawn on present-day geography

2. a large part of these rocks has been either eroded away or covered by younger rocks
3. there is a large exaggeration of vertical scale



As you can imagine, the successive processes of squeezing, uplifting and then sliding produced a highly complex arrangement of rocks. There was also a large amount of loose rubble in the system. Geologists didn't realise that these rocks were allochthonous until well into the 1970s, because at that time (just 10 years after the coming together of plate tectonic theory in the late 1960s) they were only just beginning to comprehend the huge scale on which such processes could take place. The lack of understanding was compounded by the difficulty of making geological maps in Northland, owing to the soft nature of most of these rocks, the deep chemical weathering that renders shallow exposures practically useless for the mapping geologist, and the apparently random nature of relationships between adjacent bodies of rock. When travelling around the outcrops of the Northland Allochthon (most of the unshaded areas on Figs 6.1 and 6.2), the most obvious unit one sees is the white chalky limestone, because it is exposed in many small agricultural lime quarries. It is also exposed in the very large Portland quarry near Whangarei – it makes excellent raw material for cement.

The Tangihua Volcanic slice

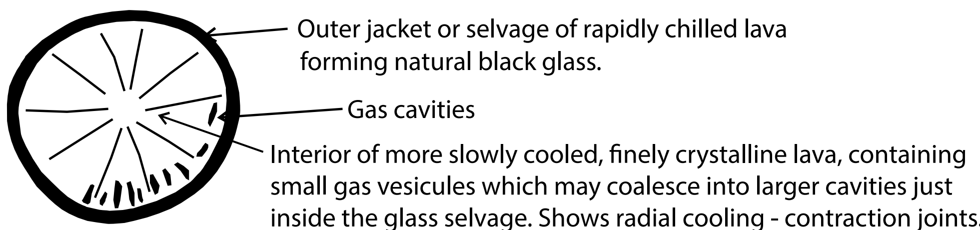
The final chapter in the story of the Northland Allochthon was the emplacement of one last slice, on top of all the others, right across Northland. This one is much more puzzling. It consists of a relatively thin layer (no more than 1.5 km thick) of volcanic rocks that were originally oceanic crust. Somehow this slice of oceanic crust was pushed up (obducted) onto Northland instead of being subducted in the usual way. Obduction happens from time to time, but usually involves a thicker slice of oceanic crust. The rocks that make up this layer are called Tangihua Volcanics, and they form most of the many hill ranges in northern Northland (Tangihua Range, Maungataniwha Range, etc.) (Fig. 6.9).

Fig. 6.9. The Tangihua Range, south of Whangarei, is made of eroded oceanic crust (Tangihua Volcanics) that was pushed up and onto Northland about 23 myrs ago. Photographer Bruce Hayward.

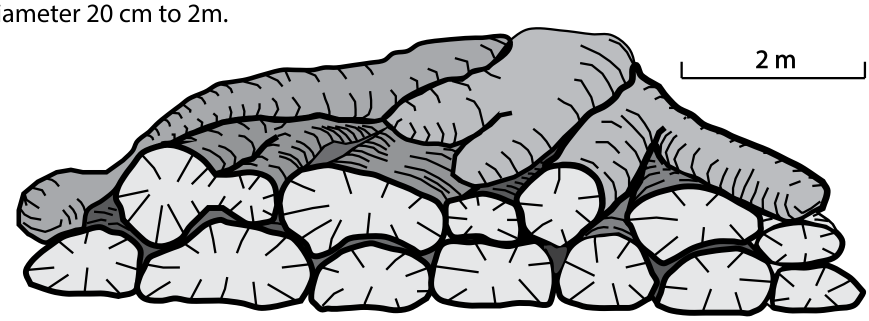
The Tangihua Volcanics present us with one of Northland's biggest geological puzzles. Most exposures of these dark basic lavas are singularly uninformative. The rocks are internally sheared and slightly metamorphosed, and no original structures can be seen, except perhaps pillow lavas (Boxes 6.4 A–C). A few rock bodies contain old fossils, dating from around 80–55 myr ago (the age of formation of the Tasman Sea). In recent years an accelerating pace of research has seen many more

Box 6.4 A. Pillow lavas *(continued on following page).*

Pillow lavas are sometimes formed when lava flows erupt under water.

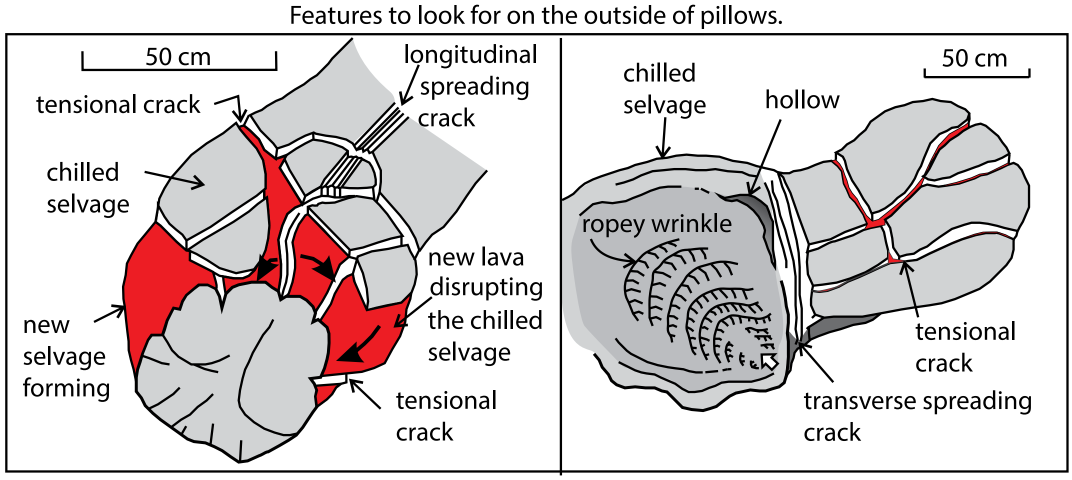


Typical pillow in cross section, diameter 20 cm to 2m.



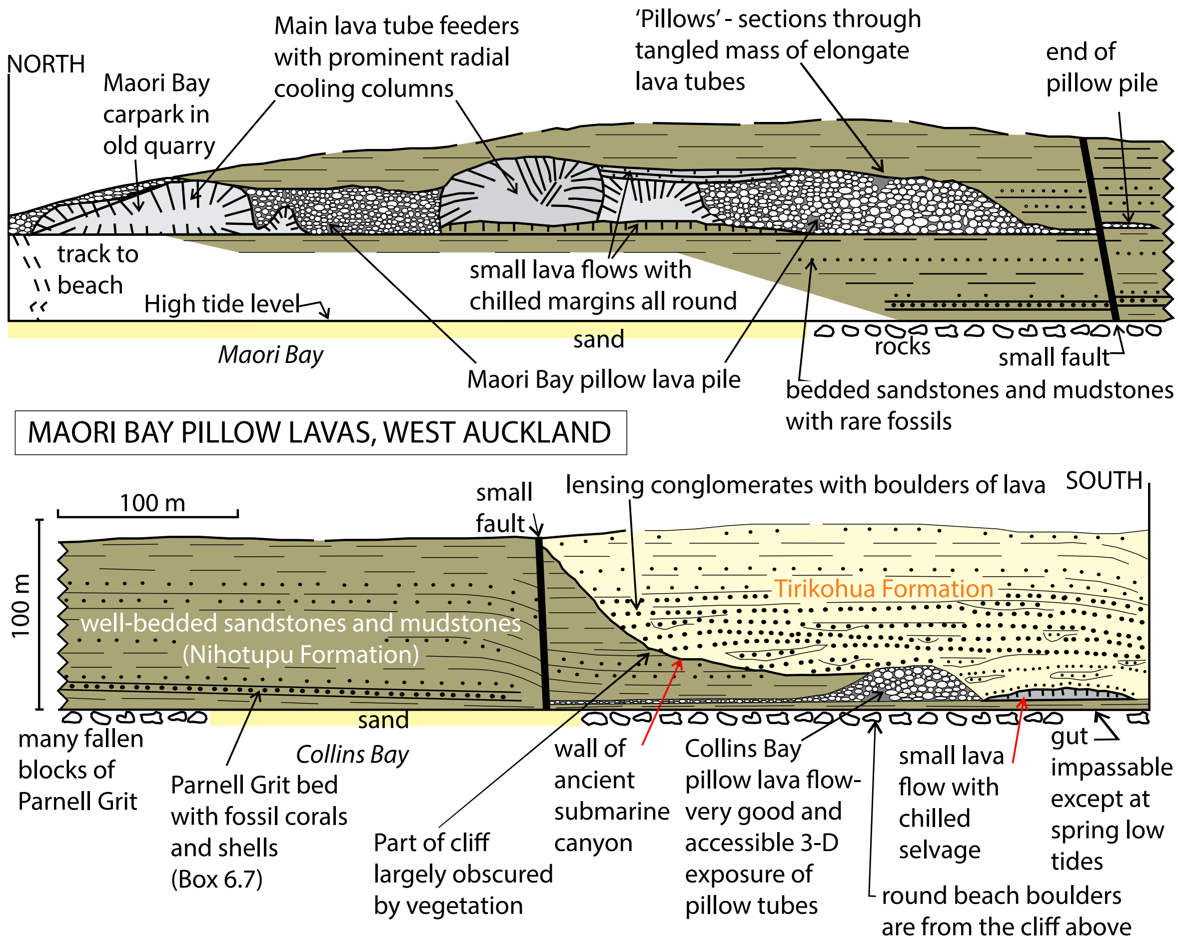
Three-dimensional view showing the typical elongate tubular shape of pillows, and the way pillows mould themselves around underlying pillows.
Molten lava continues to flow through the tube created by the chilled selvage, and is extruded at the end of the tube, causing the pillow to get longer. A pillow may swell, break the selvage, and form a new one, causing multiple selvages.

Features to look for on the outside of pillows.



The South Island's best known pillow lavas and associated volcanic deposits are at Boatmans Harbour, Oamaru.

Box 6.4 B. Pillow lavas *(following on from previous page and continued on next page).*



From Geological Society of New Zealand Guidebook 3 - "Ancient Undersea Volcanoes" by B.W. Hayward, www.gsnz.org.nz

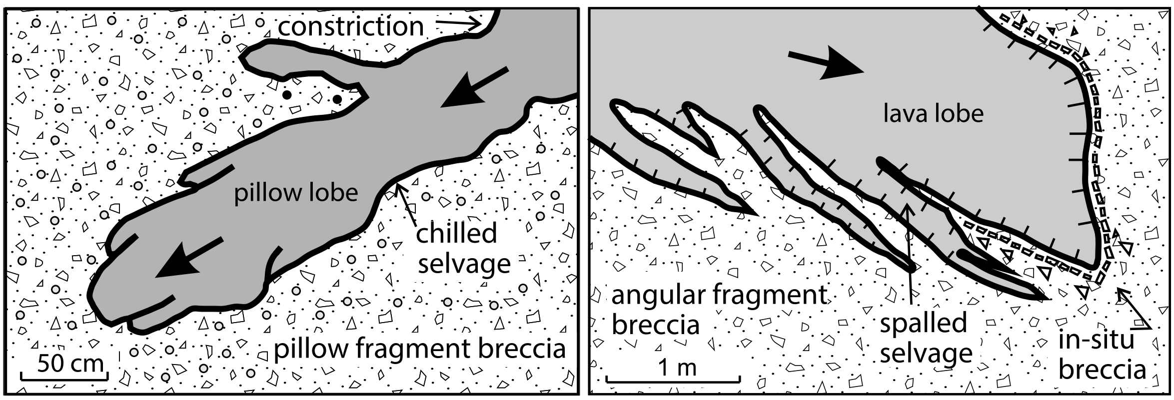
chemical analyses and radiometric dates. If the dates are correct and not reset then most of the lavas contained in the layer are younger than the fossils, and relate to a spreading basin that was active around 30 myr ago, offshore from Northland.

As the Northland Allochthon was advancing – slowly, at a rate of millimetres per year – across Northland, other effects of the new subduction zone began to appear. There is a delay between the inception of a new subduction zone and the first appearance of 'ring of fire' volcanoes, because the subducting slab first has to reach the critical depth of 100 km (Chapter 2), and it then takes some time for the new magma to stope its way to the surface. The system does not provide a chimney for magma, which has to melt rock out of the way (called 'stoping'), thereby changing its chemical composition as it goes. As the allochthon crept along, under the sea, sediment began to accumulate in hollows on its upper surface. These are called piggyback basins, for obvious reasons, and they include the first volcanic ash layers derived from the new ring of fire. This was 24 myr ago.

Box 6.4 C. Pillow lavas *(following on from 2 previous pages).*

The Maori Bay example shows a very clear and tight set of pillows. They were erupted into deep water (1-2 km depth) where the strong confining pressure prevents the pillows from breaking open.

At shallow depths the smaller confining pressure of water allows for more interaction between molten lava and water, as a result of which there is a production of quantities of volcanic glass. This material piles up around the pillow pile, such that later pillows form by intrusion of lava within the glassy debris and have highly irregular shapes. An example of this can be seen at the southern end of Bethells Beach (Te Henga) in west Auckland, close to the big cave. Visits need to be made at low tide. Afternoon light is best.



Features to look for where lava has intruded into a water-saturated pile of pillow lava debris.

The first subduction volcanoes

During the first 10 myr of volcanic activity (24–15 myr ago), more than 30 volcanoes were active in the vicinity of Northland (Fig. 6.10). Some formed an orderly line along the western side of the region, on what is now the continental shelf, from Manukau Harbour northwards to offshore Cape Reinga; some formed a parallel line along the eastern flank of Northland, from the north end of Coromandel Peninsula to North Cape; another line formed the Three Kings Ridge; some were building the Colville Ridge; and still others were dotted apparently randomly across the southern South Fiji Basin on either side of the Three Kings Ridge (look for prominent seamounts in Fig. 1.3. We do not yet have a full understanding of how the subduction zone or zones were functioning for these 10 myr, or over the previous 20 myr.

The best places to see the products of these early volcanoes are in the Waitakere Ranges west of Auckland (Waitakere Volcano), at Hokianga South Head (Waipoua Volcano), at Whangarei Heads (Taurikura Volcano) and around Whangaroa Harbour (Whangaroa Volcano). In each case the dominant rock type is a conglomerate (coarse gravel) composed of partially rounded pebbles, cobbles and boulders of dark lava. They may be cut by steeply inclined dikes of dark lava; these were the feeders for eruptive activity at higher levels. In each case, we see parts of the flanks of the original volcanoes, but not the central parts. Columnar jointing is also visible in the lava (Box 8.2).

The Waitemata Basin

The third prominent feature formed during the first 10 myr of phase four in the region was a deep sedimentary basin centred on Auckland and southern Northland (Fig. 6.10). This is called Waitemata Basin (after Waitemata Harbour), and its deposits form most of the cliffs along the north coast of Manukau Harbour and between Maraetai and Leigh, including those along the Waitemata Harbour, North Shore and Whangaparaoa Peninsula.

The sedimentary deposits that you see in these cliffs comprise an alternation of sandstones, typically 30–50 cm thick, which stick out of the cliff; and mudstones, generally 10–20 cm thick, which recede. Each sandstone was deposited in the space of a few hours from a sediment gravity flow called a turbidity current, while each mudstone accumulated very slowly over a period of

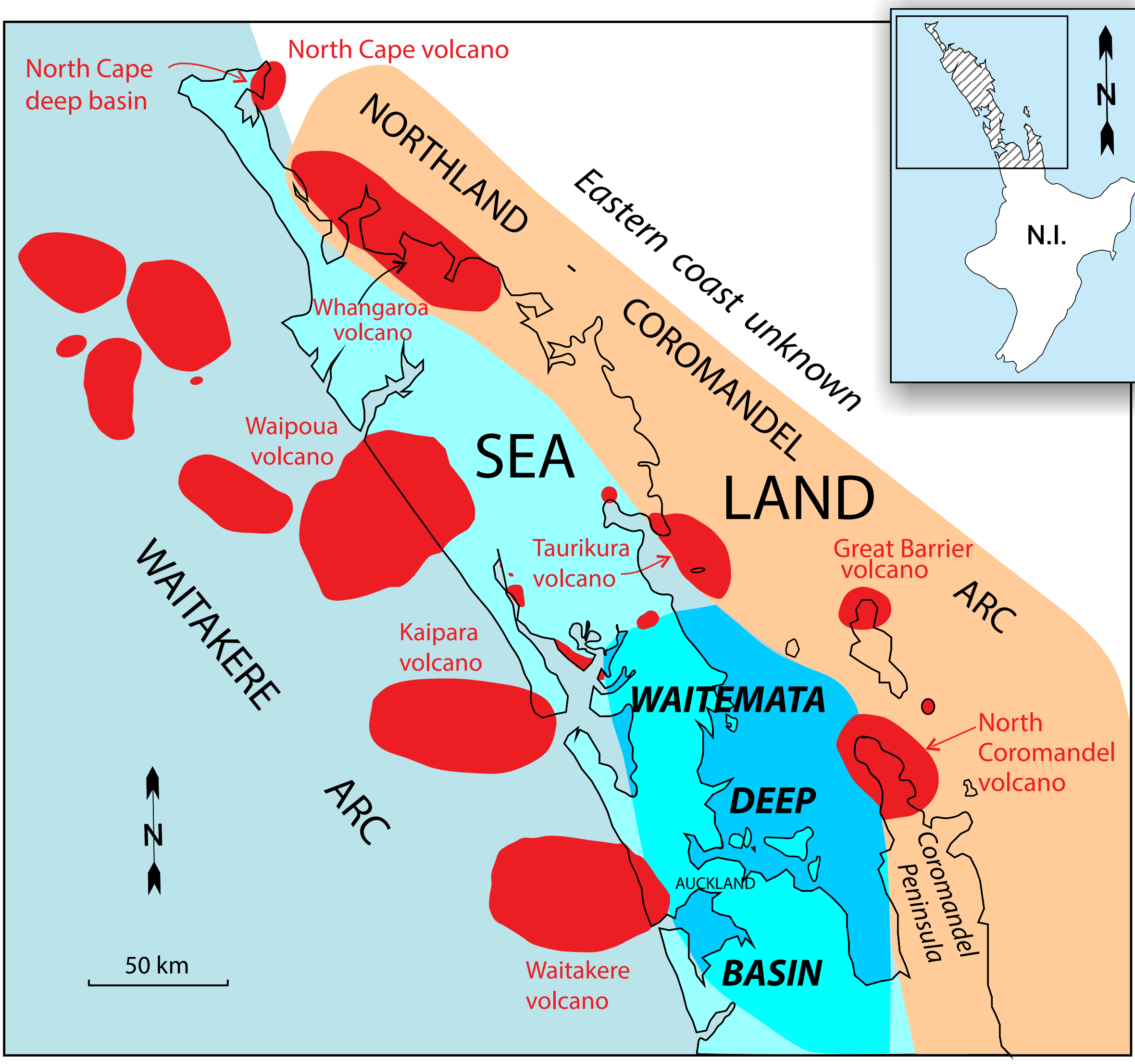


Fig. 6.10. Northern New Zealand approximate geography 20 million years ago.

hundreds or thousands of years. This kind of sedimentary sequence is known by the German-derived word flysch (pronounced 'flish' or, if you are a language purist, 'fleesh'). Boxes 6.5 A, B explain the formation of these deposits in more detail.

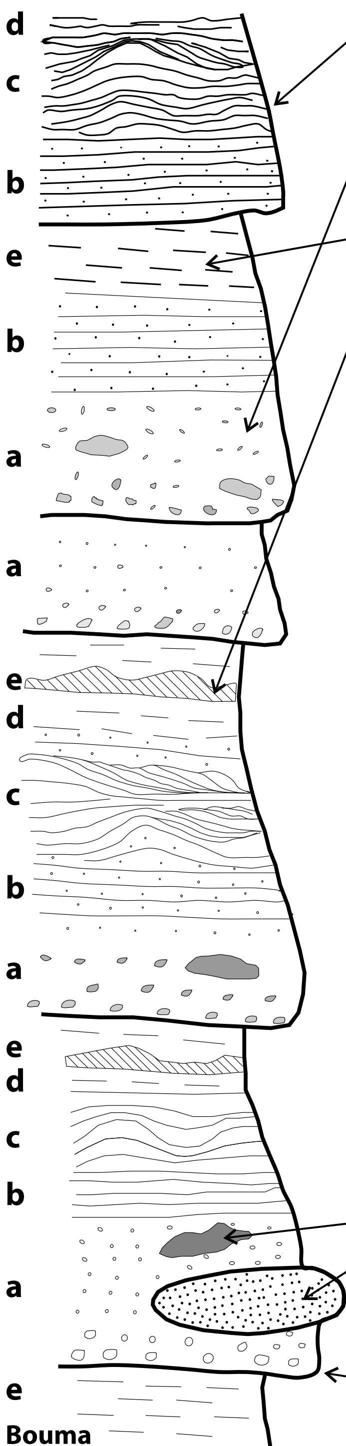
The landscape that existed before the formation of the Waitemata Basin was carried below sea-level by rapid subsidence, so quickly that coastal landforms like cliffs and seastacks were buried by sediment before erosion had time to remove them. As a result, today we can see the interesting sequence of strata from beach gravels through shallow-water fossiliferous sediments to deep-water mudstones and finally to flysch, wrapping around old seastacks made of basement greywacke. Because of the westerly tilt that was imparted to Northland 15 myr ago, we see these basal contacts mainly along the eastern coast of Auckland and Northland, where they occur mainly between basement greywackes and Waitemata strata. Good places to see them are on the northwestern coasts of Motutapu Island, Mathesons Bay, and Goat Island Bay by the marine reserve at Leigh.

While these sedimentary layers would have been more or less horizontal at the time of deposition, they are seldom horizontal when you see them now. This is because of the tectonic movements that accompanied the uplift of the basin 15 myr ago, and also because of movements of the sediment pile that accompanied the subsidence of the basin. Northland and Auckland were in the subduction earthquake zone at this time (Chapter 2). A wide variety of rock structures has resulted from these movements – Boxes 6.6 A–G illustrate some of the things to look out for.

The flysch features (Box 6.5 A, B) recur many times throughout New Zealand in different sedimentary basins of different ages. However, the Auckland rocks (known as the Waitemata Group) contain one unique kind of deposit, known as Parnell Grits (after Parnell Point in Auckland). These are strata that are thicker and darker than the surrounding sandstones and mudstones, and they contain a high proportion of pieces of lava, sometimes small pebbles, and sometimes boulders and house-sized blocks, as at Waiwera, Mahurangi Harbour, and Army Bay on Whangaparaoa Peninsula. They originated as sector collapses on the Waitakere, Kaipara and other volcanoes, and were brought into the Waitemata Basin by a different type of sediment gravity flow called a debris flow (Box 6.7).

As noted in Chapter 3, the New Zealand sector of the Pacific Ring of Fire underwent a big shake-up 15 myr ago and simplified itself to a single subduction zone/trench/volcanic arc located outside of the Northland–Auckland region. The effects in Auckland and Northland were that the arc volcanoes switched off, the Waitemata Basin was uplifted again, and everywhere north of the mouth of the Waikato River (Port Waikato) was tilted a few degrees to the southwest.

Box 6.5 A. Flysch: alternating sandstone and mudstone beds.



Very fine sandstone or siltstone (grains too small to see with the naked eye) organised into thin parallel layers or wavy (convoluted) layers with small-scale (1-5 cm) cross-bedding.

Sandstone is rough like sandpaper. Sandstone beds range from a few centimetres to several metres thick. They have abrupt lower boundaries but pass gradually upwards into grey-coloured **mudstone**, which is smooth and breaks into centimetre-size chips. Mudstone may enclose 1-5 cm beds of cross-bedded fine sandstone which formed as migrating ripples on the sea floor.

Scale - 1 to several metres

Turbidity currents

Each sandstone-mudstone pair (Bouma divisions **a** to **e** - named for the Dutch geologist who first described them) records the arrival of one turbidity current on the floor of a marine basin or lake (Fig. 6.10, for the Waitemata Basin). A turbidity current is a turbulent suspension of mud, sand, sometimes pebbles, and water (see the following diagram) that is set in motion by a sudden slope failure on the sea or lake bed. It flows down-slope until it slows, when it deposits sediment.

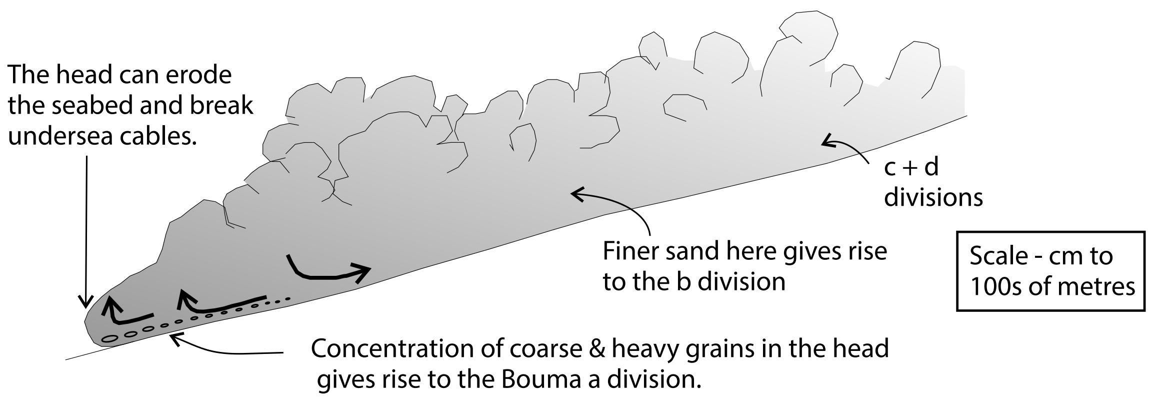
The Bouma sequence records the arrival of the different parts of the slowing turbidity current. As shown, Bouma divisions can be repeated or left out. Missing lower divisions have been left behind, already deposited. A turbidity current drops first the coarsest sediment (**a** division) from the head of the current, then the finer **b** division sand from further back, and so on. As it runs out of coarse sand, the **b** division becomes the first to be deposited, and so on until only mud is left. Mud turbidites exist, but are difficult to detect. Missing upper divisions have usually been eroded away by the next turbidity current. The head of the current sometimes erodes flutes and grooves on the underlying mud; these are filled by sand and record the direction of current flow. Each turbidity current deposits one bed of sand - a turbidite - in minutes or hours. Intervening muds settle gradually over hundreds or thousands of years. The weight of sediment accumuating on top converts sand and mud to sandstone and mudstone.

Sandstone layers may contain pieces of mudstone ("**rip-up clasts**") eroded from the seabed by the turbidity current, and **concretions -** portions of sandstone in which the sand grains have been cemented together by lime. Concretions are much harder than the surrounding sandstone. Sandstone beds are coarsest (grains up to 5 mm) at the base.

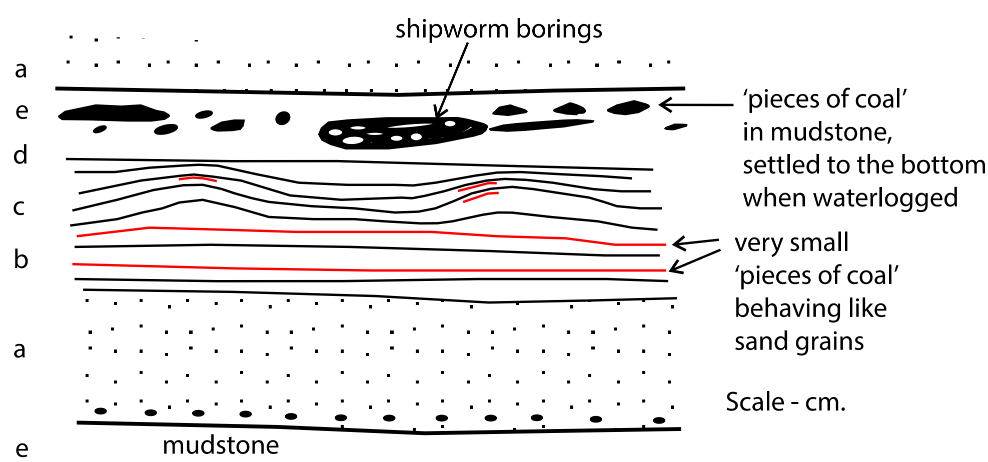
Typical weathering/erosional profile reflecting upwards fining of grain size within each bed.
As depicted, this could be either a vertical cliff exposure, or a horizontal rock platform exposure in dipping strata.

Box 6.5 B. Flysch *(continued).*

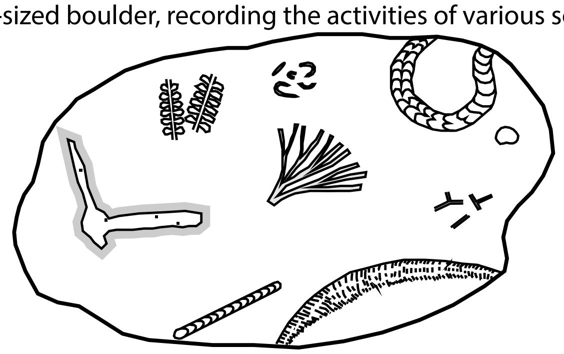
Profile of a typical turbidity current.
It may be hundreds of metres thick, and can travel hundreds of kilometres at speeds up to at least 50 kilometres per hour. A snow avalanche is a turbidity current.



PLANT FOSSILS IN FLYSCH — turbidity currents commonly carry large amounts of plant debris, preserved fossil as carbon (coal)



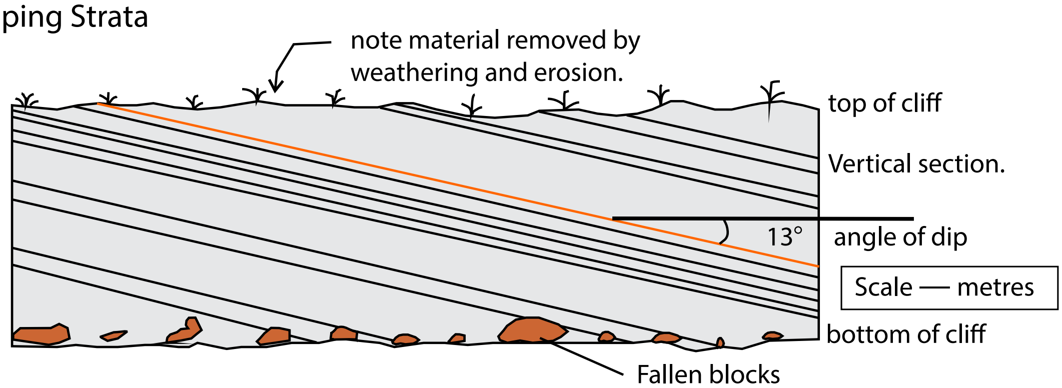
TRACE FOSSILS on a metre-sized boulder, recording the activities of various soft-bodied, burrowing organisms.



Trace fossils can be cut through in many different ways on the surface that you see. This surface is near-parallel to a bedding plane.

Box 6.6 A. Folds, faults, joints *(continued on following pages).*

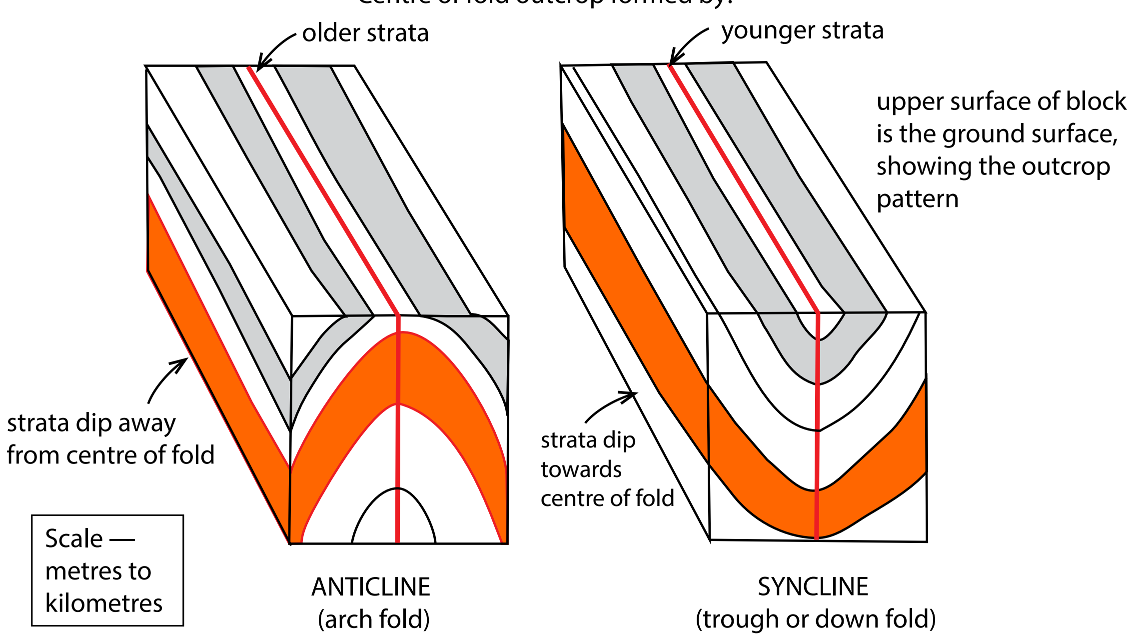
Dipping Strata



These strata have been tilted 13° from their original near-horizontal position. An anticline lies somewhere to the left and a syncline lies somewhere to the right (see folds).

Folds

Centre of fold outcrop formed by:



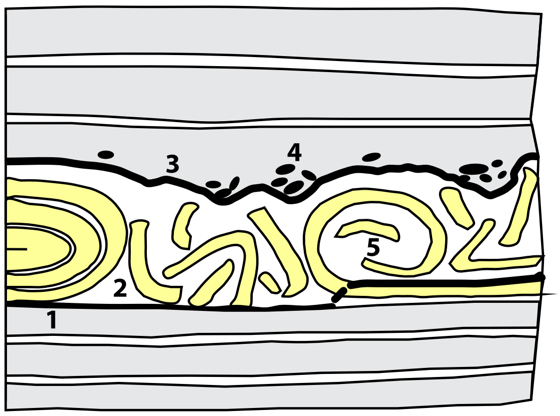
Originally near-horizontal strata have been folded by compression from left and right. These diagrams show the simplest case, where the central or axial plane (red line) is upright, and the folding axis (a line, eg. a pencil) used to outline the shape of the fold, is horizontal. Folds can also have inclined axial planes and/or tilted folding axes.

You could try drawing block diagrams of these more tricky folds. Use a rolled-up piece of paper as an aid.

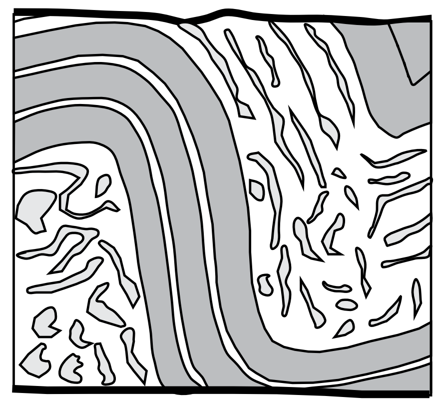
Box 6.6 B. Folds, faults, joints *(continued).*

Folds — some examples

This **slump-folded horizon** was a layer just below the sea bed that slid and was folded - probably in response to an earthquake shock.



Vertical section. Scale - metres



Vertical section.
Scale - metres to hundreds of metres

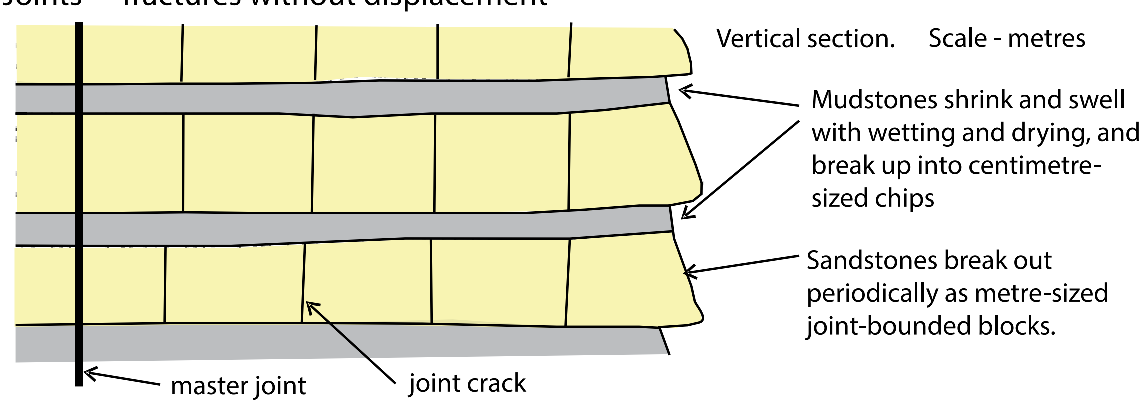
Things to look for:

1. the lower slip surface (lower heavy line)
2. the gap-filling materials - water-saturated coarse sand liquefied and flowed to fill the spaces
3. the irregular sea floor produced (the upper heavy line) and hence the irregular lower surface of the next turbidity current deposit
4. large numbers of 'rip-up clasts' in the overlying bed
5. rarely, a rolled-up bed like the one shown, which tells us that this slump slid from left to right.

This diagram shows a variation in response to the compression produced by the folding. Thicker sandstone beds were **competent**, ie. stronger, and retained their identity, whereas thinner beds with a greater proportion of mudstone were **incompetent,** ie. weaker, broke up and thickened (flowed) into the hinge area of the folds.

Joint cracks commonly form sets at right angles. They originate as drying-shrinkage structures. Their orientation may be influenced by tectonic stresses.

Joints— fractures without displacement

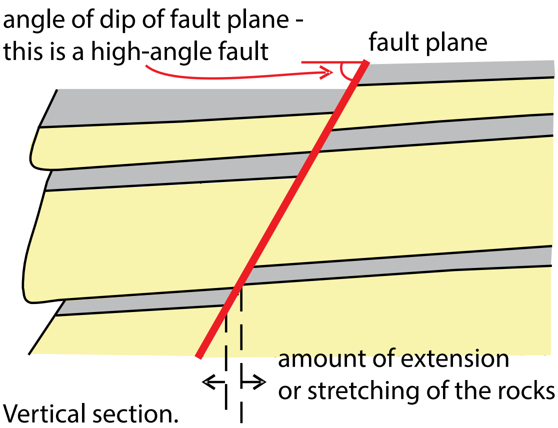


Vertical section. Scale - metres

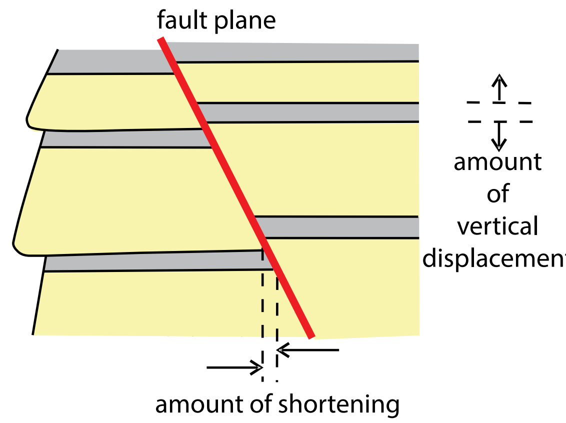
Box 6.6 C. Folds, faults, joints *(continued).*

Faults — fractures with displacements

(a) NORMAL FAULT - EXTENSIONAL



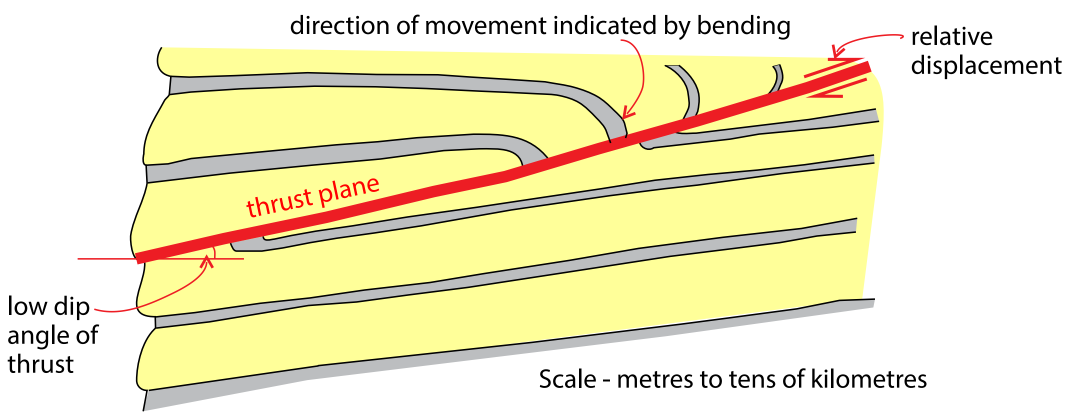
(b) REVERSE FAULT - SHORTENING



Vertical section.
Scale - metres to hundreds of metres

In both of these cases the rocks are higher to the right of the fault plane. The difference is in the direction of dip of the fault plane. If a reverse fault has a low dip angle, it is called a **thrust**.

(c) THRUST FAULT— EXTREME SHORTENING



Low-dip-angle thrust faults can have displacements of many kilometres. The diagram shows a thrust with a displacement of only a few metres, and the rocks above the fault are the same as those below. The relative displacement between the two sides of the fault is shown by the half-arrows, and by bending of layers caused by frictional drag at the fault plane. Note that this is relative displacement - we do not know whether the upper thrust plate moved to the right, or the lower thrust plate to the left. Both situations occur. Subduction of tectonic plates, for example, causes large-scale underthrusting at the plate boundary.

In the case of large displacement thrusts (the Moine Thrust of Scotland is a famous example, the first thrust to be recognised), completely different rocks are juxtaposed at the thrust plane. Mid-crustal metamorphic rocks may be thrust over upper-crustal sedimentary rocks. Displacements of tens, even hundreds, of kilometres have been established. Thrusts of this size normally form at convergent plate margins where two blocks of continental crust are in head-on collision, eg. in the European Alps. A slice of displaced rocks above a thrust is called a nappe.

Box 6.6 D. Folds, faults, joints *(continued).*

Normal faults may also have low dip-angles, and extensional displacements of many kilometres. This situation generally occurs during the crustal stretching that accompanies early stages of development of a new extensional plate boundary, beneath a continent. A present-day example is happening along the East African rift valleys, and an old example occurred when the New Zealand minicontinent began to separate from Gondwana, 100 million years ago. The rift-bounding faults flatten at depth and join in a mid-crustal plane of detachment (or décollement, the French word which is commonly used).

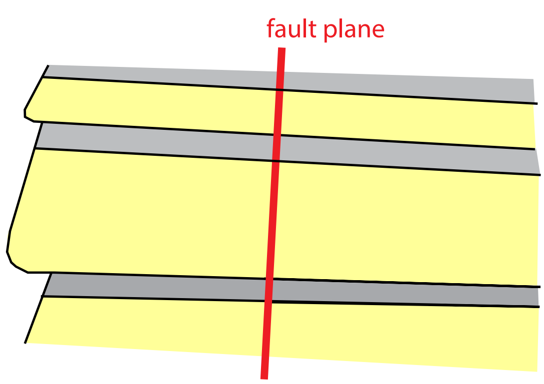
(d) TRANSCURRENT OR STRIKE-SLIP FAULT - SIDEWAYS DISPLACEMENT

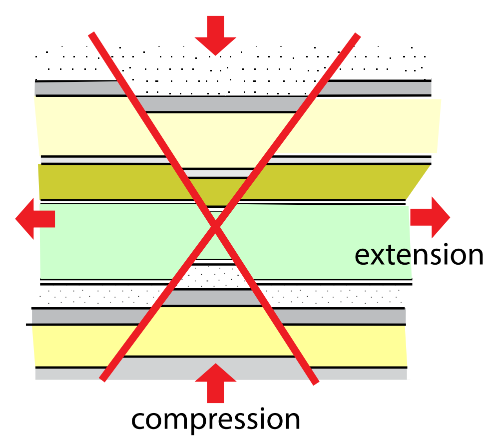
No apparent displacement in a vertical plane. To detect the offset on this type of fault requires that you can see the displacement of a steep or vertical structure in a horizontal or map view. Many faults have a combination of vertical and sideways displacement— they are **oblique-slip faults** (Box 13.3 A).

(e) A 'CONJUGATE PAIR' OF NORMAL FAULTS

Note the loss of material in the centre of the structure, and the nett extension of the rocks. This structure can occur at any orientation in the rocks, depending on the original orientation of the directions of compression and extension. Like all geological structures it is a 3-dimensional phenomenon.

Vertical sections. Scale — centimetres to kilometres





Faults are all shown here as crisp, clean fracture planes, with small displacements. However, the larger the displacement across the fault (it can be many kilometres) the more likely it is that the fault will comprise a zone of minced-up rock that may be hundreds or thousands of metres wide. That is why we seldom see major faults in natural exposures, because the minced-up rock (fault gouge) is easily weathered and eroded, and generally forms low ground.

Box 6.6 E. Folds, faults, joints *(continued).*

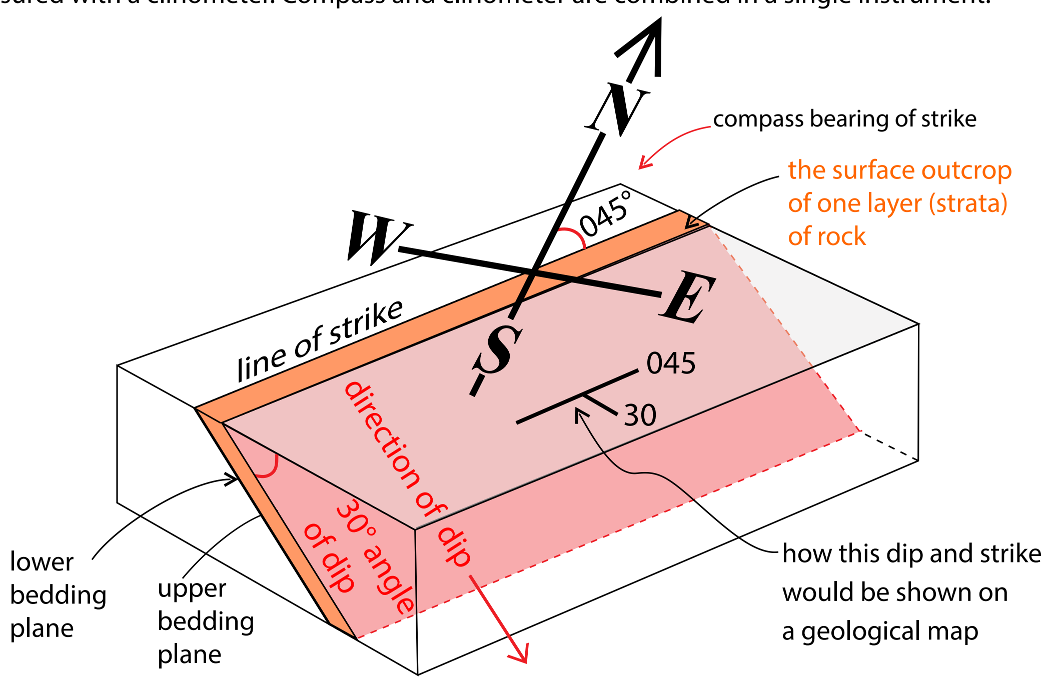
Dipping Strata

How we measure and record dipping strata — dip and strike

This is a 3-dimensional matter, requiring a 3-D diagram. The angle and direction of dip is measured, where possible, on a bedding plane. A bedding plane is the 2-D surface forming one side of a 3-D layer (stratum) of rock. Most strata are near-horizontal at the time of deposition, but are commonly tilted and folded during later earth movements.

A tilted bedding plane is defined by two lines - a horizontal line (strike line), which can be thought of as a contour line or shoreline on the sloping surface; and, at right angles to it, the line of maximum slope (direction of dip). Both directions are oriented to the compass.

Strike is measured as a compass bearing, in degrees clockwise from true north. Dip is measured in degrees from the horizontal, and dip direction by convention is down-slope, not up-slope. Dip is measured with a clinometer. Compass and clinometer are combined in a single instrument.

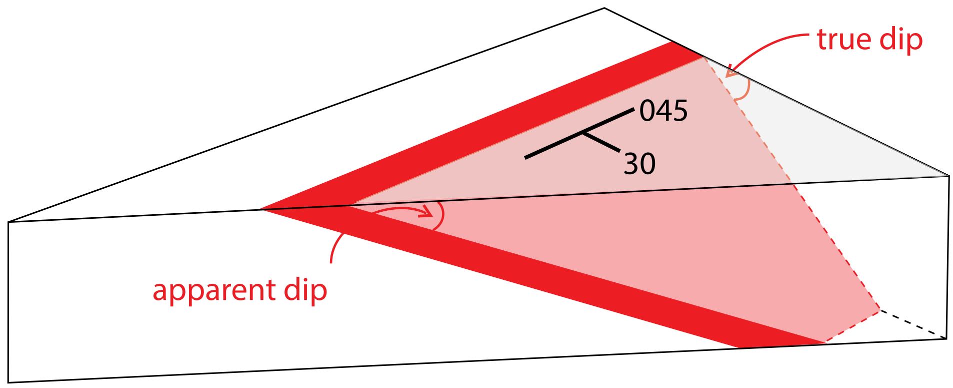


In this example, a layer of rock is dipping at 30° to the southeast. The strike is 045°, or 45° east of north. It could also be called 225° (045° plus 180°), or 045° to 225°, because strike has only an orientation, not a specific direction. Dip direction, however, is specific, at right angles to the strike, in this case to 135°, not to 315°. This measurement would be recorded in a field book as 045/30SE. It would be recorded on a map as the symbol shown on the diagram, correctly orientated, and positioned over the place where the observation was recorded.

Box 6.6 F. Folds, faults, joints *(continued).*

True dip and apparent dip

The block diagram was carefully drawn with the sides of the block parallel to dip and strike. That is, the view of the dipping bed in the side of the block shows the maximum dip, or 'true dip'. Now consider what would be seen if the block were cut diagonally from corner to corner. The bed would still dip, but the angle of dip seen would be lower than true dip, and the direction would not appear to be exactly southeast. We would be seeing an 'apparent dip'.

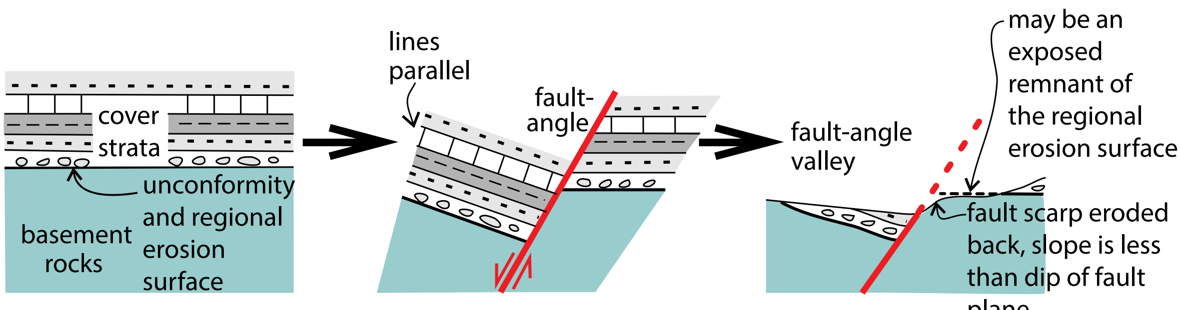


This point is important because any natural exposure of dipping strata, eg. a cliff or cutting, has an orientation which is quite random with respect to the direction of dip. The dips we see from a distance are normally apparent dips.

One final point. This measurement system can be applied to any sloping planar surface, eg. a road or a roof. In geology it is used for all planes - bedding planes, joint planes, fault planes, schistosity planes, cleavage planes, etc.

Fig. 6.11. Interbedded thin mudstone and thicker sandstone beds (flysch), that were deposited on the floor of the deep marine Waitemata Basin about 20 myrs ago, can be seen in coastal cliffs all around Auckland; seen here in cliffs between Takpauna and Narrow Neck Beach. Photographer Bruce Hayward.

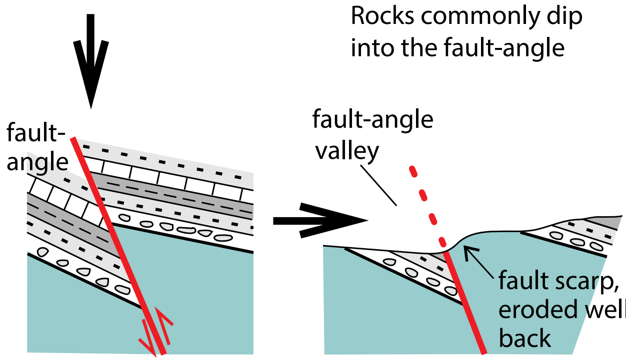
Box 6.6 G. Fault-angle valleys (fault-angle depressions).



A
cover strata deposited
on eroded basement rocks

B
extension leads to a
normal fault developing.
Rocks commonly dip
into the fault-angle

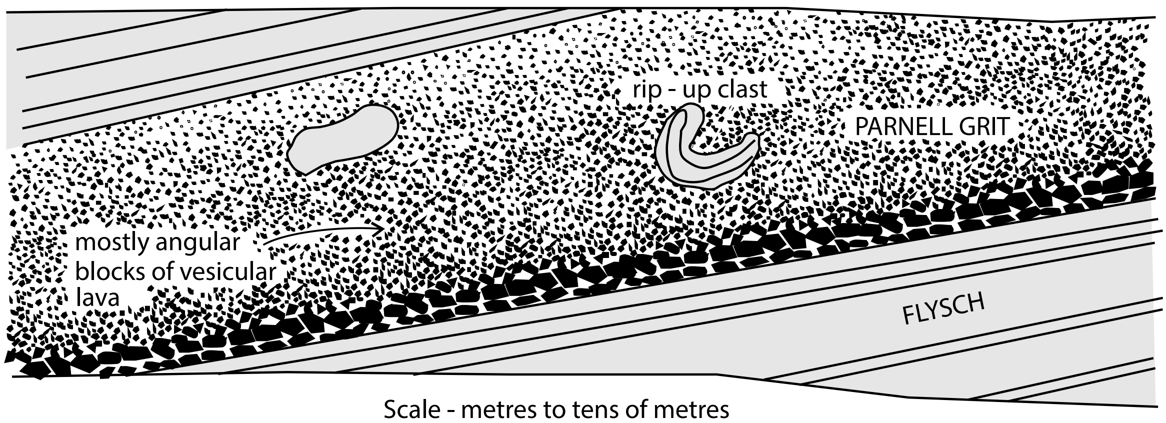
C
erosion produces an
asymmetric fault-angle valley;
cover strata may or may not
be preserved. Basement
rocks are normally more
resistant to erosion than
cover strata



B*
compression and
shortening lead to a
reverse fault. Rocks
are pushed down under
the fault, ie. they dip
into the fault angle.

C*
fault-angle valley
produced by erosion
has a similar profile to C.

Box 6.7. Parnell Grit.



Scale - metres to tens of metres

Parnell Grit beds occur within Waitemata Group flysch through most of the Waitemata Basin. Generally thicker and darker than the enclosing flysch, they are named from Parnell Point, where an 8-metre bed dips westwards behind the Parnell Baths. The dark colour reflects a high content of volcanic debris, but they also contain ripped-up blocks of Waitemata flysch, and sometimes shells of shallow-marine organisms. They have an abrupt lower contact on flysch, and an internal organisation in which particle size increases upwards to a maximum around 50 cm from the base, and then decreases upwards to the top of the bed. Maximum volcanic particle size varies from one centimetre to small house size, while rip-ups can be up to 90 metres long.

Like flysch, Parnell Grits were deposited by sediment gravity flows, but of a different kind - these were non-turbulent debris flows that originated on the Waitakere and Kaipara volcanoes (Box 6.1). Box 9.2 B describes debris flows ('lahars') on andesitic volcanoes above sea level (flowing wet concrete is a debris flow); and also the much bigger debris avalanches that occur when sectors of the volcanic cone collapse. Parnell Grit flows resulted when debris avalanches from the two volcanoes entered the sea, picked up near-shore shells, took up sand and water, and became large submarine debris flows. They were able to carry blocks of lava up to 20 m across at least 70 km into the basin and to scour the sea bed to acquire the rip-ups. However, there is little sign of scour where they deposited the Parnell Grit beds on the flat basin floor.

The difference between a Parnell Grit bed and a lahar deposit above sea level (Box 9.2 B) is that Parnell Grits lack large floating boulders near the top - the uptake of sea water reduced the overall strength of the flow. In fact, many Parnell Grit flows left all of their heavy debris behind, and carried only fine material to the point of deposition. The best places to see coarse-grained Parnell Grits are south of Waiwera beach and east of Army Bay, Whangaparaoa Peninsula.

Auckland – city of volcanoes

Despite the change that occurred 15 myr ago, the region wasn't yet done with volcanic activity. One of the interesting things about subduction zones is that the downgoing slab of lithosphere spawns activity in the overriding plate, specifically in the wedge-shaped piece of lithosphere that is in direct contact with the downgoing slab. One effect of this is to set up convective overturn in the wedge, which in turn creates a back-arc spreading basin – a small version of an oceanic spreading basin (Box 2.2). The back-arc spreading system can split the volcanic arc into an inactive remnant arc and an active arc, as we have happening at present north of the Bay of Plenty (Fig. 1.3).

Another effect seen in the overriding plate is the generation of localised production of basaltic magma. This has happened at some 10 places underneath South Auckland, Coromandel, Auckland and Northland – seemingly at random, and in some places more than once – over the past 11 myr (Box 6.8). The best known of these basaltic volcanic fields is the Auckland one, which is technically still active but dormant (Box 6.8). All the fields share common features, such that within a clearly defined area lots of separate volcanoes (>50 in Auckland) erupt at different times over the life of the field (which may be several million years) and in different places inside the field boundary. This magma-generation process produces magma only in dribs and drabs. Normally, only one volcano is active at a time, there are long periods between the formation of volcanoes, and each volcano has only one eruptive cycle. That said, recent research is suggesting that in some cases several volcanoes may have been active all at once.

The eruptive cycle of these volcanoes tends to follow the same pattern. The first slug of molten magma approaching the surface encounters water. This causes steam explosions – many of them – which blast a crater and build a tuff ring round it. Some volcanoes stop there, but if the supply of magma continues, the water that is present is dried up or excluded and the eruption moves on to fire-fountaining of gas-rich lava, which builds the familiar scoria cones with their small craters. If there is still more magma, lava flows emerge from around the bases of the cones. Boxes 6.9 A, B describe the eruption styles of these basaltic volcanoes in more detail, while Fig. 6.14, a cross section through Lake Pupuke and Rangitoto Island volcano, shows an example of the succession.

The other way in which Auckland and Northland have been affected by recent volcanic activity is by the output from the highly productive arc volcanoes of the central North Island, both directly and indirectly (Chapter 9). The products of some very large individual eruptions have reached the area directly, both as airborne ash and as ground-hugging ignimbrite flows. A particularly large and far-travelled ignimbrite flow reached the Manukau lowlands and Auckland 1 myr ago.

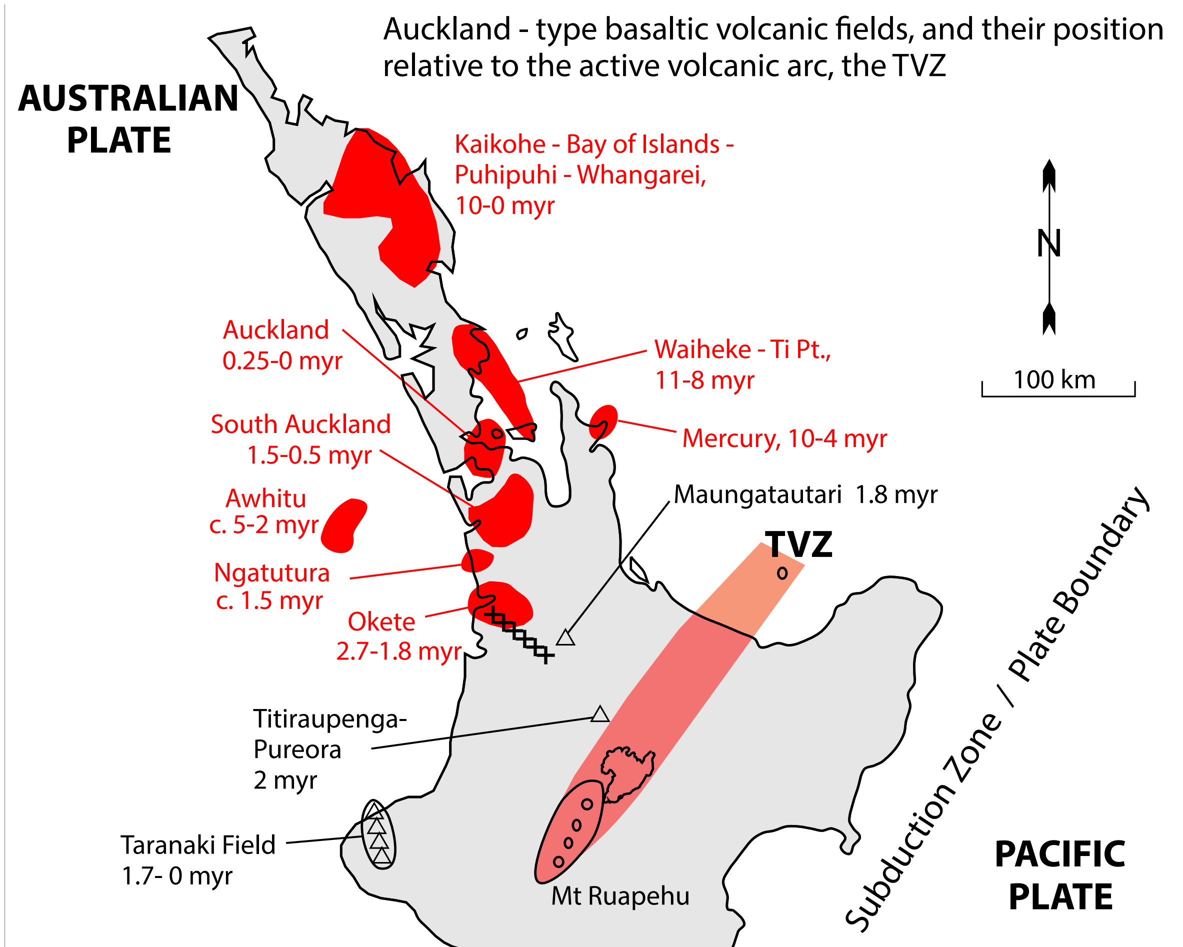
The biggest effect from the central North Island, however, has been the transportation of huge amounts of volcanic sand by the Waikato River. As described in Box 9.7, the river has delivered sand both to the Firth of Thames/Hauraki Gulf and to the west coast. At the west coast, northwards-directed longshore drift of sand over a period of 2 myr, in combination with strong westerly winds, has built the huge sand barriers that enclose the Manukau and Kaipara harbours (Box 6.10). Further north, the same sand supply has built the north head of Hokianga Harbour and the Aupouri Peninsula, a tombolo that links the former islands of Mt Camel and Cape Reinga-North Cape to Northland.

Box 6.8. Basaltic volcanoes.

Basalt volcanoes are generally small in New Zealand – Rangitoto Island, Auckland, is probably New Zealand's largest young one. However:

1) basalt can build large shield volcanoes – the island of Hawaii comprises several overlapping shield volcanoes and, measured from the ocean floor, is the world's largest free-standing mountain;
2) flood-basalt outpourings (eg. the Deccan Traps of India, the Columbia River Basalts of northwest USA, the Ontong Java Plateau in the western Pacific Ocean) are the world's most voluminous volcanic structures;
3) basalt generated at mid-ocean spreading ridges forms those ridges, and ultimately all of the ocean floor, making it the most voluminous single substance in the earth's near-surface. All other kinds of lava are derived in some way from basalt.

Auckland - type basaltic volcanic fields, and their position relative to the active volcanic arc, the TVZ



Karioi-Pirongia-Te Kawa line of andesitic volcanoes, 2.7-1.6 myr

△ subduction-related andesitic volcanoes less than 2 myr old, lying west of the TVZ

TVZ Taupo Volcanic Zone, which produces occasional small basaltic eruptions in addition to the dominant rhyolitic and andesitic eruptions

myr age in millions of years

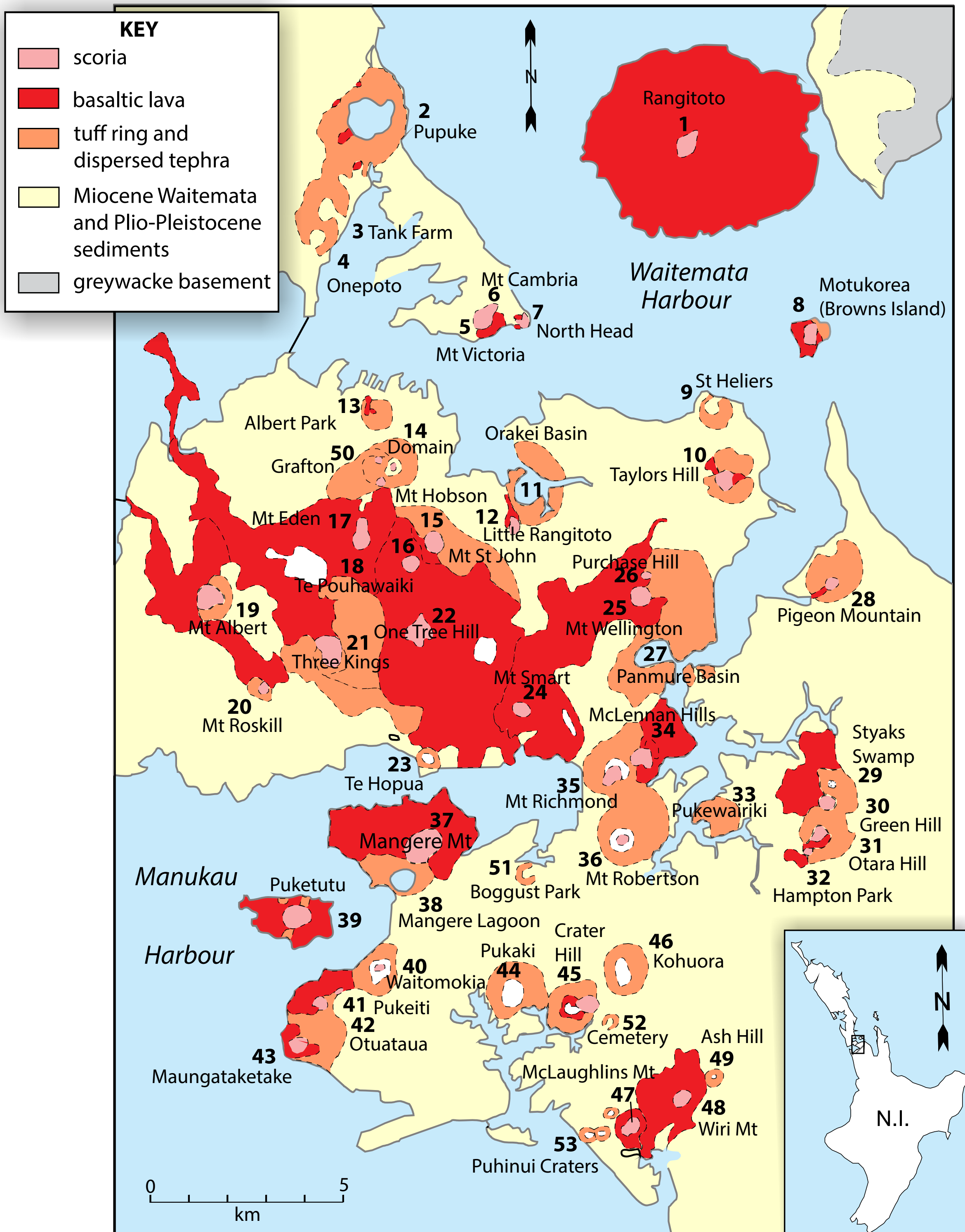


Fig. 6.12. The Auckland volcanoes.
(numbers refer to next page)

Table 6.1. Auckland's 53 volcanoes.

1. Rangitoto	28. Ohuiarangi / Pigeon Mt
2. Pupuke Moana / Lake Pupuke	29. Styaks Swamp
3. Te Kopua o Matakamokamo / Tank Farm	30. Matanginui / Green Mt
4. Te Kopua o Matakerepo / Onepoto	31. Puke o Tara / Otara Hill
5. Takarunga / Mt Victoria	32. Hampton Park
6. Takararo / Mt Cambria	33. Pukewairiki / Highbrook Park
7. Maungauika / North Head	34. Te Apunga o Tainui / McLennan Hills
8. Motukorea / Browns Island	35. Otahuhu / Mt Richmond
9. Whakamuhu / Glover Park (St Heliers)	36. Mt Robertson / Sturges Park
10. Taurere / Taylors Hill	37. Te Pane-o-Mataoho / Mangere Mt
11. Orakei Basin	38. Mangere Lagoon
12. Maungarahiri / Little Rangitoto	39. Te Motu a Hiaroa / Puketutu
13. Albert Park	40. Waitomokia / Mt Gabriel
14. Pukekawa / Domain	41. Puketapapakanga a Hape / Pukeiti
15. Ohinerau / Mt Hobson	42. Otuataua
16. Te Kopuke / Mt St John	43. Maungataketake / Elletts Mtn
17. Maungawhau / Mt Eden	44. Te Pukaki Tapu o Poutukeka / Pukaki Lagoon
18. Te Pou Hawaiki	45. Crater Hill
19. Te Ahi-ka-a-Rakataura / Mt Albert	46. Kohuora
20. Puketapapa / Mt Roskill	47. Matukutureia / McLaughlins Mt
21. Te Tatua a Riukiuta / Three Kings	48. Matukutururu / Wiri Mt
22. Maungakiekie / One Tree Hill	49. Ash Hill
23. Te Hopua a Rangi / Gloucester Park	50. Grafton
24. Rarotonga / Mt Smart	51. Boggust Park
25. Maungarei / Mt Wellington	52. Cemetery
26. Te Tauoma / Purchas Hill	53. Puhinui Craters
27. Te Kopua Kai a Hiku / Panmure Basin	

Fig. 6.13. Auckland city is built over the Auckland Volcanic Field which has erupted from more than 50 different small volcanoes over the last 250,000 yrs. Volcanic landforms are mainly scoria cones like Mangere Mt (centre) and explosion craters like Mangere Lagoon (right). The youngest and by far the largest eruption in Auckland produced Rangitoto Island (left distance). Photographer Bruce Hayward.

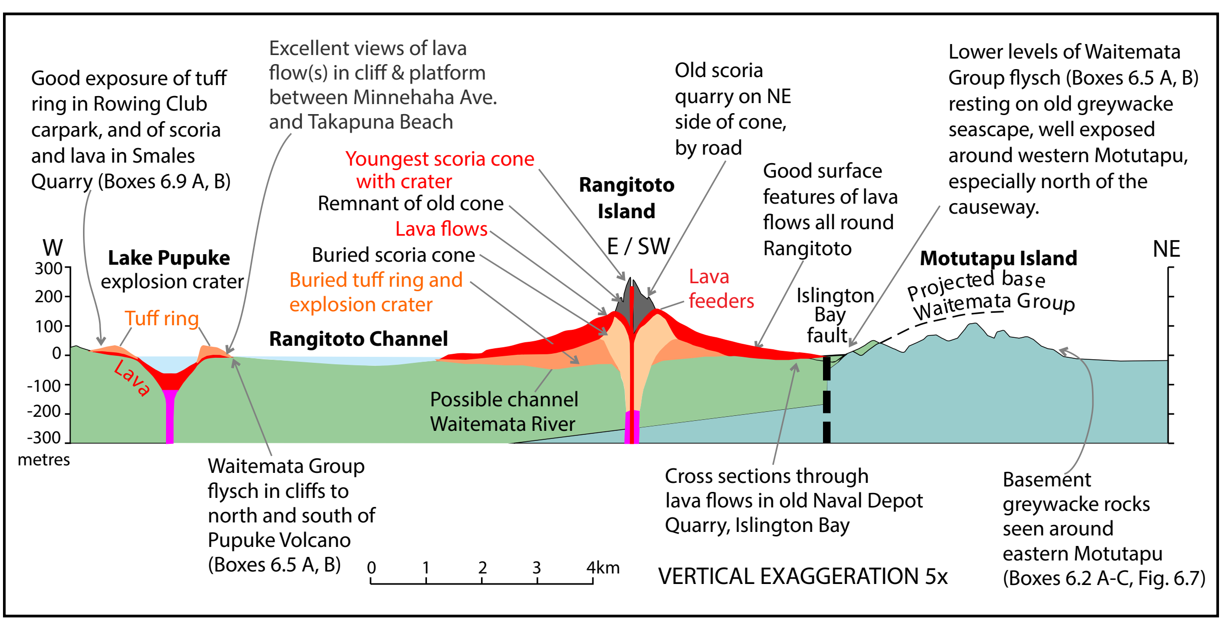


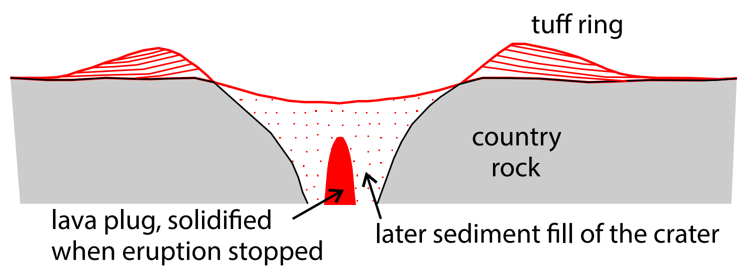
Fig. 6.14. Cross section through Rangitoto Island Volcano.

Box 6.9 A. Basalt volcanoes.

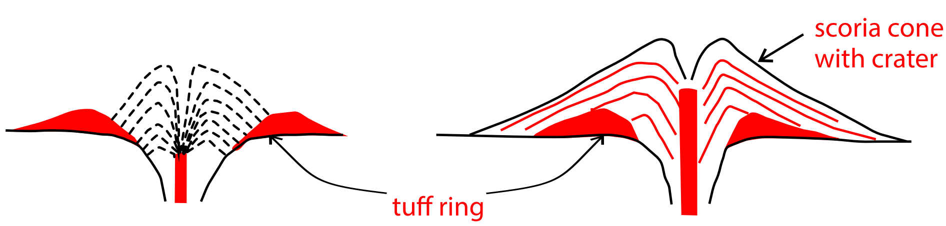
Basalt - lava contains about 50% silica.
Properties - melts at about 1000° C
- is free-running
- forms a heavy, dark-coloured rock, typically containing rounded gas cavities.

Eruption styles

1. **Explosive** - molten lava at >1000°C encounters water, commonly ground water below the surface. Water flashes to steam, causing repeated steam explosions. Product - a circular crater 1-2 km in diameter, surrounded by a low rim of layered ash, the 'tuff ring'. Ash consists of finely divided volcanic glass (rapidly chilled lava), with larger volcanic bombs, and blocks of the country rock that surrounds the vent.

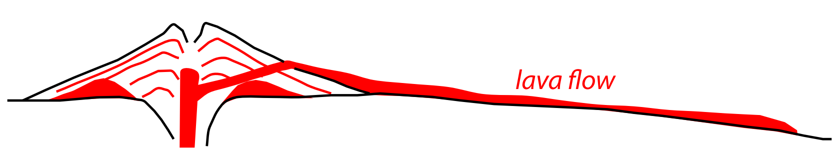


2. Some eruptions never get past this stage.
Commonly, however, supply of lava continues, water is excluded from the vent, and the eruption passes on to **fire-fountaining**. Large volumes of gas, released by the lava as its confining pressure drops to atmospheric, drive a spray or fountain of red-hot and molten lava. Lava fountains will fill and cover the tuff ring if lava supply continues. If lava supply dwindles, the tuff ring may not be buried.



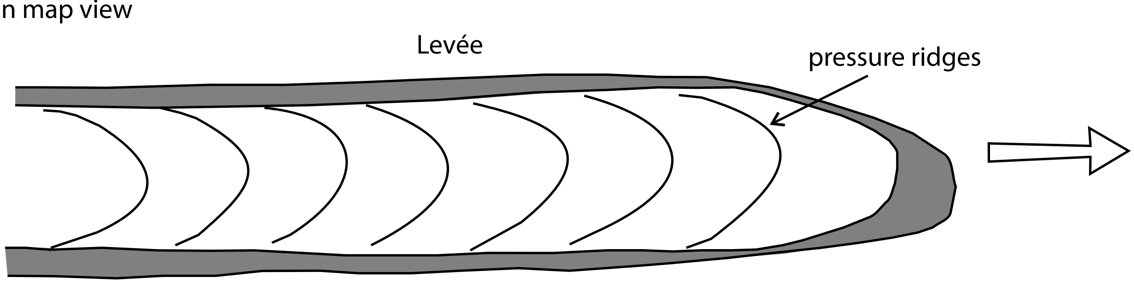
Product - a steep-sided cone of scoria.
Scoria - a deposit composed of small, centimetre sized, frothy pieces of basalt, black and glassy when fresh but commonly oxidised to a red colour (basalt contains several percent iron). The deposit typically shows layering parallel to the outer surface of the cone. The scoria pieces may be welded together to some extent.

3. Lava erupted as a fire-fountain may recombine, after falling, into a molten lava. Or less gassy lava may be extruded quietly from fissures and vents around the scoria cone. Either way, **lava flows** may extend for several kilometres from the vent. Where unconfined by existing hills and valleys, lava flows build a circular cone or shield volcano with gentle slope - e.g. Rangitoto Island, Auckland.



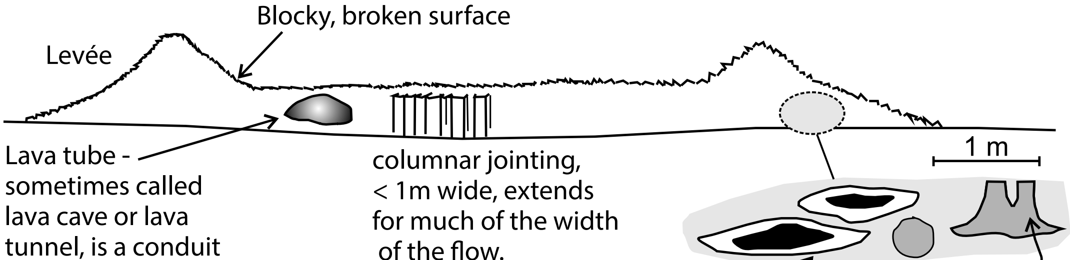
Box 6.9 B. Features of basaltic lava flows.

1. In map view



Scale :
width — 10's of metres, length — 100's to 1000's of metres.

2. In cross section



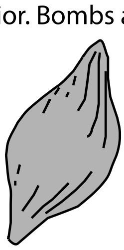
Lava tube - sometimes called lava cave or lava tunnel, is a conduit through which lava is transferred through the stationary, cooling, early part of the flow, to the actively extending front of the flow.

columnar jointing, < 1m wide, extends for much of the width of the flow.
(Box 8.2)

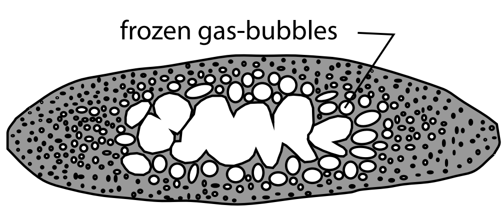
Surface sometimes smooth and folded like silk fabric -'pahoehoe' texture

VOLCANIC LAVA BOMBS

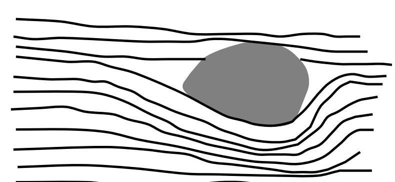
The name refers to any block of lava flung through the air. If molten, the lava may spin and form a spindle bomb, or may just splat and form a cow-pat bomb. Gas often continues to be released from the molten interior, but cannot escape through the cooled and solidified exterior, leading to a hollow interior. Bombs are commonly up to 1 metre long.



Spindle bomb



Section through cowpat bomb



Bomb-impact causes a sag structure in layered volcanic ash. Bomb came from the left.

Longshore drift involves a great deal of physical wear on the sand grains, which are nearly all mineral grains derived from volcanic ash and lava. In the process, all mineral grains except the super-tough quartz are gradually eliminated. This is why the sands of the Far North are quartz-rich, and why the white dune sands of Parengarenga Harbour contain more than 99% quartz and were the raw material for New Zealand's glass-making industry for many years.

In contrast, west coast beaches from Muriwai south to Mt Taranaki are dominated by black sand, or ironsand. The purplish-black mineral is titanomagnetite, a magnetic iron ore containing some titanium. It is also derived from lavas, of both central North Island and Mt Taranaki, where it is an 'accessory' mineral, comprising less than 1% of the volume of the lava. The presence of ironsand all along the west coast as far north as Muriwai tells us that some of the sand here has travelled up from Taranaki, although the bulk of it – including much of the ironsand – is from central North Island. Being much heavier than the other minerals on the beach, ironsand is concentrated at the top of the beach (wave uprush is stronger than backwash). Windblown dune accumulations, skimmed off the top of the beach, form the ore used by the ironsand industry to make steel; these accumulations are found at Taharoa, Kawhia (exported via ironsand tanker), and at Waikato North Head, which feeds the steel mill at Glenbrook.

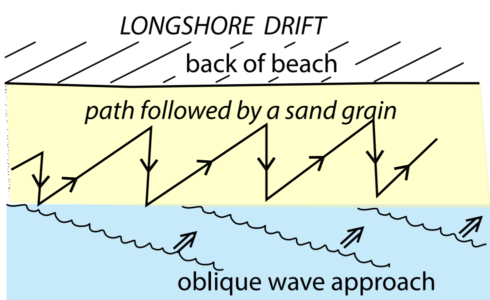
The Ria Coastline

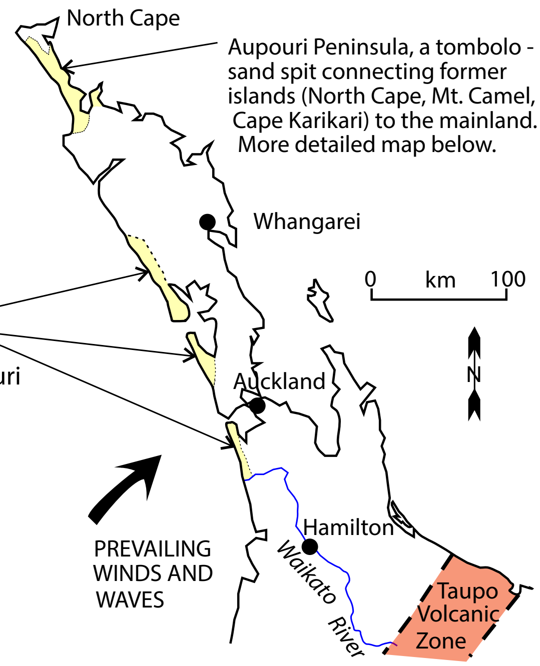
As noted above, Auckland and Northland have been above sea-level for about 15 myr. The main geological process for all of that time has been erosion. For the past 2 myr, both erosion and coastal sand accumulation have been greatly influenced by up and down movements of sea-level resulting from global glaciations. There has been a glaciation roughly every 100,000 years, with the last one peaking 20,000 years ago. Water locked up in ice sheets lowers sea-level by up to 130 m. This greatly lowered base level of erosion rejuvenates rivers, which cut down their valleys to the new level. Thus, although Auckland and Northland were well outside the region affected by ice sheets, they were nevertheless strongly affected by glaciations. In particular, while the Northland peninsula is quite narrow at the present time and does not support any big rivers, it was much wider when the sea-level was 130 m lower than it is today. As a result, existing rivers were much longer then, and correspondingly more powerful, so that valleys were excavated deeply, especially across the coastal plain of that time (Box 6.12).

For reasons connected with the climatic controls on glaciation, each glacial peak is followed by rapid warming to a climatic maximum, which is where we are at the present time. A correspondingly rapid melting of ice drives a rapid rise of sea-level, at rates averaging 1.3 m/100 years. Deeply incised valleys are then drowned, giving us the strongly indented ria coastline that we presently enjoy. This is particularly the case in Northland and Auckland, where sediment supply to the coast by rivers is small, and the rias have not been filled to the extent that they have in many other parts of New Zealand. There is more on the sea-level story in Chapter 10. All the harbours and estuaries here formed in this way. The great contrast in 'roughness' between the eastern and western coastlines exists because the smaller erosional indentations on the west coast have been 'smoothed' by the huge movement of sand.

Box 6.10. Longshore drift, sandspits, tombolos and sand barriers.

Longshore drift of sand results when waves approach the coast at an oblique angle, as shown. Wave uprush moves sand grains obliquely up the beach, while backwash, starting from zero momentum, always runs down the steepest slope. Individual sand grains therefore move along the beach in a zig-zag fashion. Longshore drift moves huge quantities of sand around the New Zealand coast, because the prevailing winds and waves are oblique to most of the coastline.



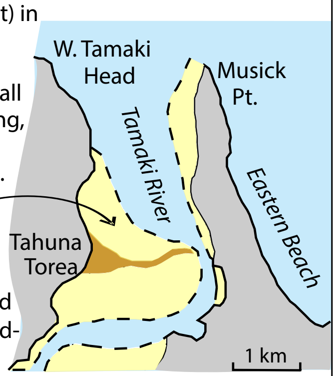


These three major sand barriers, along with the Aupouri tombolo, are between 1 and 2 million years old, and are built mostly from sand brought from the Taupo Volcanic Zone by the Waikato River, and moved up the west coast by longshore drift and wind.

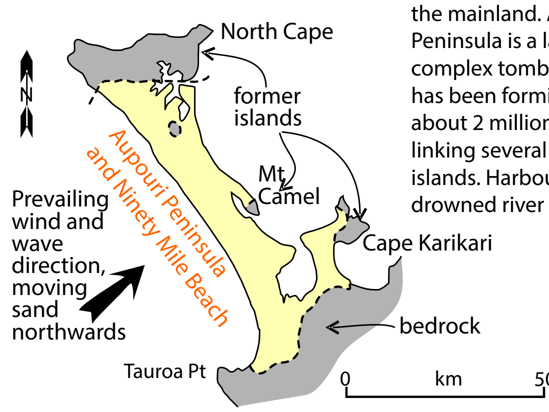
There are many small sandspits built across estuaries in the past 7500 years, like the example below

TAHUNA TOREA (Glendowie sandspit) in the Tamaki Estuary, eastern Auckland. An example of a small sand spit, 1.5 km long, built since sea-level rose 7500 years ago.

This intertidal platform is partly rock-cut (by cliff retreat, Box 6.11) and partly sand-and-mud-built, all in the past 7500 years.



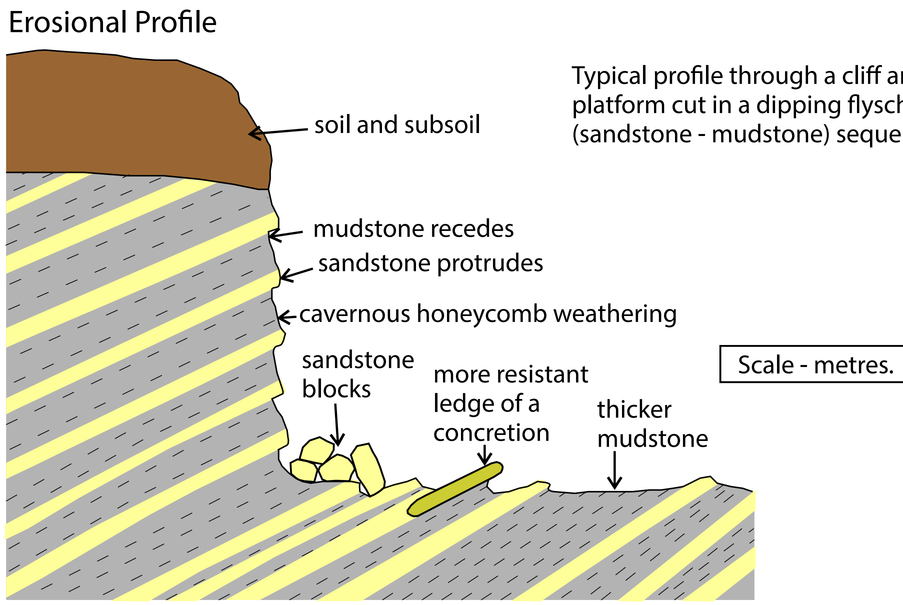
NORTH CAPE TOMBOLO (northernmost North Island)



A tombolo is a sandspit connecting an island to the mainland. Aupouri Peninsula is a large and complex tombolo, that has been forming for about 2 million years, linking several former islands. Harbours are drowned river valleys.

Sandspits are built when longshore drift moves sand into a recessed piece of sheltered water, e.g. across the mouth of a drowned river valley. The world is richly supplied with drowned valleys as a consequence of the glacial low sea level of 20,000 years ago. Box 6.12 explains why.

Box 6.11. Intertidal rock platforms.



The intertidal platform exists because the cliff retreats under erosion faster than the platform rocks do. Rocks below platform level are kept more-or-less permanently wet, both behind the cliff by the water table, and in front of the cliff by the twice-daily high tide. Chemical weathering processes, as described in Box 6.1 B, are hindered by the lack of oxygen, and physical erosion is minimal. However, above platform level, the rocks are subjected to high levels of physical weathering, by wetting and drying, salt crystallization, tree roots, and oxidation (Box 6.1 A). Waves remove the debris from the foot of the cliff. There is always a small-scale relief caused by variations in hardness of the rocks, and the presence of defects like joints and faults (Boxes 6.6 B-D). In the 7500 years since sea-level stabilised at its present level, some cliffs have retreated a kilometre.

Fisherman's Rock — high-tide level platforms

Some rock types, for example thick, coarse, homogeneous sandstones, lend themselves to the formation of platforms at around high tide level. The controlling factors seem to be the level of the water table behind the cliff, and salt pan weathering around the margin of sea-water puddles. Evaporation of sea-water in the puddles at low tide, combined with the secretions of tiny marine snails that live in the puddles, gives rise to a corrosive brine which causes precipitation of salt crystals around the margin of the puddle, and possibly dissolves some parts of the rock. The puddles enlarge, forming horizontal salt-pan steps at various levels. The most famous example of a high-tide platform is probably Fisherman's Rock at Muriwai, where the outer part is the highest, because it is kept the wettest by constant spray. The surface is swept clean of debris by waves. Joints in the rock (Box 6.6 B) control the shape of the platform and the location of blowholes.

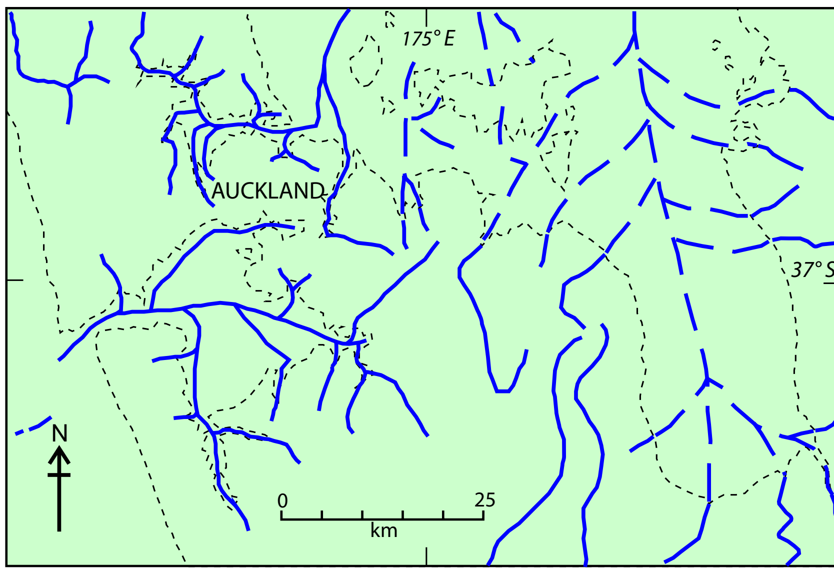
Box 6.12. Drowned coastlines.

Most of the world at present has a 'drowned' coastline. The cause is the glaciations. At the height of a glaciation (the last two were 20,000 and 140,000 years ago) the large quantity of water locked up in continental ice caps causes sea-level to fall up to 130 metres. Rivers excavate their valleys to that low base level. The valleys are then 'drowned' by the post-glacial rise in sea-level as ice sheets melt, giving us the many harbours and embayments.

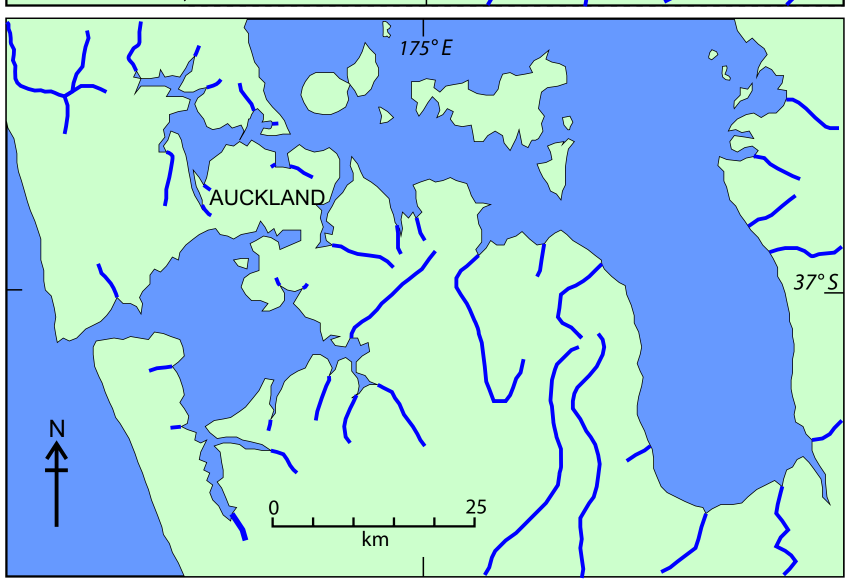
The maps illustrate the effects in the Auckland region. Sea-level returned to the present level between 18,000 and 7500 years ago - a rise of more than one metre per century. This figure is important in the context of the current debate about global warming, and the likely rise of sea-level.

A - low sea-level.

A greatly enlarged land area.



B - high sea-level



See Boxes 10.2 A, B for some more consequences of changes of sea-level driven by glaciations.

Chapter 7

South Auckland, the Waikato and King Country

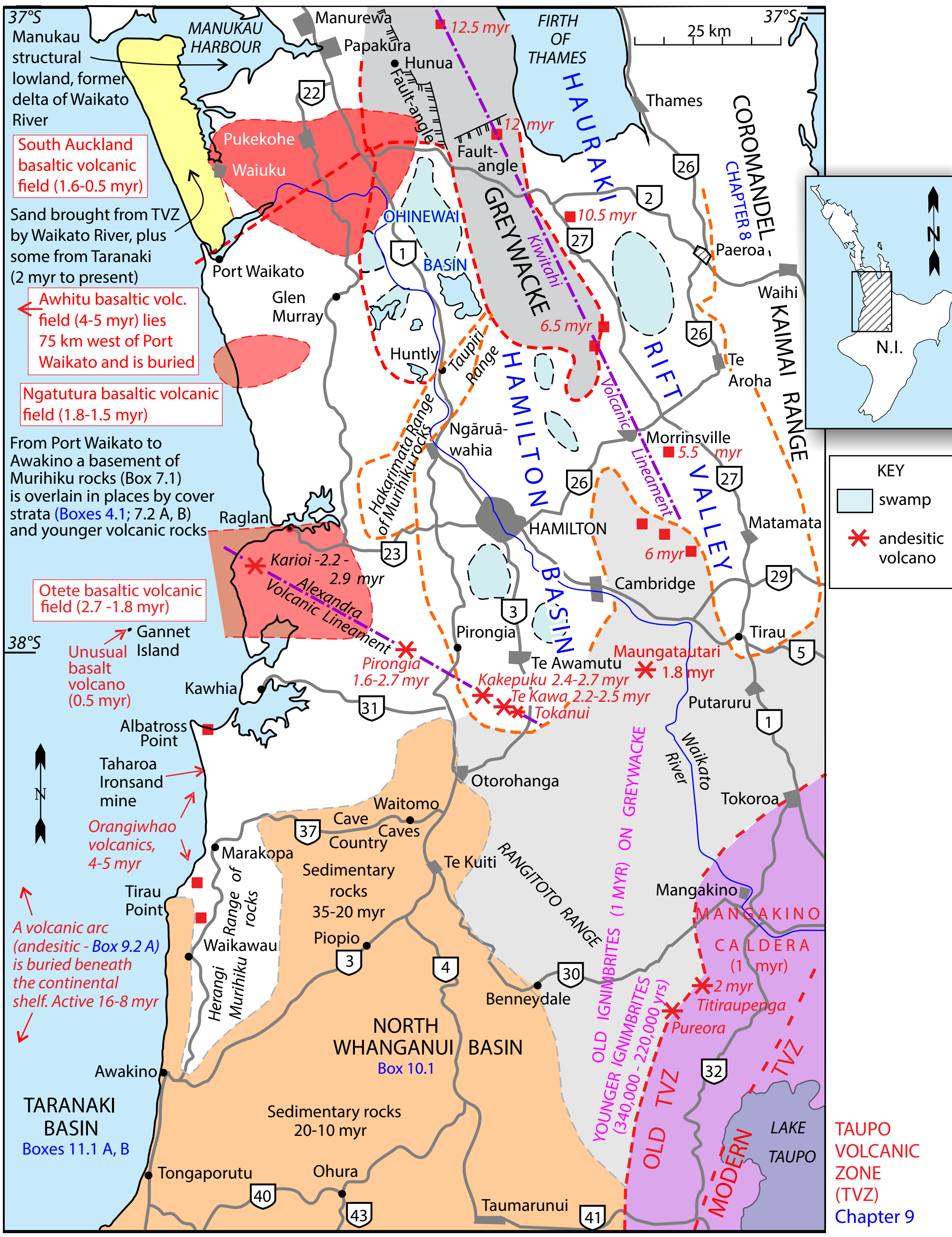


Fig. 7.1. Simplified geology of South Auckland, the Waikato and King Country.

In this chapter we move south of the Waikato River mouth. Here, the geology is mostly a continuation of that found in Auckland and Northland, but without the westerly tilt, without the Northland Allochthon, and with a more prominent representation of the phase three limestones (Chapter 6 - section on passive subsidence of New Zealand). As in the north, there are volcanoes here, including basaltic fields like that in Auckland (Box 6.8), a few small arc volcanoes and some new types as typified by Mt Pirongia (see below). The region is bordered to the north by the Waikato Fault, which follows the lower part of the Waikato River and is the 'scissors' contact between land that is tilted to the west and land that is not. The southern part of the region ends at Mt Taranaki.

Geological History

Starting with the basement rocks (Chapter 5), we have the same two terranes in this region that underlie Northland, corresponding to phase one of New Zealand's formation (Box 5.3). Without the westerly tilt, however, both of these terranes are visible at the surface. Thus greywacke rocks of the Waipapa Terrane (Boxes 6.2 A–C, Fig. 6.7) form the eastern ranges of the region, from Hunua Range in the north to Rangitoto Range in the south; and Murihiku Terrane rocks (Box 7.1) form the western belt, from Port Waikato southwards to the Herangi Range near Awakino. They adjoin along a major tectonic boundary or terrane suture, the Waipa Fault, which like many major faults is not exposed to view anywhere.

The region's cover strata comprise representatives of both phases three and four of New Zealand's development. During phase three, extensive coastal swamps formed as the low-lying landmass slowly subsided, these resulting in the coal measures found today (see below). As in Auckland and Northland, they were overlain by shell limestones and associated glauconite-bearing sandstones deposited on a wide continental shelf. Where the land was higher, swamps did not develop and so here limestone rests directly on basement, as it does between Port Waikato and Awakino. Limestone is prominent in the landscape, forming whitish bluffs, all down the west coast as far as Awakino. It is particularly strongly developed around Waitomo, where it hosts the famous Waitomo Caves (Box 7.2 A, B).

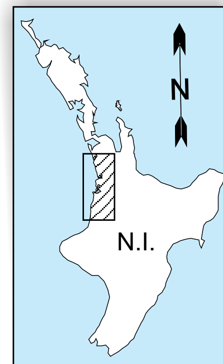
The coal measures that formed from coastal swamp deposits are confined to areas that were old valleys, chiefly around Huntly and Maramarua. Coal here has been mined by both underground and open-cast methods, and until recently has been of considerable economic importance (Box 7.3 A, B).

Sedimentary basins

Phase four of the country's development is represented here by the laying down of sedimentary deposits, recording the southern extension of Auckland's Waitemata Basin (Chapter 6). These strata can be seen in road (e.g. SH1) and rail cuttings between Pokeno and Te Kauwhata, and are very similar to the Auckland strata.

Further south, we leave the Waitemata Basin and enter another major sedimentary basin, the North Whanganui Basin. The Whanganui Basin as a whole is gradually moving south and is still actively forming beneath the South Whanganui Bight (Chapter 10), so we are looking here at a part of it that formed early on, around 20 myr ago. The deposits in this basin, which overlaps the

Box 7.1. Murihiku rocks.



Murihiku rocks comprise one of the constituent 'terranes' of the New Zealand basement rocks (Boxes 5.1, 5.2, 5.3). Their chief distinguishing characteristics are a great thickness of well-stratified, volcanic-derived sandstone and mudstone (more than 15 km thick) and very large scale folding.

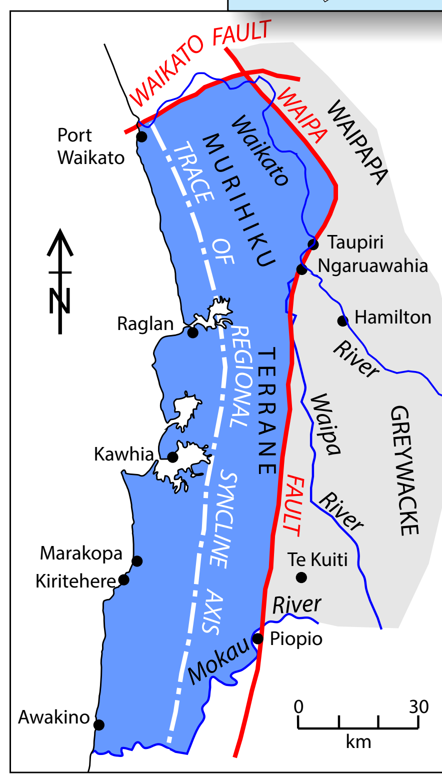
Between the Waikato Fault and the Mokau River, Murihiku rocks and Waipapa Greywacke are concealed in places by younger rocks.

South of the Mokau River both Murihiku and Waipapa rocks are totally concealed by younger rocks until they reappear in Nelson.

The Waikato Fault drops Murihiku rocks 2.5 km to the north, so that they are concealed beneath younger rocks everywhere north of the Waikato River. They extend at least as far north as Hokianga Harbour.

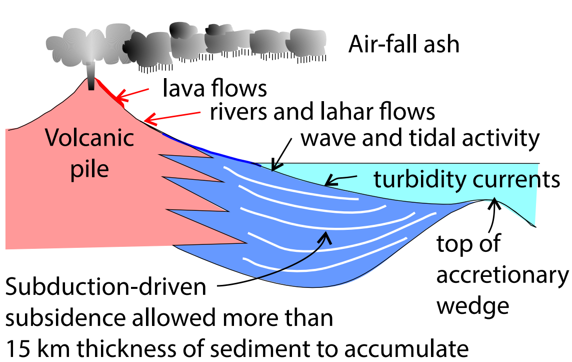
The Waipa Fault separates Murihiku rocks from Waipapa Terrane greywackes.

Murihiku rocks range in age from middle Triassic (240 myr) to latest Jurassic (146 myr). They comprise well-bedded mudstones, sandstones and conglomerates. Fossils are abundant in places. The best places to see Murihiku rocks are Port Waikato and between Kawhia and Kiritehere. You can guarantee seeing the fossil scallop shell *Monotis* on the south side of Kiritehere Beach.

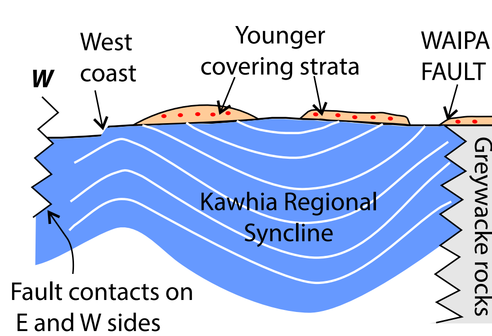


Origin of Murihiku Rocks
They formed over a period of more than 100 million years in a long-lived forearc basin. The basin was part of the subduction system on Zealandia's sector of the Gondwana continental margin. Compare Boxes 6.2 A-C, Fig. 6.7 for the contemporary greywackes, and Boxes 12.1 A, B for the present-day analogue.

Base of forearc basin is not exposed anywhere. The volcanic pile is not known now.



Present Set-up
Cross section showing the present set up.
Scale- 10s of km, vertical scale exaggerated.

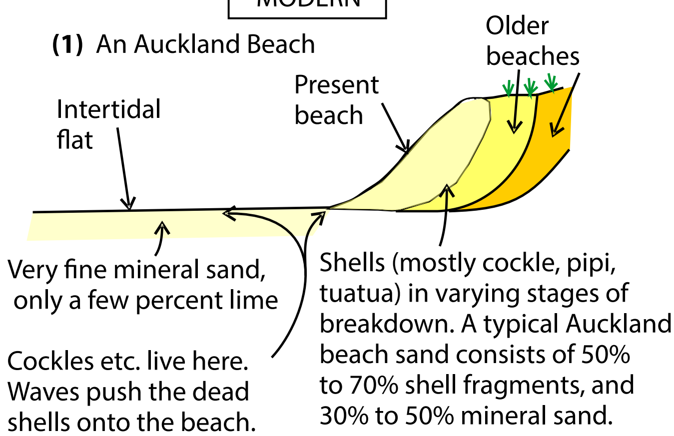


Box 7.2 A. Limestone.

Limestone is a rock made mainly of **lime.** Lime (calcium carbonate, $CaCo_3$) is used by many shellfish, corals, moss animals (Bryozoa), forams (Foraminifera, single-celled microscopic animals) and echinoderms (sea eggs, sea lilies, starfish) to make their shells. Some plants secrete lime (some marine algae and the very tiny but prolific marine Coccolithophorida).
Most of these creatures live in the sea. Their shells and secretions are the raw material for limestone, which is thus a sedimentary rock typically marine in origin.
The term raw material implies that further processing has to take place before we arrive at the end product of limestone. Taking two examples, one modern, one ancient:-

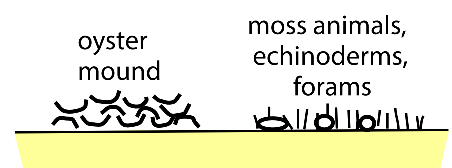
MODERN

(1) An Auckland Beach



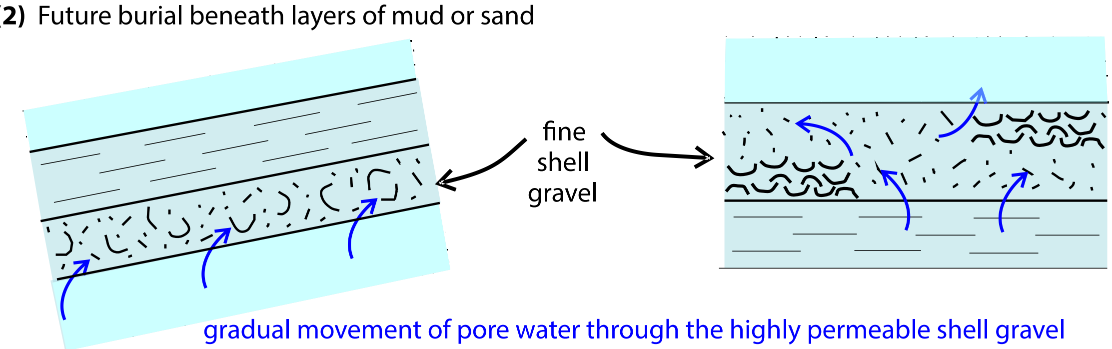
ANCIENT

(1) A continental shelf 30 million years ago, western North Island.



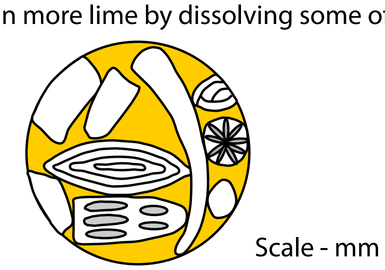
There was a much-reduced supply of mineral sand from the New Zealand land mass at this time (Boxes 4.1, 5.5 B). The sediment that built up on the sea floor included layers of pure shell gravel.

(2) Future burial beneath layers of mud or sand



(3) Pore water is sea water, and contains dissolved lime. It can obtain more lime by dissolving some of the shell material.

When the chemical environment (pH) is suitable, lime slowly crystallises and fills the spaces, both within and between the shell fragments and mineral sand grains. (Lime can also fill the spaces between grains in a layer of mineral sand, forming a sandstone with a cement of lime, as in a concretion (Box 6.5 A).



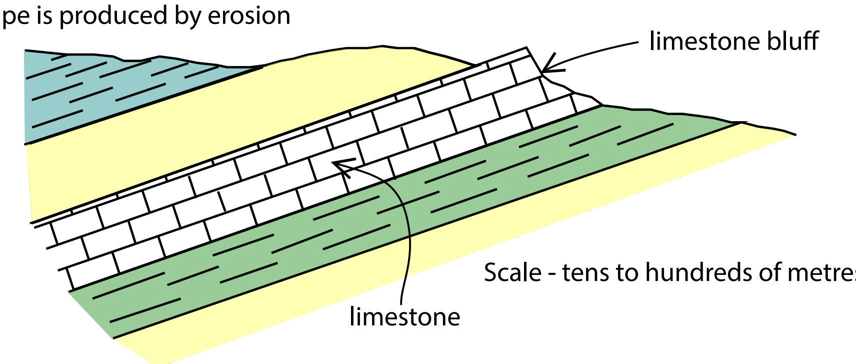
(continued below)

Box 7.2 B. Limestone *(continued).*

(4) The shell gravel is now solid limestone. For reasons that are not well understood, cementation with lime is commonly variable, in a regular, rhythmic fashion.

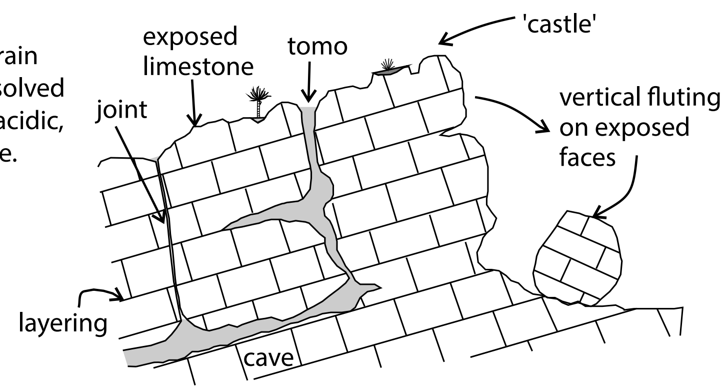
(5) Later, earth movements fold and tilt the pile of strata and carry them above sea level. There may be more than one layer of limestone.

A landscape is produced by erosion



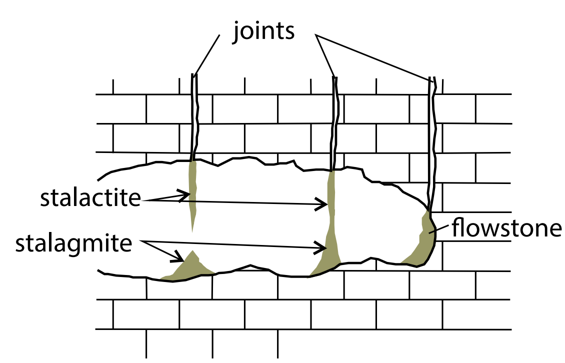
(6) The limestone is now attacked by rain water and soil water. They contain dissolved carbon dioxide (CO_2) and are slightly acidic, thus they slowly dissolve the limestone.

Scale - metres to hundreds of metres.



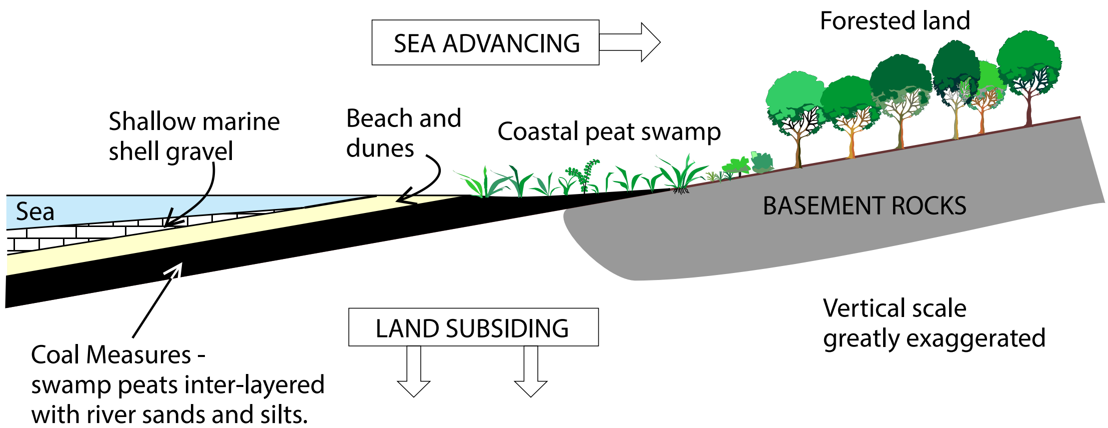
Penetration of water, and hence solution of limestone to form caves, is controlled by two structures - the original layering in the limestone, and a system of joint cracks which develop, generally at right angles to the layering.

(7) Fresh water dripping into the caves is saturated with lime. As the drips hang from the cave ceiling, or trickle down the walls or along the floor, some of the dissolved carbon dioxide is released from the water. This lowers the acidity of the water, which is then less able to hold dissolved lime, and causes a small amount of lime to be deposited. In this way the various cave formations are slowly built up. The same thing can happen where a spring discharges at the surface, causing a build-up of lime (travertine, or calcareous tufa) often around plants.



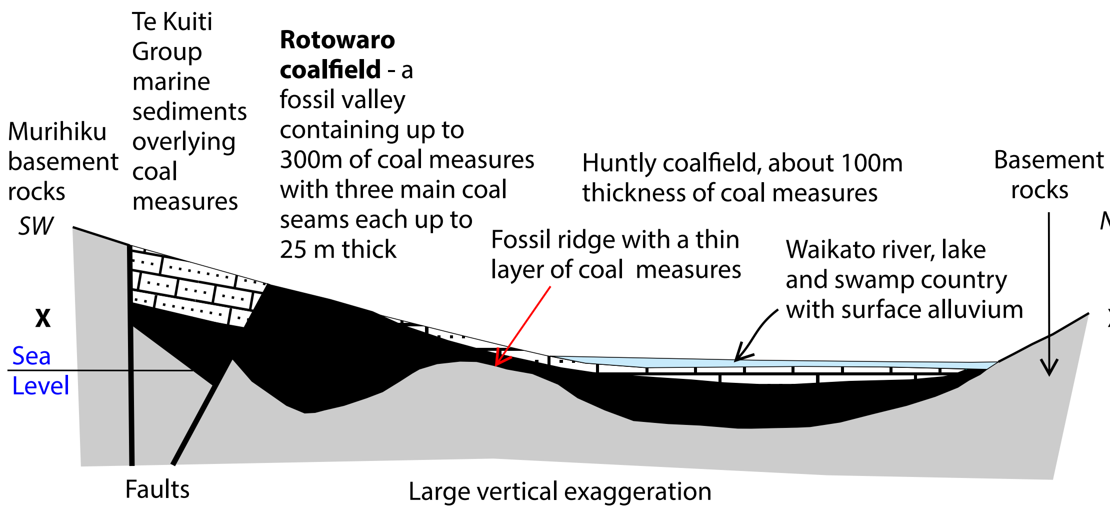
Scale - metres to tens of metres

Box 7.3 A. Waikato Coal Measures around 40 million years ago, and the Waikato Coalfield now. Cross section X – X′ is shown in Box 7.3 B.



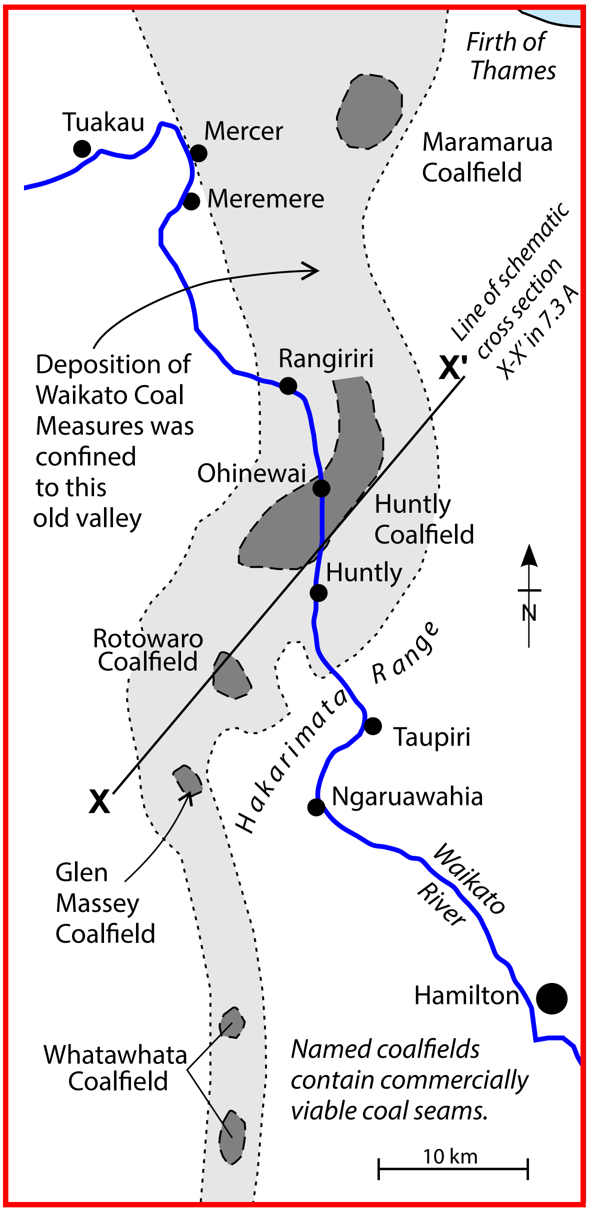
With burial beneath marine sediments, peat is converted to lignite and eventually to coal. Beach and dune sands later become sandstone. Shell gravel later becomes the Te Kuiti Limestone.

THE WAIKATO COALFIELD NOW



Schematic cross section through two coal basins within the Waikato Coalfield. In detail the coalfield is very complicated, and underground mining is difficult. Most mining is now open-cast. The coal is of sub-bituminous rank.

See 'Extent of Waikato Coal Measures' for location of the cross section.



Box 7.3 B. Extent of Waikato Coal Measures. The schematic cross section in Box 7.3.A, lower illustration, is from X – X' diagonally across this map.

Waitemata Basin in time, are similar to the latter. They are flysch strata (Box 6.5 A, B), deposited by south-flowing turbidity currents, but lack the prominent volcanic component that is present in the sandstones and Parnell Grits of the Waitemata Basin (Box 6.7). These strata are well exposed between Te Kuiti and Taumarunui along SH4. To the south, they dip underneath still younger deposits of the Whanganui Basin (Box 10.1).

Volcanic activity

The volcanic story of phase four comes after the formation of the sedimentary basins (Fig. 7.1). Subduction-related arc volcanoes of the period 15–5 myr ago actually extended from northeast to southwest right through Coromandel–Kaimai and South Auckland (Box 5.5 D). They linked the offshore Colville Ridge in the northeast (the oceanic volcanic arc of the time; Chapter 3) with a series of arc volcanoes that lie offshore to the southwest between Raglan Harbour and the North Taranaki Bight, and that are now completely buried underneath continental shelf sediment.

Curiously, while volcanic rocks are plentiful in Coromandel and the Kaimai Range, and also offshore to the west, they are sparse in South Auckland. Here, we have two lines of small andesitic volcanoes, forming the so-called Kiwitahi and Alexandra lineaments. The Kiwitahi Lineament is more than 100 km long, extending from the eastern end of Waiheke Island in the Hauraki Gulf to near Cambridge. The volcanoes here have lost their cone shape and are not obvious to the eye, but there is an interesting age progression whereby they get younger from north to south. Ages are shown on Fig. 7.1, and range from 14 myr on Waiheke to 6 myr near Cambridge. An age progression like that is thought to record trench rollback, reflecting subduction zone retreat (Chapter 2).

The Alexandra Lineament, on the other hand, is much younger, shorter, more obvious and differently oriented, and has no age progression. Known ages of Mts Karioi and Pirongia, as well as other volcanoes along this line, are all around 2 myr, which places them within the timeframe of the existing subduction/volcanic system. The volcanoes are andesitic in composition, which relates them to an arc, and they retain their cone shapes, though these are somewhat degraded by erosion. In relation to the active arc of this time (see below), they are behind-arc in position. This may indicate that they originated in the same way as the currently active Mt Taranaki, from a magma source on the subducting slab that is much deeper than the normal one – a depth of 200 km, compared with the usual 100 km.

The main discussion of New Zealand's active volcanic arc, the Taupo Volcanic Zone (TVZ), is reserved for Chapter 9, but here in South Auckland we have some of its older manifestations. The three andesitic cones of Maungatautari (near Cambridge), and Titiraupenga and Pureora (near Lake Taupo), are all dated around 2 myr, the same as the Alexandra Lineament. They are the 'normal' arc to the Alexandra 'behind-arc'. This arc position was short-lived, however, and records part of the shift

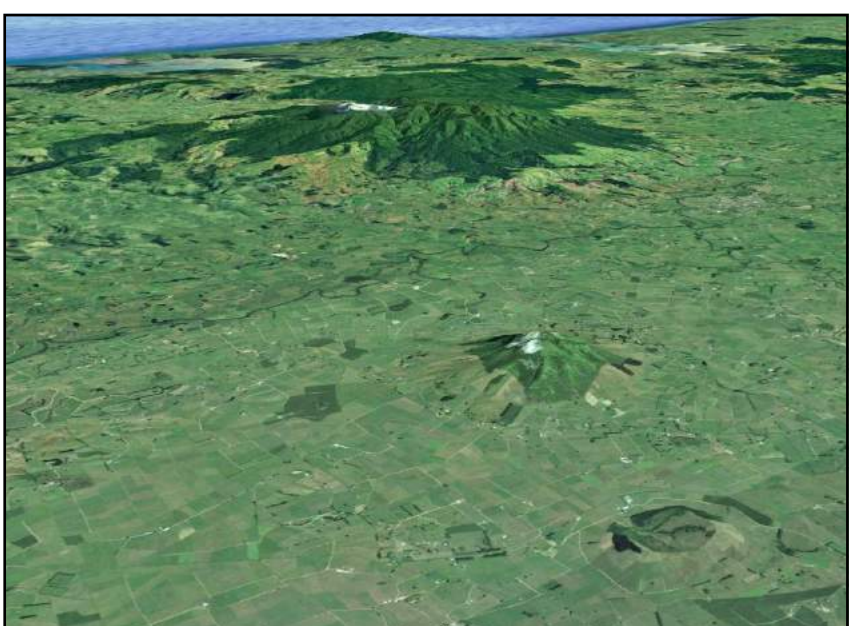


Fig. 7.2. View west across the 2 myr-old Alexandra lineament of four volcanoes southwest of Hamilton. The cones, from bottom right, are Te Kawa, Kakepuku, Pirongia (forested) and on the coast Karioi (near Raglan). Photograph courtesy of Google Earth.

from the Colville–Taranaki Arc (15–5 myr; see above) to the Kermadec–Ruapehu Arc that is active today. In addition to the cones, the southeastern portion of the region is blanketed by welded ignimbrite flows that originated in the older TVZ around 250,000 years ago (Fig. 7.3).

Still in the behind-arc domain, we have four basaltic volcanic fields similar to Auckland's. Aged between 5 myr and 0.5 myr, they are the Otete (around Raglan), Ngatutura (south of Port Waikato), Awhitu (offshore from Awhitu Peninsula and buried by sediment) and South Auckland fields.

Finally on the subject of volcanoes, there are three more andesitic volcanoes, aged 4–5 myr this time, that occur on the west coast south of Kawhia: the Orangiwhau Volcanics. They have lost their cone shapes, and seem to record an early stage in the progression from the wholly buried offshore Taranaki portion of the Colville–Taranaki Arc to the early TVZ, 2 myr ago.

Fig. 7.3. Bluffs of welded ignimbrite at Hinuera are sawn into blocks for use as a building stone.
Photographer Adrian Pittari.

Fig. 7.4. This columnar-jointed sea stack and adjacent cliff at Ngatutura Point, south of Port Waikato, are eroded remnants of a volcanic plug, part of the behind-arc Ngatutura Volcanic Field that was active 1.8–1.5 myr ago.
Photographer Bruce Hayward.

The present disposition of all these different rock units, and thus their appearance or otherwise at the surface, is governed by the regional geological structure – especially a series of fault-angle depressions (Box 6.6 G) – and by the ever-present surface weathering and erosion. Cover strata tend to be preserved in fault angles in the north, around the southern Hunua Ranges (including the Ohinewai Basin), and in broader sag basins like the Hamilton Basin. In the far south of the South Auckland region, all the non-volcanic rocks start to bend southwards into the Whanganui Basin, as discussed in Chapter 10.

Chapter 8

Coromandel Volcanic Zone

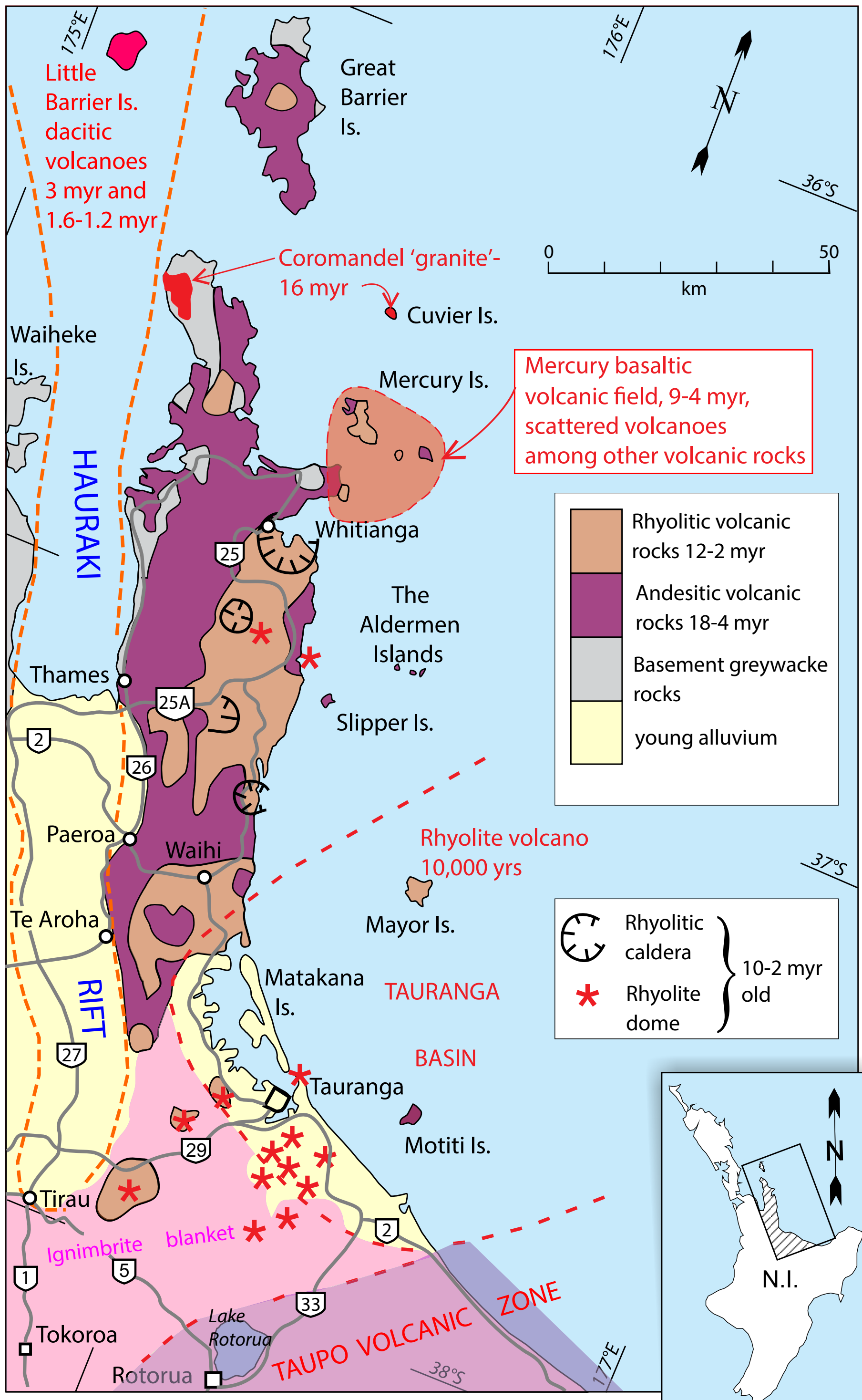


Fig. 8.1. Simplified geology of the Coromandel Peninsula and western Bay of Plenty.

Great Barrier Island, Coromandel Peninsula and the Kaimai Range are sometimes referred to as the Coromandel Volcanic Zone (CVZ), giving them a status apparently equivalent to the Taupo Volcanic Zone (TVZ). The TVZ is New Zealand's currently active volcanic arc (Chapter 9), and is 2 myr old. The CVZ does, indeed, include parts of the predecessors of the TVZ, but the nomenclature is confusing in that it has never been a single coherent volcanic arc in its own right. Activity spanned about 16 myr (18–2 myr ago), and is more complicated than that in the TVZ because it contains portions of two separate volcanic arcs that happen to overlap on the ground. Box 5.5.D, illustrating the geological development of New Zealand, shows the volcanic arcs that preceded the TVZ. The older, northwest-trending arc of eastern Northland (Chapter 6) was active from around 24 myr to 15 myr ago. Towards the end of its life it extended as far south as Great Barrier Island and northern Coromandel, where it was subsequently overprinted by a younger arc trending at right angles to the first one. This was the northeast-trending Colville–Taranaki Arc, which was active between 15 myr and 5 myr ago (Chapter 7). The CVZ is situated at the crossroads of these two arc systems, and therefore the younger volcanic rocks were superimposed at right angles across the older. Globally this is a rare situation. The Colville-Taranaki Arc interpretation presented in this chapter is still the subject of much debate and not currently accepted by all those who study the Coromandel Volcanic Zone volcanoes.

The geology of this region is not easily unravelled. Adding to the great variety of volcanic rocks, there is poor rock exposure owing to extensive rainforest cover and deep chemical weathering as discussed in Chapter 6. A further obscuring factor is hydrothermal alteration of many of the older rocks, which comprises mineral changes in the rocks and was caused by the underground circulation of hot water while volcanoes were still active. Like chemical weathering, hydrothermal alteration tends to change rocks to clay, and as such it is difficult to separate the effects of the two processes without detailed study. A colleague who specialises in hydrothermal alteration divides rocks into two kinds: nicely altered and badly fresh. It all depends on one's perspective! The rocks involved are both volcanic and non-volcanic, the latter comprising basement greywacke rocks that lie underneath the volcanic rocks.

The result of this chemical weathering and hydrothermal alteration is that road cuttings in the region commonly show clays in shades of yellow, red and brown, indicating the presence of water-bearing iron oxides (rust) derived from iron-rich minerals. Shades of purple to black, meanwhile, indicate the presence of manganese oxides. It is not easy to distinguish the original rocks, although one feature of greywacke that does survive early stages of weathering or alteration (but not more advanced stages) is its closely jointed character.

Hydrothermal alteration does have one mitigating feature. It was accompanied by gold and silver mineralisation in quartz veins, which created great wealth in the region during the late nineteenth and early twentieth centuries. Mining activity has left many marks on the region, and at the time of writing in 2009 continued at one gold mine in Waihi. Lead and zinc were also mined, at Te Aroha.

Geological History

The geological history of the CVZ is best thought of as encompassing five stages (Box 5.5 D). First came the basement greywacke rocks (Chapter 5), shared with most of New Zealand and underpinning everything else. Next, following New Zealand's separation from Gondwana 80 myr ago, the area shared in the Zealandia mini-continent's long period of erosion and gradual subsidence (phase three of New Zealand's development; Chapter 4), and by 30 myr ago was covered by shallow marine sediments of the Te Kuiti limestone association. Only one small patch of these limestones survived subsequent events (at Torehina in north Coromandel), and there are no significant coal measures here.

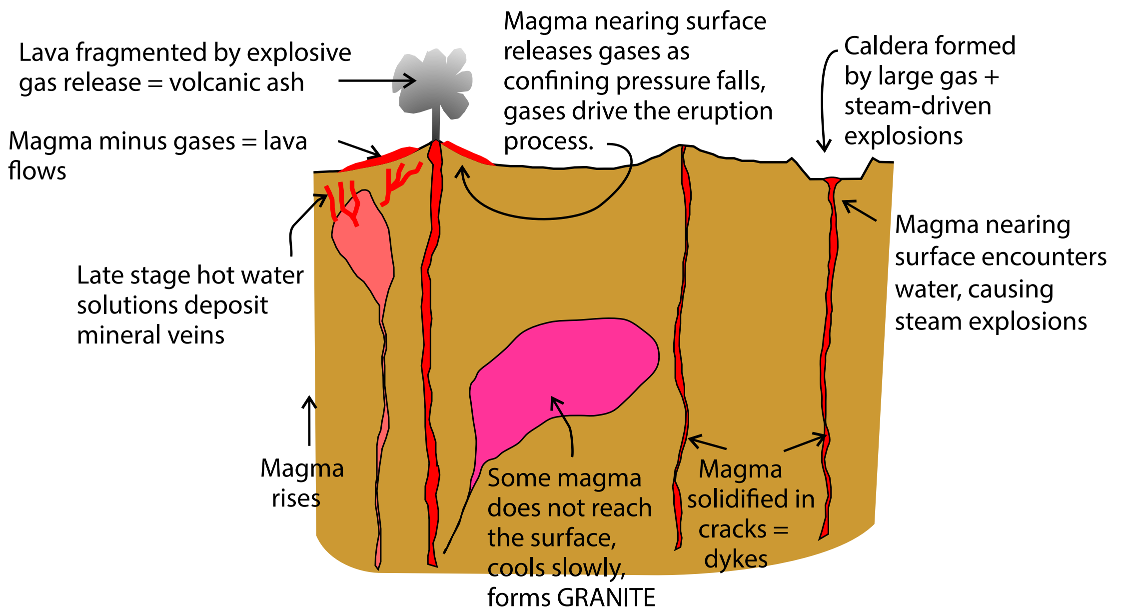
Around 25 myr ago, the region entered its third stage of formation when Coromandel became involved in the maelstrom of tectonic and volcanic events related to the development of the Pacific–Australian plate boundary subduction zone (phase four of New Zealand's overall geological development; Chapter 3). Initial uplift and erosion, which removed most of the Te Kuiti-type sedimentary rocks, was followed by subsidence of a deep basin in northern Coromandel. This basin was actually the eastern portion of the Waitemata Basin (Chapter 6). Interestingly, there are Parnell Grit volcanic debris flow deposits (Box 6.7) among the strata, raising the question of whether these debris flows extended all the way from the Waitakere or Kaipara volcanoes, requiring a continuous downhill slope for 120 km, or were derived locally from the northern Coromandel volcano whose lavas appear to be younger and overlie the basin deposits. These rocks are seen at Fletcher Bay, at the northern tip of the Coromandel Peninsula.

Subduction-related volcanic activity on the northwest-trending arc of eastern Northland (see above) extended south to Great Barrier and northern Coromandel only about 18 myr ago. From 18 myr to around 4 myr ago the volcanoes – initially from the Northland Arc, later from the Colville–Taranaki Arc – were andesitic in nature (Box 9.2 A). There are two exposed sub-volcanic 'granites' from this period, at Paritu and Cuvier Island, both of which are about 16 myr old. Paritu 'granite' (more commonly known as Coromandel Granite) forms most of Mt Moehau. Quotes are used because technically the rock is not granite, as its chemical composition is not quite right (Box 8.1). Strictly speaking, these 'granites' vary between quartz diorite, tonalite and granodiorite, all of which are intrusive equivalents of andesite lava. These were sub-volcanic plutons, bodies of magma that did not make it to the surface and cooled slowly, allowing centimetre-size crystals to grow. Plutonic rocks are rare in the North Island, but common in the South Island (Box 8.1).

Coromandel Granite was used extensively in public buildings of the early twentieth century (e.g. the old Central Post Office in Auckland and the Parliament Buildings in Wellington). It is technically tonalite, a dark rock made of the minerals hornblende and black mica, white plagioclase feldspar, and small quantities of colourless quartz, like window glass. The old jetty in Paritu Bay is built of the rock, which is exposed all around the bay and in Fantail Bay to the north. Either at Paritu Bay or in buildings made from Coromandel Granite, look for xenoliths. These are dark blocks of finer-grained rock enclosed in the tonalite, and merging into it to varying extents – they were plucked from the walls of the expanding magma chamber, baked, and melted to varying degrees. When melted completely, of course, they became part of the magma.

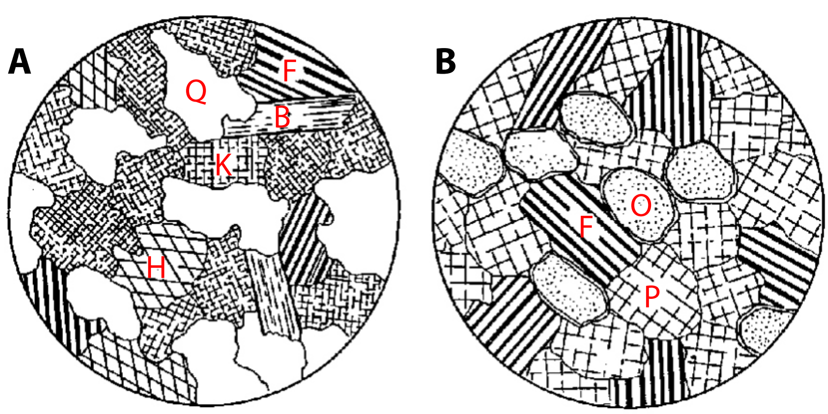
Box 8.1. Granite – its place in the process.

Magma is molten rock including dissolved gases. 'Granite" is used here as a catch-all term for all rocks formed from magma and made of large crystals. An alternative name is "plutonic", from Pluto, god of the underworld.



Magma source at depth of 30km plus

'Granite' contains crystals that are visible to the naked eye. They had time to grow while the magma slowly cooled. Many different combinations of minerals occur, reflecting differences in chemical composition of magma. The commonest minerals are quartz, feldspar and mica. Granites are commonly used as decorative facing stones and paving slabs. Go on a granite hunt in your local downtown or cemetery.



Q = quartz (colourless, like glass)

K = potash feldspar (commonly pink or red)

F = plagioclase feldspar

B = biotite mica or white mica

H = hornblende

P = pyroxene

O = olivine

Polished surfaces on two kinds of 'granite'.
A - true granite with free quartz, and an overall silica content of more than 70%.
B - gabbro, darker coloured with no free quartz and an overall silica content of about 50%.
The volcanic equivalent of granite is rhyolite (Boxes 9.3 A-C), and of gabbro is basalt (Boxes 6.9 A, B)
The actual size of the crystals may range from 1mm to several cm.

Fig. 8.2. This disused stone jetty at Paritu, northern Coromandel Peninsula, is made of Coromandel Granite blocks that were shipped out from here to many part sof New Zealand for use in many public buildings. Photographer Bruce Hayward.

Heat from the Paritu magma baked the surrounding greywacke very thoroughly, causing new minerals to form (a process known as 'contact metamorphism'). The contact between the 'granite' and baked greywacke can be seen just south of the old Paritu stone jetty.

Following the switch from the Northland Arc to the Colville–Taranaki Arc 15 myr ago, the type of andesitic volcanic activity did not change, but the volcanoes extended further south into southern Coromandel and the Kaimai Range. Table Mountain's flat top is the upper surface of a young andesitic crater lake.

The fourth stage of geological development began about 12 myr ago, when the lava chemistry changed to include rhyolitic – i.e. to a greater silica content. Volcanic activity shifted to the eastern side of the peninsula, and was accompanied by the formation of several calderas (Fig. 8.1 and below). Remains of this activity can be seen in the form of rhyolite domes and ignimbrite deposits on the east coast. The pattern of volcanic activity in this area was identical to that taking place at the present time in the central portion of the TVZ (Chapter 9). Extensive hydrothermal activity also took place at this time. It caused much alteration of rocks towards clay and much silicification of rocks (enrichment in silica), and it deposited mineral-bearing quartz veins in cracks. Residual heat from volcanic activity still drives the hot springs of Hot Water Beach and Te Aroha.

Fig. 8.3. The flat top of Table Mountain, in the middle of the Coromandel Range, is the original surface of solidified andesite lava that cooled in a crater lake 8 myr ago. Photographer Bruce Hayward.

In the final stage of formation, behind-arc basaltic activity built localised volcanic fields, as in Northland, Auckland and South Auckland (Chapter 6). The Mercury Islands Basaltic Volcanic Field (Box 6.8) was active from 9 myr to 4 myr ago, possibly in two separate periods of activity in the same area, as happened in Northland. Thus it overlapped on the ground with primary rhyolitic volcanism. Such an association of volcanoes of contrasting chemistry (silica contents of 50% and 75%) is found elsewhere and is called a bimodal association. It occurs in a small way in the TVZ (Chapter 9) and Northland (Chapter 6).

Interpreting the Volcanic Rocks and Features of the CVZ

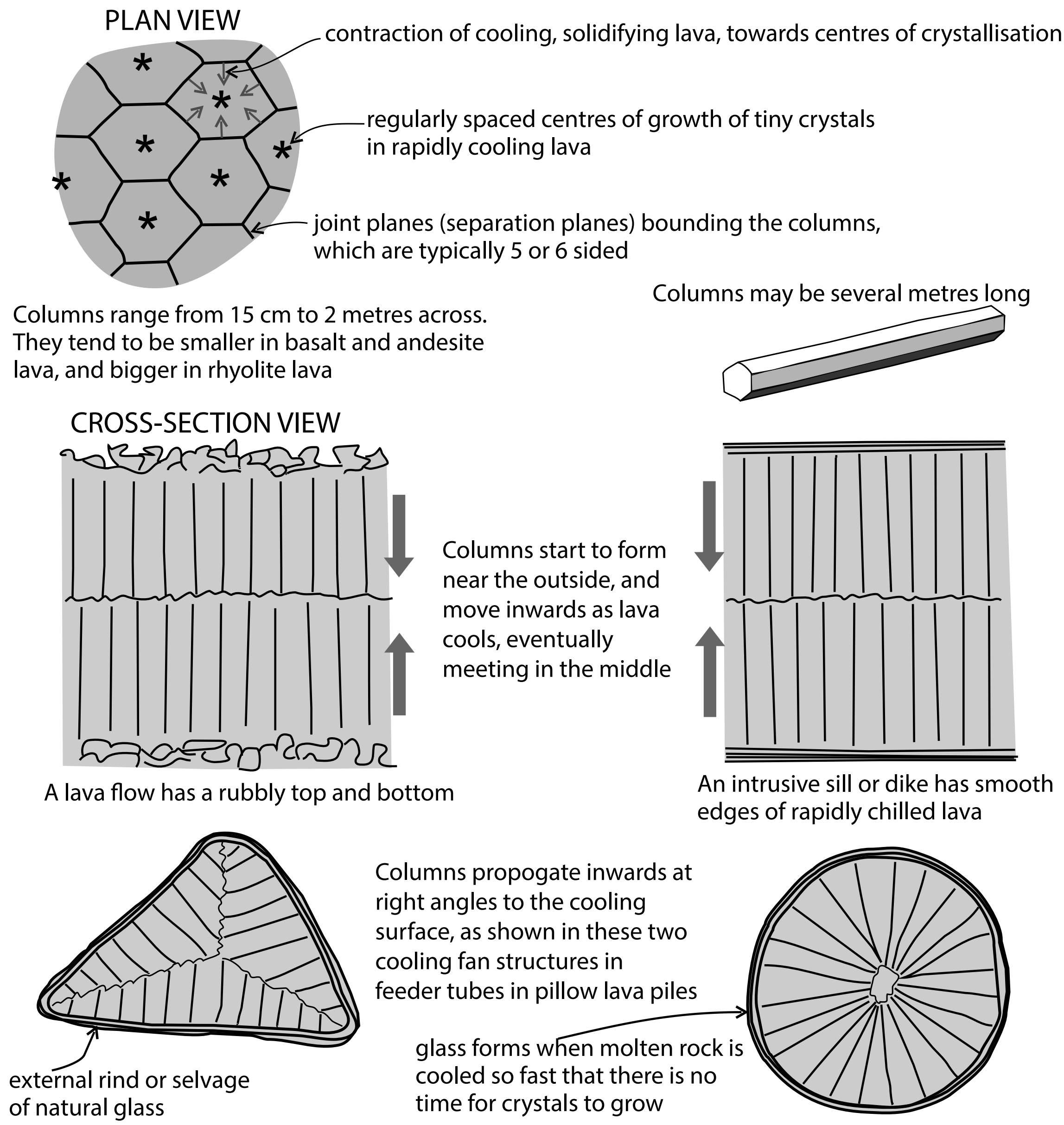
The region's dark-coloured andesitic rocks originated in exactly the way that we see andesitic lava flows and breccias (deposits of angular stones, pronounced 'bretchias') forming on Mts Tongariro and Ruapehu today (Chapter 9). However, the original volcanic landforms have long succumbed to erosion, and the rocks have suffered changes that have tended to obscure their original features, as noted earlier. There has also been tectonic disturbance, leading to folding and faulting (Boxes 6.6 A–G). To interpret the rock exposures, look for columnar-jointed lava flows (Box 8.2) and breccia textures (Fig. 8.4).

Some of the pale-coloured rhyolitic rocks on the east coast of the Coromandel Peninsula have not been altered and are easily interpreted (Box 9.3 A–C). Examples of such rocks include those found on the southern shores of Mercury Bay as far as Hot Water Beach, within the Whitianga caldera.

Fig. 8.4. Weakly stratified andesite breccia is a common rock type in the Coromandel Ranges. The breccia seen here in The Pinnacles, east of Fletchers Bay, is the eroded remnants of a stratovolcano that erupted in this vicinity 18–17 myr ago. Photographer Bruce Hayward.

Box 8.2. Columnar jointing in lava.

Lava which cooled evenly, in the interior of flows and small intrusions (sills and dikes), commonly displays columnar jointing. That is, it is broken into parallel 5-sided or 6-sided columns. Columns may be many times longer than they are wide. The most famous example is probably the Giant's Causeway in Northern Ireland. Columns are cooling-crystallisation-shrinkage structures, which propagate from the outside towards the middle of the body of lava, forming at right angles to the cooling surface.



Silicified wood

One common by-product of terrestrial volcanism is petrified or silicified wood, formed when wood from forests is incorporated into lahar flows (Boxes 9.2 A–C) and ignimbrite deposits (Boxes 9.3 A–C). Volcanic glass – lava that was chilled so quickly that there was no time for crystals to grow – is abundant in these deposits, and is chemically unstable. It starts to break down soon after deposition, and in the process releases dissolved silica. There is an affinity between silica and cellulose, such that silica replaces the cellulose in the wood, molecule by molecule, and hence can perfectly preserve the open, cellular structure of the wood. If the supply of silica continues, it can fill all remaining spaces and destroy the cellular structure, producing opal. You are likely to come across silicified wood all over Coromandel, as it is durable and lasts well in stream and beach gravels. It is popular with rockhounds, who cut and polish the stones they find.

Calderas

These are volcanic collapse features that are larger than erupting craters, and they form after very large eruptions. They are easier to recognise in the younger TVZ than in the CVZ, simply because of all the weathering and erosion that has taken place in the latter over the past 5 myr. However, mapping geologists have recognised some old calderas in the region; these are shown on Fig. 8.1.

Pinnacle landforms

These landforms are common throughout the Coromandel Peninsula, but less so in the Kaimai Range. Sometimes they appear to be old volcanic plugs (lava that hardened in the feeder pipe), and sometimes they are the remaining fragments of once extensive sheets of resistant lava (Box 12.6). At any rate, they are made of rock that resists erosion and has prominent vertical joints or fractures.

Castle Rock near Whangapoua is one of the most striking landforms in Coromandel, and consists of several pinnacles of dacite lava (midway in chemical composition between andesite and rhyolite). It is unclear whether it represents a single volcanic plug or several feeder dikes. The feature is dated at 11–12 myr, and therefore belongs to the region's early andesitic phase.

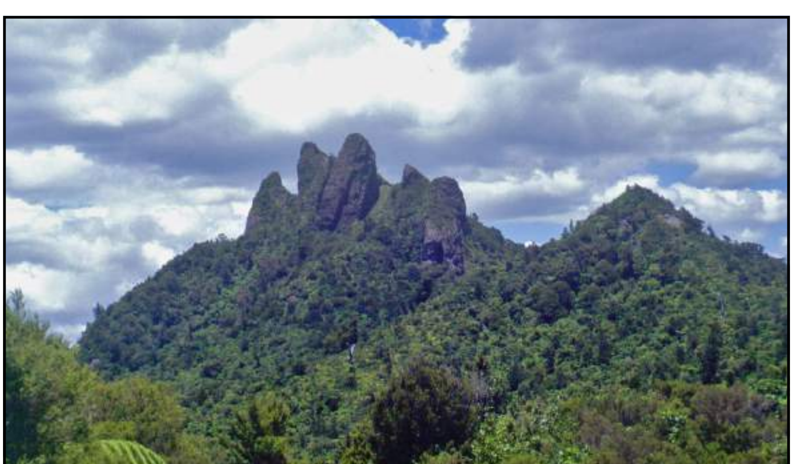


Fig. 8.5. The jagged pinnacles of 520 m-high Castle Rock, east of Coromandel township, is composed of vertical dikes thought to be the compound plug of a large dacite stratovolcano that erupted 11–12 myr ago. Photographer Bruce Hayward.

Coastal landforms in rhyolitic rocks

The rhyolitic lavas and ignimbrite deposits along Coromandel's east coast have many joint and fault defects that are exploited by wave action. The result has been the formation of a range of caves, blowholes, arches and seastacks, notably along the coast south of Whitianga.

Mayor Island

Mayor Island volcano is unique. It is much younger than the Coromandel and Kaimai Range volcanoes, at a mere 10,000 years. Its age suggests that it should be included in the TVZ, but it is well outside this region. Although it is rhyolitic, its chemical composition is not typical, containing more sodium and potassium than is normal. Its position within the developing Tauranga Basin may be significant. The outstanding geological feature of the island is the abundance of fine black natural glass (obsidian), which made it of great cultural and practical importance to Maori.

Kaimai Range, Mamaku Plateau and Tauranga structural depression

The rugged, pinnacled Coromandel Range merges southwards into the Kaimai Range, and then, south of Tauranga, into the less dissected Kaimai–Mamaku Plateau. Here, the products of the two successive volcanic arcs, the Colville–Taranaki and Kermadec–Taupo (TVZ) arcs, overlap in a messy sort of way. Ignimbrite sheets emanating from the older parts of the TVZ have buried the southernmost part of the CVZ, forming the Kaimai Plateau. The Mamaku Plateau, meanwhile, was formed by the Mamaku Ignimbrite deposit from the Rotorua caldera in the TVZ (Chapter 9). Individual rhyolite domes are more easily distinguished here, including Paku Island near Tairua, and Bowentown at the northern head of Tauranga Harbour.

Young structural depressions influence the landscape in this southern part of the region. Waihi sits in the Waihi Basin, which measures approximately 12 km east–west and 5 km north–south. It is filled with rhyolitic rocks and alluvial sediment. As shown in 8.a, the Tauranga Basin is a semi-circular structure 40 km long and 15 km wide, located between the Kaimai Range and the sea. It is actually the southern end of a submarine structural basin that extends more than 200 km to the northeast. This basin is part of the story of the opening of the Bay of Plenty (Chapter 12).

Tauranga Harbour, like most harbours in New Zealand, is a drowned river valley (Box 6.12), in this case a whole series of them. This large feature reflects the extent of the Tauranga structural depression. Matakana Island is a post-glacial offshore barrier constructed entirely of sand. In fact, a prominent feature of the whole Katikati–Tauranga area is intense alluviation – sediment galore derived from the erosion of voluminous young volcanic deposits. Thick, unconsolidated sediment holds much groundwater, fed from the ranges; some of this water at deeper levels is warm enough (35–55°C) to be used for recreation. Combined with the moist mid-latitude climate, the situation is ideal for the intensive horticulture practised locally.

Mt Maunganui

'The Mount', the iconic cone at the entrance to Tauranga Harbour, is one of a cluster of rhyolite domes aged 2 myr and more that form the southern end of the CVZ (Fig. 8.1). Walks around the Mount

Fig. 8.6. Oblique view south over the Tauranga Basin, between Kaimai Ranges and the sea, and partly occupied by the Tauranga Harbour. The young sand dune barrier that forms Matakana Island mostly accumulated over the last 7000 yrs. Photograph courtesy of Google Earth.

Fig. 8.7. Mt Maunganui is a rhyolite dome that was extruded onto the surface at the southern end of the Coromandel Volcanic Arc about 4 myrs ago. Photographer Bruce Hayward.

reveal the flow-banded rhyolites making up the dome. Being 4 myr old, the dome has been weathered and eroded to some extent, and it has also been capped by younger ash layers from the TVZ.

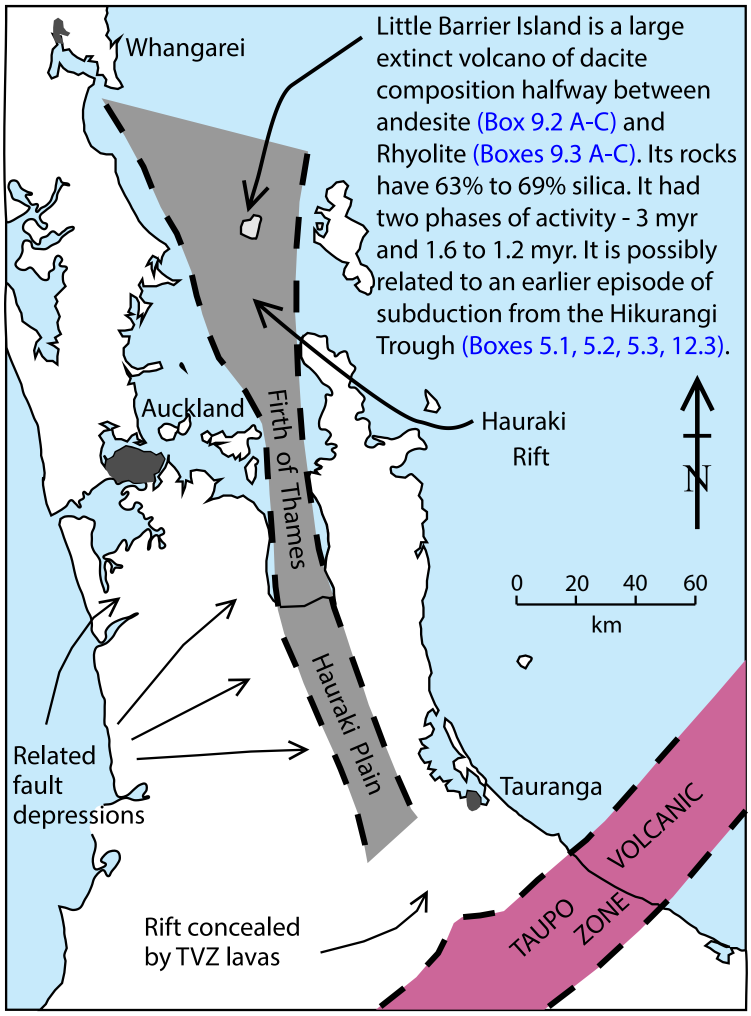
The Hauraki Rift

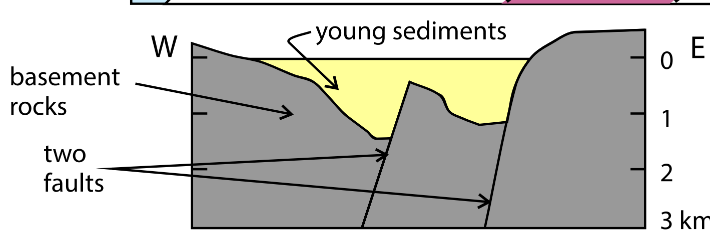
The North Island's only rift valley separates the CVZ from Northland, Auckland and South Auckland. It is a scenically stunning major structure, more than 200 km long. Its full length is not known, because the southern end is infilled by volcanic sediments and buried by ignimbrite flows from the TVZ. It is active today, with frequent shallow earthquakes and hot springs, and forms the central part of the Hauraki Gulf, the Firth of Thames and the Hauraki Plains. It is described more fully in Box 8.3.

Fig. 8.8. Oblique view south down the Hauraki Rift with the Firth of Thames in the foreground and the sediment-filled Hauraki Plains portion beyond. The Coromandel and Kaimai Ranges lie to the east (left) and Hunua Ranges to the west. Photograph courtesy of Google Earth.

Box 8.3. Hauraki Rift.

A rift is an elongate depression caused by active or recently active faulting, and comprising some combination of fault-angle depressions. The faults are usually normal, i.e. extensional, but they can also be reverse, i.e. shortening (Box 6.6 C).





Representative cross section of the rift. It is a double fault-angle depression.

Hauraki Rift is still actively forming, as indicated by frequent shallow earthquakes, active fault traces, and hot springs. It is extensional. Its age is uncertain. It may be several million years old.

See also Box 9.7, Waikato River.

Some special volcanoes

As mentioned above, the Hauraki Rift Valley extends northwards through the Hauraki Gulf. It is not easy to see there, of course, but it is known to reach nearly all the way to Whangarei. There is a volcanic point of interest here in the form of Little Barrier Island, a prominent nature reserve that sits in the middle of the rift. Little Barrier is a double volcano – a young volcano dumped on top of an older one. Here, an older dacitic dome (close in composition to rhyolite; Boxes 9.3 A–C), 3 myr old, is overlain by a more voluminous dacitic stratocone, analogous to Mt Tauhara at Taupo (Chapter 9) and dated at 1.2–1.6 myr.

Geologically, Little Barrier volcano is a conundrum. Its rock chemistry relates it to a subduction zone, but 1.5 myr ago the subduction zone was located in its present position and the active volcanic arc was the current one, the TVZ. Little Barrier lies 220 km northwest of the TVZ, which should be far beyond the range of these kinds of lavas. It is possible that the lavas were residual, left over from the earlier volcanic arc prior to 5 myr (Box 2.3).

8.9. Little Barrier Island is a slightly eroded 1.5 myr old dacite stratovolcano that erupted on top of a 3 myr old dacite dome in the middle of the Hauraki Rift.
Photographer Bruce Hayward.

To add to the Little Barrier volcanic conundrum, there are two more dacite dome volcanoes in Northland. Mts Parakiore (800,000 years) and Hikurangi (1.25 myr) are located just north of Whangarei. Their age, rock chemistry and, presumably, origin are very similar to Little Barrier. To compound the mystery, they are even further from the TVZ – 320 km northwest in fact. And to cap it off, while the two Whangarei dacitic domes were erupting, completely different lavas (the Whangarei Basaltic Volcanic Field) were erupting around them.

Chapter 9

The Taupo Volcanic Zone

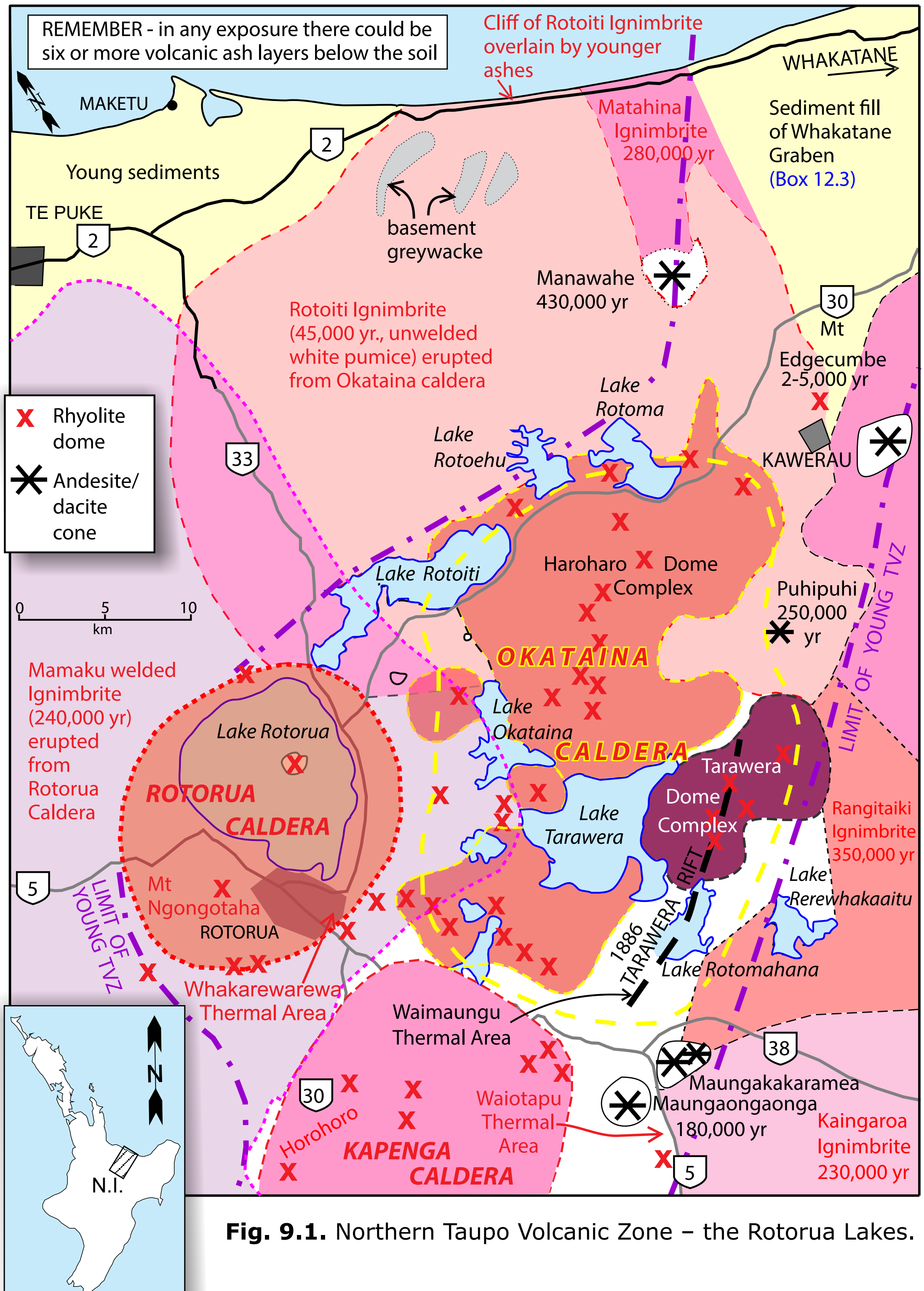


Fig. 9.1. Northern Taupo Volcanic Zone – the Rotorua Lakes.

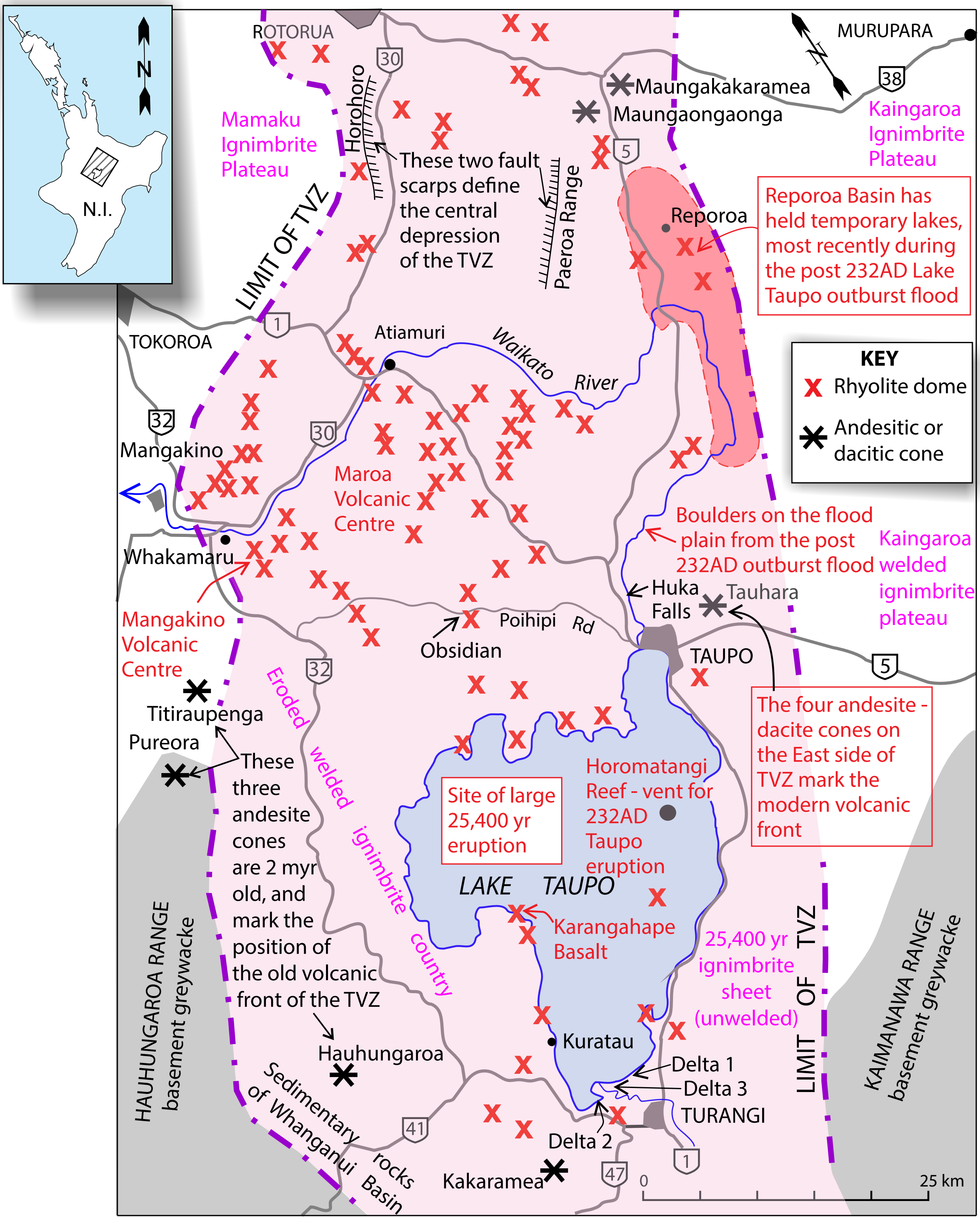


Fig. 9.2 A. Central Taupo Volcanic Zone. Preamble to accompany this map is on following page.

Information you need to make sense of the central Taupo Volcanic Zone (TVZ) – Rotorua to Turangi – in addition to Boxes 9.1, 9.2 A–C, 9.3 A–C.

1. North of Taupo the extent of the younger TVZ (less than 350,000 years) is defined quite closely by the large number of rhyolite domes (see map). Each of these is a separate vent, and the zone is defined as the area of active and recently active vents. South of Taupo there are fewer domes, and the limits of the zone are shown best by the faults which bound the uplifted greywacke basement rocks of the Kaimanawa and Hauhungaroa Ranges. The TVZ is actually a zone of crustal stretching and subsidence, on account of the widening (about 7 mm per year).
2. The area between Tokoroa and Taupo contains two dense clusters of domes, marking the western rim of the buried Whakamaru caldera and the Maroa Volcanic Centre. The domes in some cases have coalesced, forming high plateaux of rhyolite lava.
3. The zone of currently active vents (currently active = the last 50,000 years) is from 20 to 50 km wide (why is it where it is? – see Box 12.1 A). It is flanked on both sides by extensive plateaux made of welded ignimbrite sheets (Boxes 9.1, 9.3 A–C). On the east side the Kaingaroa Plateau is not much dissected by erosion, so is planar surfaced. These ignimbrites were erupted from the Reporoa caldera, north of Taupo (Box 9.1) 240,000 years ago, and have a volume of about 100 cubic kilometres. In the northwest the Mamaku Plateau is also little dissected. These ignimbrites were erupted from the Rotorua caldera 220,000 years ago, and have a volume between 100 and 300 cubic kilometres.
4. Welded ignimbrite – recognised by its crude vertical columns one or two metres wide – rests directly on basement greywacke rocks in many places, and underlies the more recent ignimbrite deposits which are not welded. It can be seen in many places, and where it is being eroded it forms Stonehenge-like clusters of detached and fallen columns (e.g. West Taupo Road, SH32, and the Mangakino – Kihikihi road). It forms spectacular river cliffs by the Waikato River dams downstream from Mangakino, where the river is deeply incised into the ignimbrite plateau (Box 9.7).
5. The Mangakino caldera is the oldest one in the TVZ, and was active around 1.6 myr ago, and again from 1.2 to 1 myr ago. Its centre was where Mangakino township is now located, i.e. it is outside the present TVZ, and tells us that the TVZ has migrated eastwards with time, but the caldera landform is long-since buried beneath younger deposits. The old ignimbrites that erupted from the Mangakino caldera are exposed at the surface in the area to the west of the Central TVZ map, where they extend from Morrinsville in the north, through Kapuni and Arapuni to Mount Pureora, a distance of more than 100 km. They include the well-known Hinuera building stone, a lightly welded ignimbrite (look for the white, rounded, unflattened pumice fragments in the stone) which is quarried near Lake Karapiro. One of Mangakino caldera's products, the 1 myr Rocky Hill Ignimbrite, has the distinction of being the world's most extensive known ignimbrite, being preserved from Auckland to Cape Kidnappers on the east coast, a distance of 360 km.
6. The Mangakino caldera has been infilled by younger ignimbrites and overprinted by numerous rhyolite domes, all belonging to the Whakamaru caldera and dating from around 330,000 years.
7. There are two key events to recognise in the more recent, non-welded ignimbrites. Immediately beneath the soil, pumice deposits of Persil whiteness belong to the most recent event, the Taupo eruption of approximately 232AD. These vary enormously in thickness and coarseness from place to place. There is a characteristic sequence of layers, recording stages of the eruption.
8. The 25,400 year event (it has various names, including Oruanui) was an enormous eruption from a vent in the northwestern bays of Lake Taupo; the bays are the 25 kyr (short for thousand year) caldera collapse. Several hundred cubic kilometres of fine-grained pumice were erupted, and air-borne deposits are found all over the country, including the Chatham Islands where it is 15 cm thick. However, the ignimbrite sheet itself covered an area smaller than the map area, but is tens of metres thick. Where exposed in cliffs or cuttings it is pinkish coloured, fine-grained with maximum size of scattered pumice fragments around 1 cm, and quite massive and unstructured. Like all big ignimbrite eruptions it filled valleys and created a new desert landscape; exposures are commonly associated with sand dunes and stream deposits resulting from the surface reworking of ignimbrite material. Being older than the Taupo ash, the 25 kyr deposit underlies it.
9. Separating the 25 kyr and the Taupo eruptions there were about 30 other small eruptions from northern Lake Taupo. You will see their deposits in many cuttings. A good locality is opposite the De Brett Thermal Hotel, on SH5 where it climbs away from Lake Taupo. A section here was once cleared and labelled, but has deteriorated.

Fig. 9.2 B contd. Central Taupo Volcanic Zone – preamble.

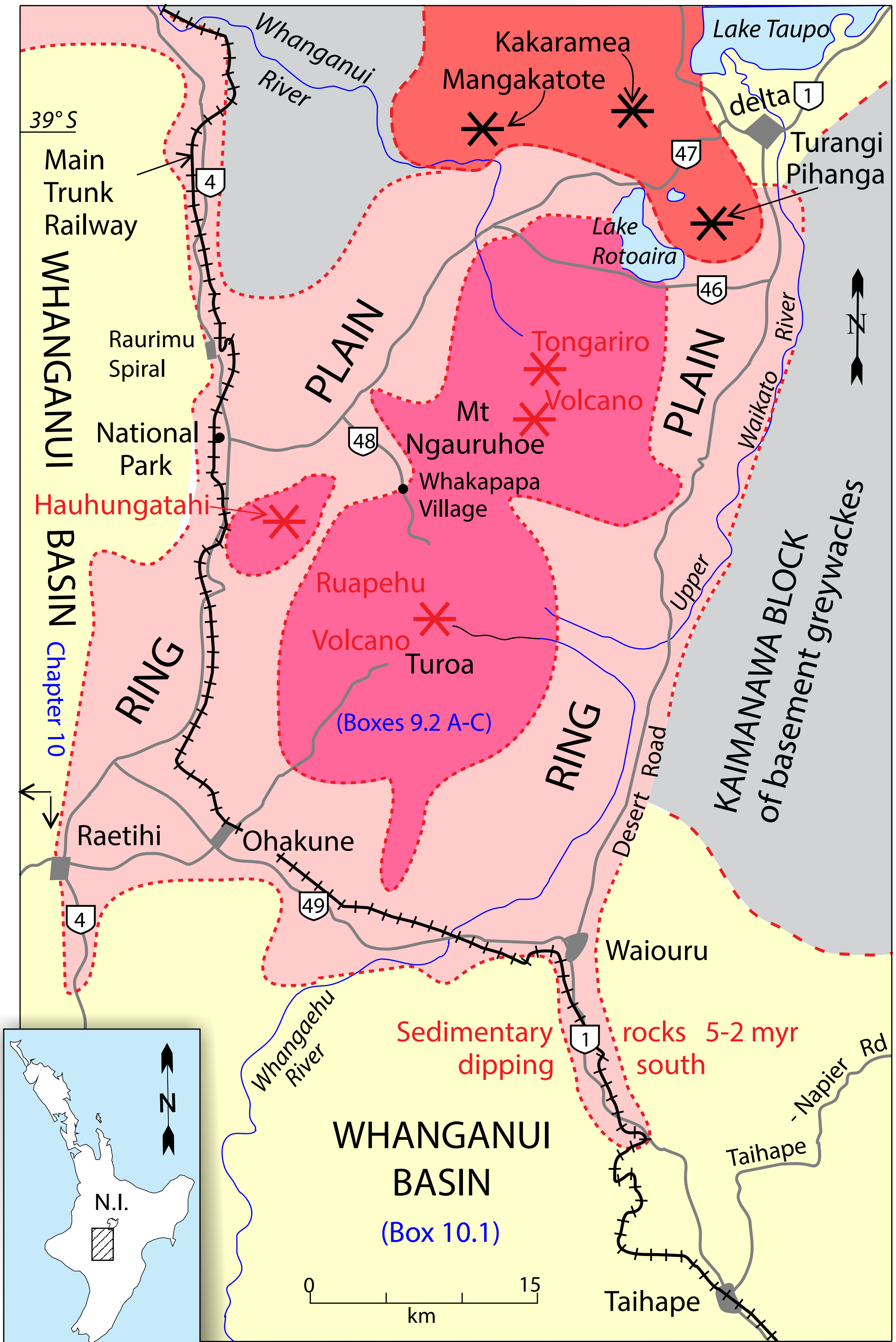


Fig. 9.3. Southern Taupo Volcanic Zone.

The Taupo Volcanic Zone (TVZ) is without doubt the most exciting part of the North Island, geologically speaking, and marks the southern end of the Tonga–Kermadec–New Zealand active volcanic arc. This arc enters the North Island at White Island, the active volcano in the Bay of Plenty, passes through the Rotorua and Taupo volcanic districts, and ends at Mt Ruapehu in the centre of the island (Box 9.1). Just why it should end here is a matter of some interest, as we shall see.

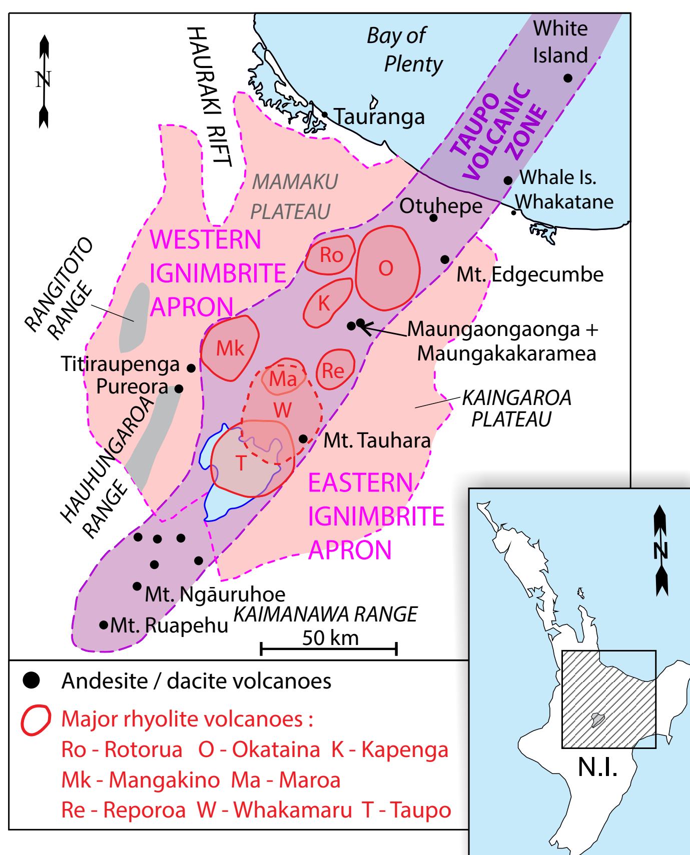
Being a volcanic arc and forming part of the Pacific Ring of Fire, the TVZ is intimately associated with a subduction zone, which is, of course, part of the Pacific–Australian plate boundary that passes through the entire country. As described in Chapter 3, lithosphere of the Pacific Plate passes beneath that of the Australian Plate, the latter carrying continental crust of the North Island. This occurs as far south as Cook Strait, where things change.

A volcanic arc (actually a straight line in the case of the TVZ) is only one of several manifestations of a subduction zone. Others are the deep-sea trench, subduction earthquakes and the so-called forearc region, which lies between the trench and the arc. We address the forearc, which comprises the North Island's axial ranges and the whole East Coast, in Chapter 12. Back-arc phenomena are also associated with subduction zones, such as the back-arc basin that enters the North Island as the Bay of Plenty (Chapter 12), and the 10 back-arc basaltic volcanic fields that were described in chapters 6, 7 and 8. Box 2.3 shows all these phenomena on one map.

Box 9.1. The Taupo Volcanic Zone (TVZ).

The Taupo Volcanic Zone (TVZ)

- is the active volcanic arc of the New Zealand subduction zone
- is the southern end of the Tonga-Kermadec-New Zealand volcanic arc
- is an actively widening rift structure (7 mm per year)
- contains all of New Zealand's active rhyolitic volcanoes
- has two rhyolitic centres currently active - Taupo and Okataina and six dormant or extinct centres
- lava production from these two centres over the past 50,000 years has averaged 1 cubic metre every second
- rhyolitic eruptions typically occur at thousand year intervals
- the zone contains several active andesitic/dacitic volcanoes and a few small basaltic ones, aligned along the eastern side of the zone, marking the 'volcanic front'.



Volcanic Time Perspectives

There is nothing simple about volcanic arcs. When we talk about the TVZ we need to bear in mind two different time perspectives. First, there is the 'instantaneous' TVZ, comprising the handful of volcanoes that are active now. Even 'now' needs to be defined, because volcanoes do not erupt every day; in this sense, 'now' becomes a period of time defined by the typical gap between eruptions. With regards to the two currently active rhyolitic volcanoes in the TVZ, Taupo and Okataina, Taupo hasn't erupted for nearly 2000 years, so 'now' in this case means in the last few thousand years.

Second, there is the long-term TVZ. 'Long term' means the past 2 myr, this being the time that the volcanic arc has existed more or less in its present location. During that time, lots of volcanoes have switched on and off again, and the line of the arc of active volcanoes has shifted, moving eastwards and extending further south. Prior to this, as we saw in Chapter 3, the TVZ has antecedents stretching back 25 myr.

Types of Volcano in the TVZ

The volcanoes in the TVZ are typical of volcanic arcs worldwide. There are two main types: andesitic and rhyolitic. Andesitic volcanoes erupt frequently and violently, and build large cones like Mt Ruapehu (Boxes 9.2 A–C). Rhyolitic volcanoes either build domes, by extruding very sticky lava, or they explode, infrequently but very violently, blasting huge quantities of volcanic ash into the air and leaving large caldera landforms (Boxes 9. 3 A–C). Basaltic volcanism (Box 6.8) is very uncommon in the TVZ.

In most road cuttings in the region, the soil is underlain by a number of rhyolitic (pale-coloured) ash showers, in layers up to 2 m thick. As these showers fall, they drape the landscape, and therefore lie parallel to the overlying land surface. Each ash shower may show the remains of a dark soil layer that developed on it (Box 9.4).

The TVZ is divided into three distinct sectors: the central sector (Rotorua–Taupo), which comprises mainly rhyolitic volcanoes; and the northern and southern sectors, which comprise andesitic volcanoes. In addition, there is a line of widely spaced andesitic to dacitic volcanoes that forms the eastern limit of the TVZ – the 'volcanic front' nearest to the subduction trench. The reasons for the north–south division into sectors are unknown. The basic difference between them is that in the two andesitic sectors there is little interaction between the magmas (which originate at around 100 km depth on the subducting lithospheric slab) and the continental crust (around 25 km thick) that they pass through on their way to the surface. The central rhyolitic volcanoes, on the other hand, erupt material that is basically melted continental crust; in this case, magma rising from 100 km depth is operating through some kind of heat-exchange system at the base of the crust. The plumbing of this heat exchange, and the reason for it being confined to Rotorua–Taupo, are not understood.

A caldera is an extra-large volcanic crater, produced by the collapse of a magma chamber following a large eruption. The biggest calderas are associated with rhyolitic volcanoes. This is the place to introduce the idea of the 'upside-down volcano', which is clearest in the case of the Taupo

Volcano. Eruptions from the northern half of Lake Taupo in the past 50,000 years have been very voluminous, ejecting more than 1000 cu km of pumice and fragmented lava, and have also been very powerful. The material has been blasted far and wide, so that instead of building a familiar cone, Taupo Volcano has excavated a large hole in the ground – an inverted cone. Following the eruptions, collapse of the caldera formed the lake basin itself. If you face Taupo town from the lookout on SH1 on its northern outskirts, you are actually looking down into the inverted cone, with the outer rims far to the right and left.

Calderas can be filled up, partially or completely, by continued activity, particularly the growth of rhyolite domes. Whereas the Taupo inverted cone/caldera has just a few domes, the other active volcano in the central sector, Okataina, to the east of Rotorua, has many more, which makes recognition of this caldera difficult (Fig. 9.1).

The Northern Sector

Fig. 9.a A shows the eastern limit of the TVZ, the 'volcanic front', as being defined by the younger andesitic cones of Mts Edgecumbe and Maungakakaramea. Note that there are older andesitic volcanoes, Manawahe and Puhipuhi, behind (i.e. to the west of) the volcanic front, and that these mark previous positions of the front.

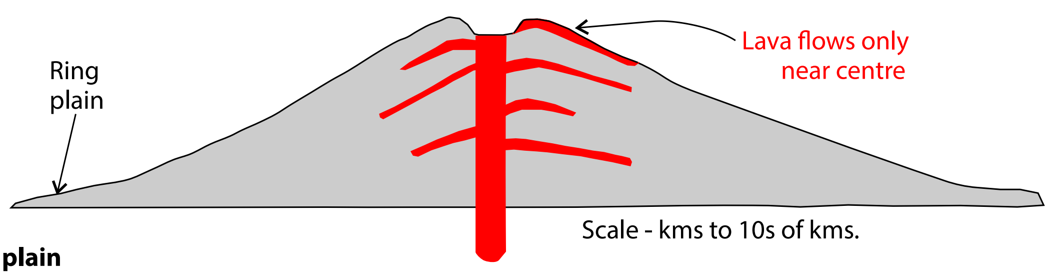
The northern andesitic sector of the TVZ lies to the north of the rhyolitic Okataina Caldera, and includes the extinct volcanoes Manawahe (onshore) and Motuhora or Whale Island (offshore), as well as the active volcanoes Mt Edgecumbe (onshore) and White Island (offshore). However, most of the sector is blanketed by rhyolitic ignimbrite deposits erupted from Okataina Caldera, or by alluvial sediment in the Whakatane Graben.

The Whakatane flats, the delta of the Rangitaiki River, are the alluvial fill of a tectonic graben, or elongate, fault-bounded depression. This marks the extension onshore of the back-arc basin behind the Kermadec sector of the volcanic arc (Fig. 1.3). During the Edgecumbe earthquake of 1987, the Whakatane flats widened by 1 m – part of the 'unzipping' of the North Island and opening of the Bay of Plenty (Chapter 12).

Box 9.2 A. Andesitic volcanoes.

Andesite is lava containing about 60% silica.
Properties:
- melts at about 950°C
- is stiff and does not flow easily, contains much gas
- activity is dominantly explosive
- only 20-25% of lava is extruded as lava flows, the remainder is thrown out as blocks and ash.

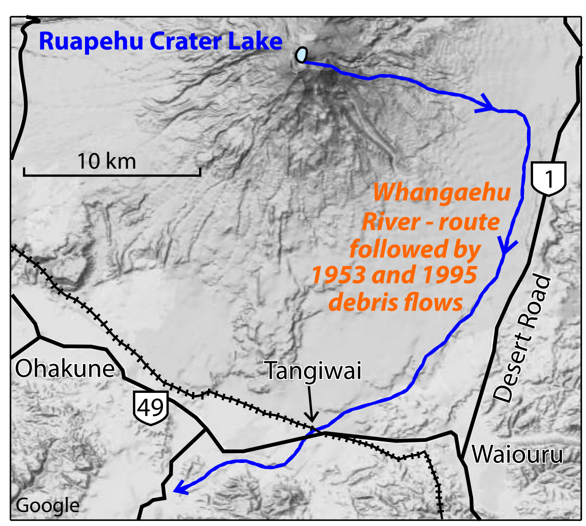
Eruption style
Typically from a single, central vent, forming a large stratocone with a crater. The crater may hold a hot lake, e.g. Mt. Ruapehu. Eruptions are generally explosive, and may send dust and gas aerosols high into the stratosphere (e.g. Mt Pinatubo, Phillippines, 1991). Ash is dispersed many kilometres downwind. Although lava flows are uncommon, they form a skeleton, the ribs, which help to brace the cone.



Ring plain
The most characteristic feature of andesitic stratocones is the ring plain. It exists because the combination of height, steep slopes, lots of fragmental material, lots of water, and explosive eruptions, favours the generation of gravity flows.

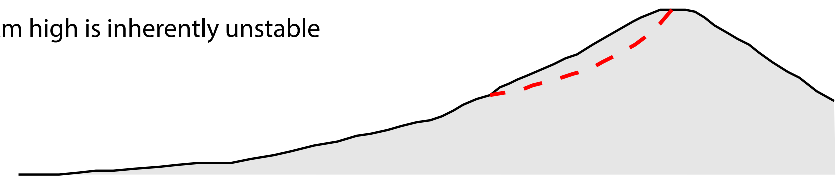
A **gravity flow** is a mobile mixture of rocks, ash and fluid (gas and water) which can be generated in several ways - e.g. sudden ejection of water from a crater lake, sudden melting of ice and snow, heavy rain, collapse of part of the cone (stratocones are inherently unstable), collapse of an eruption column. Andesitic volcanoes can generate hot, dry ignimbrite flows analogous to rhyolitic volcanoes but on a much smaller scale - e.g. Mt. Ngauruhoe, 1974/5).

Technically, most of the gravity flows on stratocones are debris flows, non-turbulent mixtures of water, mud and boulders. They are highly mobile and very destructive, even at tens of kilometres from the vent (e.g. the Tangiwai rail disaster of Christmas Eve, 1953 was caused by a sudden release of water from Ruapehu crater lake, which picked up mud and boulders, transformed into a debris flow, and flowed more than 30 kilometres down the Whangaehu River valley before demolishing the rail bridge at Tangiwai just minutes before the packed overnight train arrived. The debris flow was still flowing, and carried the train engine some distance downstream. More than 150 people were killed). Debris flows are often known by the Indonesian name 'lahar'. They cause most of the deaths and property damage associated with andesitic volcanoes. As Mt Pinatubo showed, after a large eruption destructive lahar flows can continue for at least a decade.

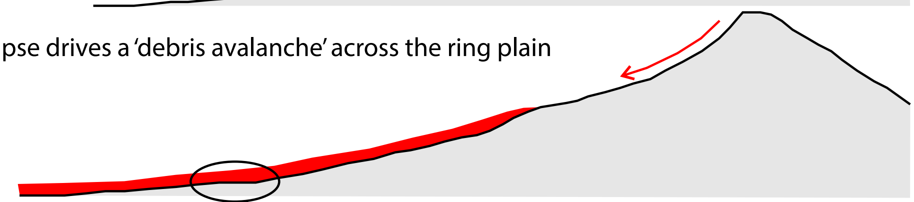


Box 9.2 B. Cone-collapse events.

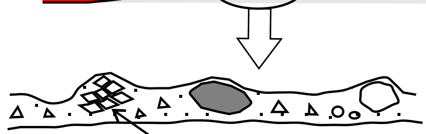
(1) Stratocone 2 to 3 km high is inherently unstable



(2) Sector collapse drives a 'debris avalanche' across the ring plain



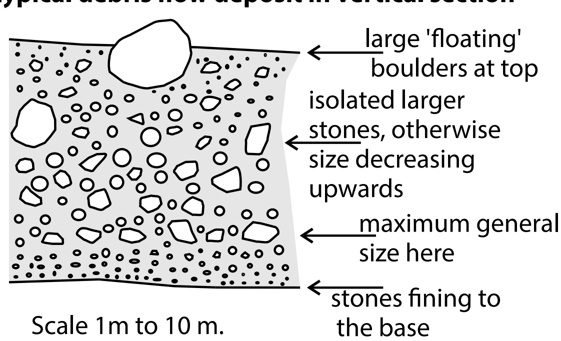
(3)



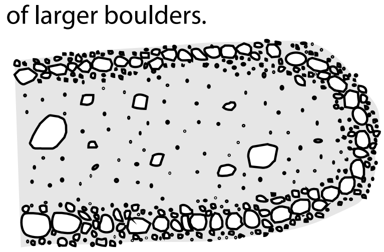
Deposit forms a mound field. Each mound comprises a large piece (10s to 100s of metres) of the cone.

(4) Repeated sector collapses produce the Mt. Taranaki profile, a slender, central, pointed peak.

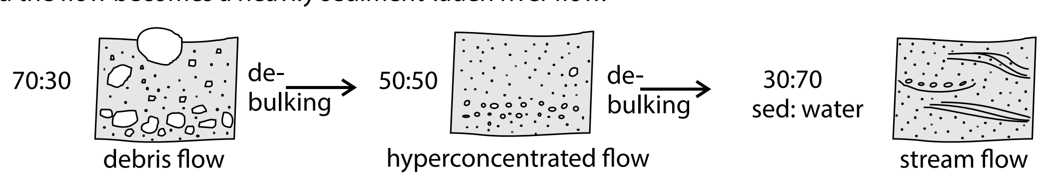
A typical debris flow deposit in vertical section



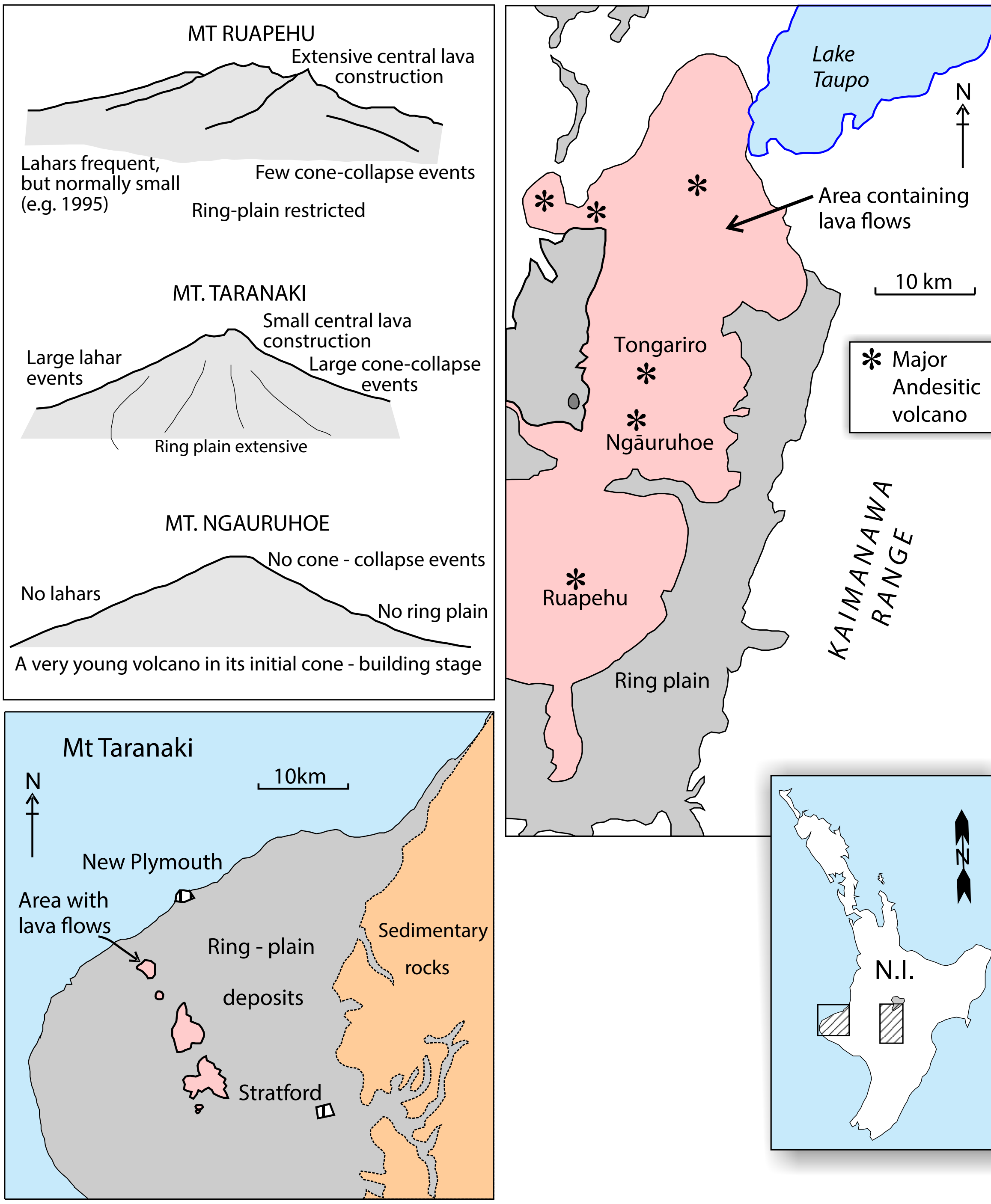
In plan view a debris flow deposit has a lobate shape and a marginal levee of larger boulders.



What happens when a debris flow (lahar) follows a river?
A debris flow commonly begins when a sudden flood of water bulks up by picking up rocks and sediment. A typical debris flow has about 70% sediment + 30% water by volume. If the debris flow then follows a river valley it moves faster than the river water, and so progressively takes in water - it de-bulks. When the proportions are about 50-50 the flow changes character and becomes a 'hyper-concentrated flow'. Large boulders are abandoned, but the deposit of a hyper-concentrated flow will still contain floating small boulders and a general size distribution like a debris flow deposit. Overall, though, it is better sorted, i.e. it has fewer fine and large particles.
If debulking continues, a further change of character takes place at about 70% water, 30% sediment, and the flow becomes a heavily sediment-laden river flow.



Box 9.2 C. Comparison of Mounts Ruapehu, Taranaki and Ngauruhoe.



Box 9.3 A. Rhyolite volcanism and ignimbrites.

Rhyolite is lava containing more than 70% silica.

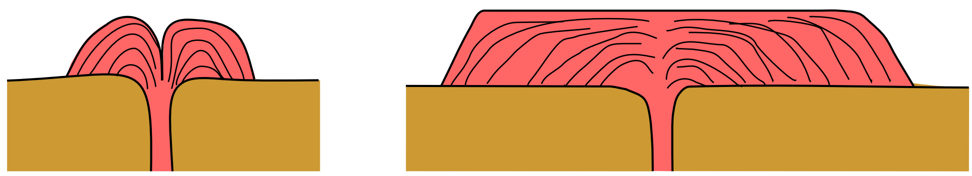
Properties:
- melts at c. 900°C
- is very stiff and contains much dissolved gas.

Products:
- obsidian (dark-coloured natural glass)
- rhyolite (light-coloured fine-grained lava)
- pumice (light-coloured solidified froth).

Eruption style

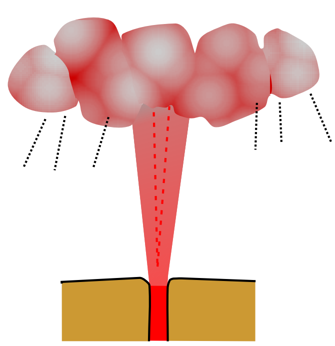
There are two contrasting eruption styles:

(1) Extrusion of very stiff lava makes rhyolite domes. These are typically circular in plan view, flat-topped and steep-sided e.g. Mt Ngongataha, Rotorua. There are hundreds of domes in the Taupo Volcanic Zone. Mt Maunganui is an older dome.

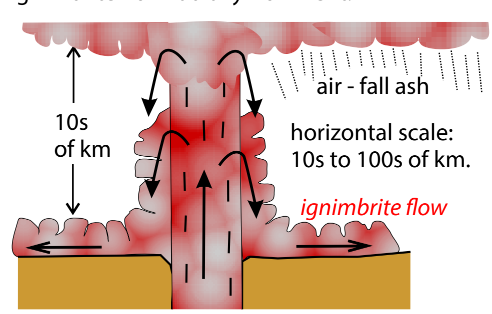


(2) Highly explosive eruptions of gas and pumice, plus steam if water is involved, cause ignimbrite flows and deposits.

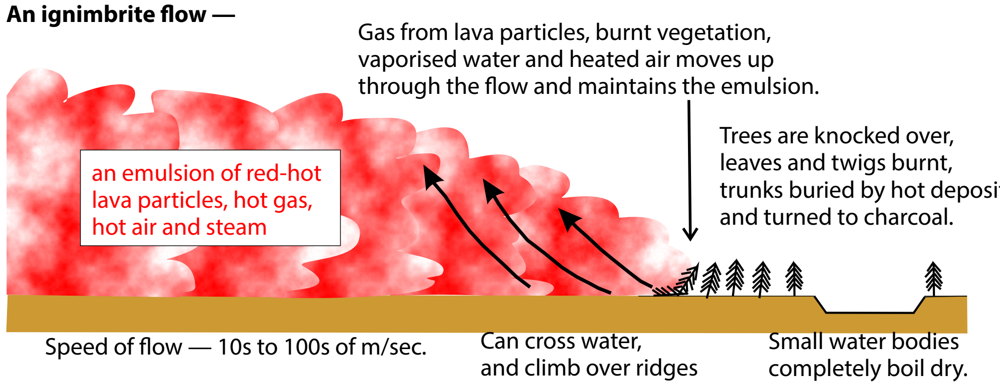
(a) initial activity, small vent, airfall ash



(b) eruption enlarges vent, becomes very powerful. Collapse of eruption column drives ignimbrite flow radially from vent.



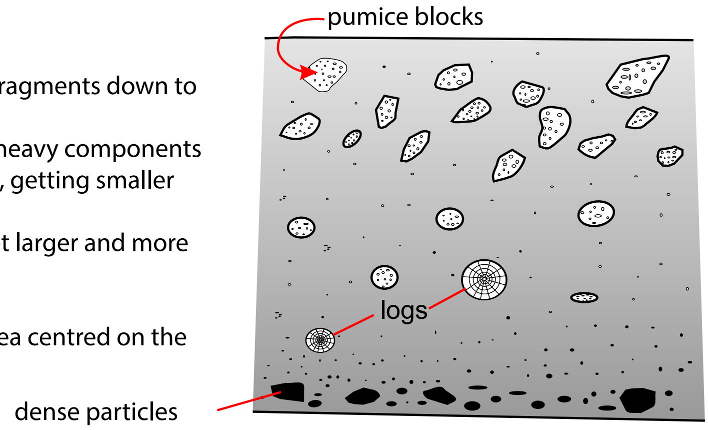
An ignimbrite flow —



Box 9.3 B. Ignimbrite deposits.

An ignimbrite deposit:

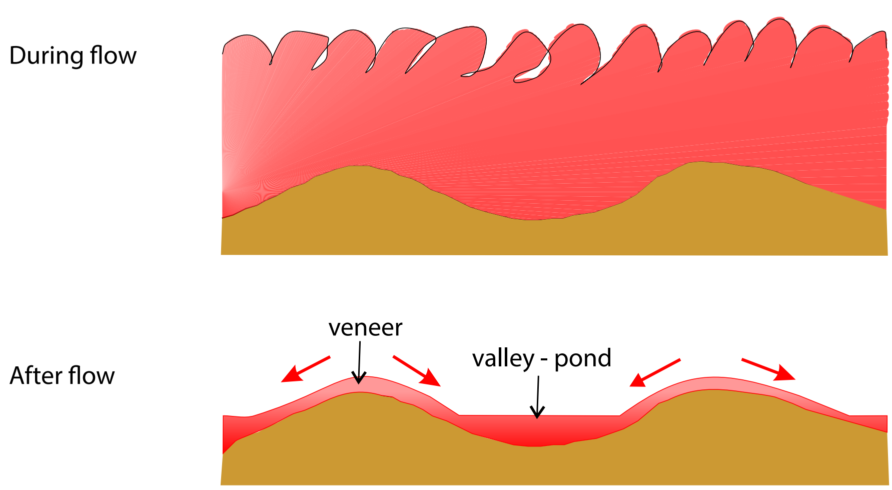
- is typically 2 to 50 m thick
- consists mostly of small pumice fragments down to dust size
- contains a concentrated layer of heavy components (rocks and crystals) near the base, getting smaller upwards
- contains pumice blocks which get larger and more numerous towards the top
- contains carbonised logs
- is distributed in a near-circular area centred on the vent.



After the flow has slowed and deposited its material, it may simply cool, or the material may still be so hot that it welds together into a solid rock. In the latter case, the pumice fragments usually collapse into thin lens-shaped pieces of glass, and the deposit acquires metre-sized cooling joints. Thus an ignimbrite deposit is either welded or unwelded, and the two kinds of end product look quite different.

Landforms produced by ignimbrite deposits -

When the ignimbrite flow slows and eventually stops, the material is still highly gas-charged and quite fluid. While it is deflating it tends to drain from ridges into valleys, building up flat-topped valley-pond deposits, and leaving just a veneer on the ridges.



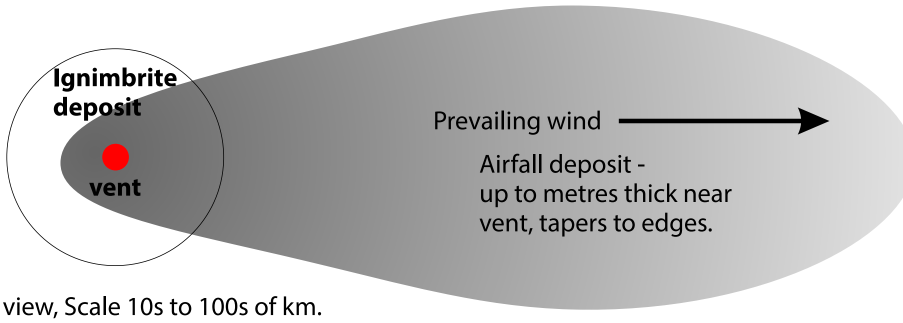
The height of the valley-pond surface in each valley is independent of its neighbours. Note that after the valley-pond deposit has been dissected by stream erosion the remnants resemble river terraces (Box 10.3).

Box 9.3 C. Airfall ash associated with ignimbrite deposits.

Associated with an ignimbrite flow there is typically -

a) an extensive airfall ash distributed from the top of the eruption column by high-level winds, and

b) a fine airfall dust deposit on top of the ignimbrite deposit, blown out by the hot gases in the still-moving flow.



Map view, Scale 10s to 100s of km.

The combined volume of the ignimbrite and airfall deposits may exceed 1,000 cubic kilometres (km^3). Historic eruptions of this type (Valley of Ten Thousand Smokes, Alaska, 1910; Mt. Pelee - St. Pierre, Martinique, 1902) have produced only a few km^3.

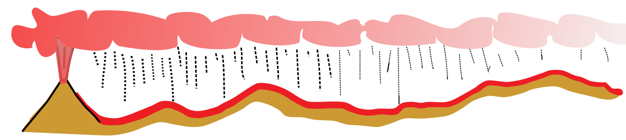
RHYOLITIC VOLCANOES

Rhyolitic domes have long been recognised as volcanoes. However, it has only recently been recognised that the very large, violently explosive, ignimbrite-forming eruptions give rise to a negative volcanic landform, i.e. a large hole. This is because the violence of the eruption carries most of the volcanic material tens of kilometres away from the vent. For example, the northern half of Lake Taupo is the vent area for several major eruptions, while the wider crater rim is located 5 to 15 km from the lake shore. (See Box 9.1 for the location of rhyolitic volcanoes in the Taupo Volcanic Zone.)

FREQUENCY OF ERUPTIONS

Large ignimbrite eruptions (more than 50 cubic kilometres) tend to occur every few thousand years from each volcano. The last one in New Zealand was the Taupo Ignimbrite of approximately 232 AD. They inject so much dust and gaseous aerosol into the stratosphere that they cause global cooling for some years.

Box 9.4. Volcanic ash and wind-blown dust.



Volcanic ash layer:

- technical name tephra
- deposited in hours or weeks,
- distribution is dictated by wind of the day,
- deposit gets thinner and finer grained away from the volcano,
- thickness ranges from less than 1mm to maybe several metres
- drapes the existing landscape, kills vegetation if sufficiently thick
- consists of fragments of rapidly chilled lava (natural glass) and separate crystals of feldspar and other minerals
- may be internally layered (shower bedding)

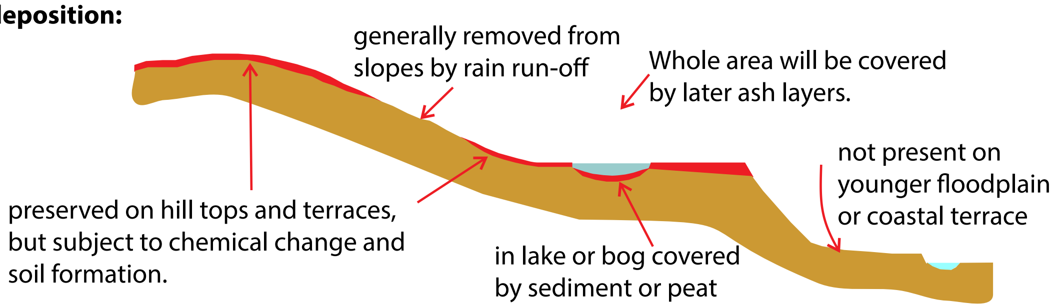
Can be dated:

- by various physical methods (e.g. radiocarbon, fission track), by placement in a succession of ash and dust layers, rarely by fossils.

Can be identified:

- by minerals present, by chemical composition of the volcanic glass.

After deposition:



The Central Rhyolitic Sector

The Rotorua Lakes area comprises the northern half of the rhyolitic sector of the TVZ. It is also the widest part of the TVZ, largely because the Rotorua Caldera lies alongside the Okataina Caldera and makes a bulge to the west. The Okataina Caldera is definitely active – it is one of two presently active rhyolitic volcanoes in the TVZ, the other being Taupo. The Rotorua Caldera may also be a separate, active rhyolitic volcano – although it may equally be regarded as an extension of the Okataina Caldera. Rotorua and Okataina belong to the 'instantaneous' TVZ, and have formed during the last 350,000 years (Fig. 9.1).

The youngest caldera in the area is the Rotorua Caldera (240,000 years), which is the most obvious such landform. Next in terms of age is Okataina Caldera (280,000 years–present), which has been extensively modified by domes and younger deposits, while the oldest, Kapenga Caldera (900,000–240,000 years), has also been heavily overprinted by dome complexes.

The lakes around Rotorua exist for the same reasons that lakes exist in all volcanic terrains – volcanoes blast holes in the ground and block rivers. In this case, the Rotorua Caldera contains a single large lake that reflects its shape, though it has been bigger and deeper in the past. The

Fig. 9.4. Lake Rotorua and the city of Rotorua occupy the collapsed Rotorua caldera formed during the eruption of Mamaku Ignimbrite, 240,000 yrs ago.
Photographer Lloyd Homer, GNS Science.

lake contains just one rhyolite dome, the famous Mokoia Island (there are other domes inside and around the caldera but they are not inside the lake, e.g. Mt Ngongataha). The Okataina Caldera is bigger, but it is less obvious in the landscape because it contains three rhyolite dome complexes that have created lots of hollows. Lakes Okataina and Tarawera were once joined, but were separated by the growth of a textbook rhyolite dome – from the air, it looks just like a large cowpat (Fig. 9.5). There were some adjustments to the lakes following the 1886 Tarawera eruption (see below) – Lake Rotomahana became bigger, and an ash barrier raised Lake Tarawera until the lake washed it away and flooded Kawerau in 1904. Right now, Lakes Tarawera and Rotomahana are at different levels and are separated only by a ridge of loose volcanic ash.

A line of small explosion crater lakes, accessible via walking tracks from the Lake Okataina road, lies between Lakes Rotoiti and Okataina. The 1886 Tarawera eruption rift also contains three small crater lakes (one cold, two hot) that were created in the June 1886 Tarawera eruption; these can be seen in the Waimangu thermal area.

The line of three domes forming Mt Tarawera represents the youngest rhyolitic eruptions in the TVZ; the most recent, the Kaharoa eruption, took place 700 years ago, in the earliest days of human occupation of New Zealand. Dome formation was accompanied by widespread ash showers, which provide a distinctive marker level in archaeological excavations. The three domes were formed by eruptions up the bounding fracture of the Okataina Caldera, i.e. they mark part of the caldera margin.

Fig. 9.5. This rhyolite dome, like a giant cowpat, separates Lakes Okataina (distance) from Lake Tarawera. Photograph courtesy of Google Earth.

The Tarawera eruption

The spectacular Tarawera eruption, on 10 June 1886, is the only one to have occurred in the rhyolitic sector of the TVZ since European settlement of New Zealand. Ironically, the eruption was not of rhyolitic lava, but was basaltic. The minute-by-minute story of the famous eruption, and its human aftermath, is well told in various books. It is also told at various locales around Rotorua, but particularly at the Wairoa buried village site. It was a spectacular eruption by any standard: there was very little warning; it was very short, lasting just a few hours, all during the night; and it was extremely violent (the explosions were heard loudly in Auckland, more than 200 km away). The world-famous pink and white silica terraces of the old Lake Rotomahana were also totally destroyed.

There are three things of particular scientific interest about the eruption. First, the material erupted was basalt instead of the dominant rhyolite – in fact, this is the largest known basaltic eruption from the whole TVZ. Basalt is normally a fluid lava that generates explosive eruptions only when it interacts with water. This eruption did not form a lava flow, but blasted black basaltic lapilli (pebble-sized material) over a wide area. In terms of the heat-exchanger at the base of the crust (see above - "Types of Volcano in the TVZ"), it would appear that the basalt of this eruption (and a few others that have taken place in the TVZ) escaped involvement in it and made its way directly to the surface.

Second, the explosions ejected a great quantity of grey mud from the floor of the old Lake Rotomahana. This mud was blasted over all the hills around the lake, and was the material that buried Wairoa village. Third, the eruption created a brand-new physical feature, the 17 km-long Tarawera Rift. This split the Tarawera dome complex, extended through Lake Rotomahana, and created the Waimangu thermal valley, the only major geothermal area in the world to have come into existence in historic time. All this took place in the early hours of 10 June 1886.

A trip to Rotorua is incomplete without visiting places associated with the Tarawera eruption, including Waimangu, Lakes Rotomahana and Tarawera, the rift on the summit of Mt Tarawera (Fig. 9.6), and Wairoa buried village. There are also many geothermal areas to see in the area, including that at Whakarewarewa – when you are here, check out the cliff on the south side of the site, which is part of the wall of the Rotorua Caldera. In most cases the source of the heat beneath the geysers, hot springs and boiling mud around Rotorua – as in other areas of the TVZ (Box 9.5) – is residual hot rock left over from previous volcanic activity. Rock is a very poor conductor of heat, and as such can stay hot for thousands of years.

Profile of an Ignimbrite Deposit

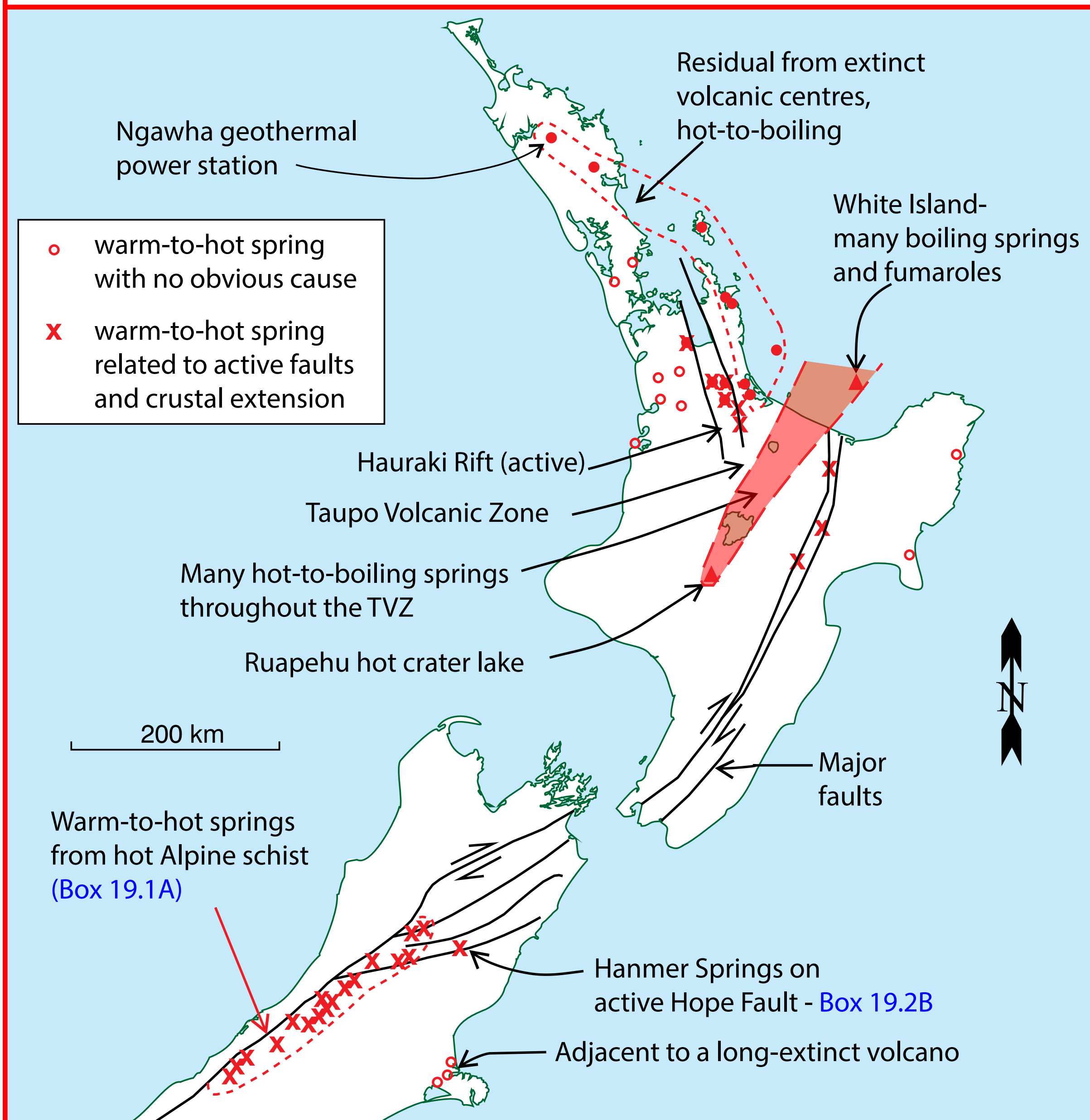
One of the most interesting developments in vulcanology in the latter part of the twentieth century was the surge in understanding of the eruption phenomenon called ignimbrite, which is a type of pyroclastic flow. Box 9.3 A explains the mechanism of a continuously fed, high eruption column of gas and frothed-up molten lava, which collapses, also on a continuous basis, on all sides. The collapse drives a ground-hugging gravity flow of hot gas and glowing lava fragments, radially outwards, for as long as the eruption continues. The flow also contains pieces of heavier rock torn from the sides of the eruption crater.

Fig. 9.6. The 17-km-long Tarawera Rift was created during the 1886 Tarawera eruption. Photographer Lloyd Homer, GNS Science.

There have been no large ignimbrite eruptions anywhere in historic time that are well documented. There was probably one of modest volume during the famous Krakatoa eruption of 1883 (Indonesia), and the major eruption of Tambora in Sumatra in 1815 is thought to have produced more than 100 cu km of material, some of which constituted ignimbrite flows and deposits. The best-known historic example, however, occurred during the eruption of Mt Pelée on the island of Martinique in the Caribbean in 1902. This was a small flow, only a fraction of a cubic kilometre in volume, which escaped from beneath a large lava spine that was being pushed up from the crater, but it was directed down a valley and overran the town of St-Pierre. About 30,000 people were killed, and when the flow continued over the harbour, ships were burned and overturned, and the water surface boiled. The flow was described at the time by the French term *nuée ardente*, meaning 'glowing avalanche'.

Prehistoric ignimbrite eruptions reconstructed from their deposits have been far larger – quantities of material exceeding 1000 cu km are not unusual. When the ignimbrite gravity flow finally comes to rest, the material may weld together to form a solid deposit, or remain unwelded and unconsolidated, depending on the residual heat. For unexplained reasons, many of the early

Box 9.5. Hot springs of New Zealand.
Hot springs arise when circulating groundwater encounters hot rock. As the map indicates, hot rock can have several causes. Active volcanoes in the Taupo Volcanic Zone may introduce small amounts of new water from the earth's interior, but most spring water is rain water.



ignimbrite deposits from the TVZ are welded, whereas none of the younger ones are. Either way, ignimbrite eruptions are intensely destructive, and can cover huge areas.

If you travel along SH33 from Lake Rotoiti to the Bay of Plenty coast, you traverse a 30 km longitudinal profile through the Rotoiti Ignimbrite deposit of 45,000 years ago. The eruption was centred in Lake Rotoiti (part of the Okataina Caldera), and produced a modest 50 cu km of unwelded material. Exposures start at the highest point on the road, near to the lake, but they are ephemeral by their nature, so it's a matter of luck what you see on a particular day. All the exposures consist of a matrix of fine white pumice (too fine to distinguish grains with the naked eye), containing larger blocks of both dense rock and lightweight pumice.

Near to the lake source, heavy, generally dark-coloured stones are prominent, because these dropped out of the gas-supported suspension early in its travels. Further away, heavy stones decline in size and number while white pumice blocks become more prominent. At intermediate points along the road, no internal divisions are visible in the white deposit. However, at the coast, 30 km from source, the roadside cliffs on SH2 show the deposit divided into several flow units in which larger pumice blocks are concentrated at the top. This reflects the fact that gravity flows in general become internally organised as they flow, in this case into discrete pulses. Here, internal shear in the pulses pushed the larger pumice blocks to the rear and top of the flows. The slope on SH33 between Lake Rotoiti and the coast is the depositional slope on which the ignimbrite travelled.

One of the unanswered questions about the Rotoiti Ignimbrite flow is what happened when it encountered the sea. At the time, 45,000 years ago, the world was entering its last period of glaciation and the sea-level was lower than it is today, although not at its lowest level of -130 m (Box 10.2 A, B). Unfortunately, no information is available in this case, but the interaction between large, very hot ignimbrite flows and large bodies of water is a topic of great scientific interest. For example, a flow whose bulk density is lower than that of water could skate across the water surface – there is evidence that this has happened in Italy and Japan. In contrast, a flow of density greater than water could plunge beneath the sea, which is thought to have happened in the Caribbean. There is also the potential for large steam explosions if an ignimbrite flow mixes thoroughly with water. This would set up a volcano-like event that is not a volcano!

Taupo Volcano – Eruptions and Post-eruption Floods

The combination of lakes and volcanoes is a fascinating one with respect to science, but a deadly one in relation to hazard. The point to note here is that, while volcanic eruptions themselves are spectacular and hazardous, the aftermath also presents major hazards that can remain for quite some time. The flood at Kawerau in 1904 (related to the 1886 Mt Tarawera eruption) and the 1953 Tangiwai disaster (related to the 1945 Mt Ruapehu eruption) are both examples of this.

When it comes to really big volcanoes and big water bodies, the scale of the water hazard becomes mightily impressive. Lake Taupo contains about 60 cu km of water, making it the most voluminous lake in New Zealand. Its narrow southern half is primarily tectonic in origin, downfaulted between more or less parallel faults that are part of the TVZ rift structure. There are domes in the southern half, so there is certainly a volcanic influence here.

The bulbous northern half of the lake is a series of overlapping volcanic vents. It and the surrounding country comprise the Taupo Volcano. This is actually one of the world's biggest volcanoes, but you would never guess that when looking at it, because as we noted earlier it is an upside-down, or inverted, volcano. It is also the world's most productive volcano, measured by volume of product over time, averaging about 1 cu m/sec of liquid magma over the past 50,000 years.

The combination, therefore, of the world's biggest and most productive volcano, and New Zealand's largest lake, holds some sobering hazard implications. We have good knowledge of these hazards, as there is quite a clear record of the last two big eruptions from Taupo and the floods that followed them: the Oruanui eruption, 25,400 years ago; and the Taupo eruption, 1800 years ago (approximately AD232). The Oruanui eruption was enormous, producing more than 1000 cu km of material. It was not, however, a powerful eruption, so although the airfall deposit covered the whole of New Zealand (it is 15 cm thick in the Chatham Islands), the ignimbrite deposit covers a relatively small area centred on northern Lake Taupo but is very thick – 20 m plus. The ignimbrite is very fine-grained, pinkish in colour and unwelded.

The Taupo eruption

The AD232 Taupo eruption is the best-documented ignimbrite eruption in the world, and in contrast with Oruanui it was modest in volume (about 50 cu km) but enormously powerful. The main, ignimbrite-driving part of the eruption was preceded by a number of throat-clearing events that produced a distinctive sequence of airfall ash layers, shown in Box 9.6 and Fig. 9.7. Some of these layers were clearly influenced by interaction between lava and water, and it is thought that Lake Taupo was effectively emptied during these events.

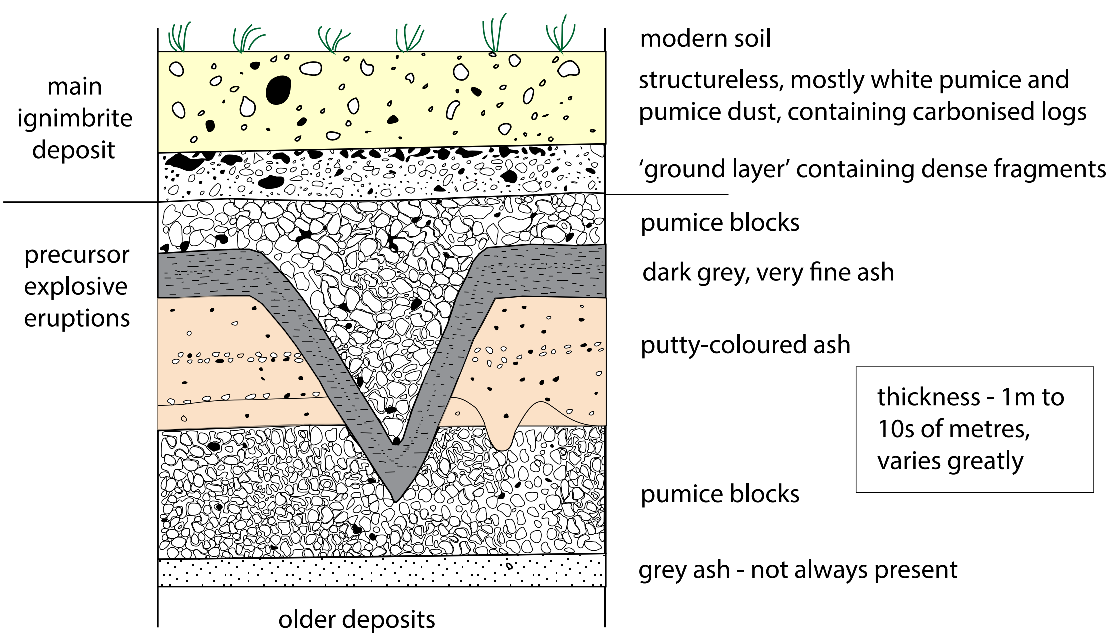
Much of the lake water would have gone down the Waikato River, presumably as a series of flood surges, but there is no geological record of these. There is a simple reason for this – any deposits left by the surges were wiped out by the post-eruption flood, which was much bigger. The importance of those precursor surges is that they highlight the vulnerability of much of the North Island's hydroelectricity generation plant, which is located either in or on the banks of Lake Taupo or the Waikato River, to even a minor volcanic hiccup in Lake Taupo.

The main part of the Taupo eruption took place from a vent in the northeast corner of the present lake, now occupied by Horomatangi Reef. At this stage the lake was empty. The eruption column went up so high (perhaps 50–60 km) that collapse around its margins drove the ignimbrite flow outwards at speeds greater than the speed of sound (1235 kph). The consequences were that the ignimbrite deposit is spread very wide (in a 150 km-diameter circle) and thin (1–2 m in most places), and that it climbed over the top of Mt Tongariro. It may have climbed over Mt Ruapehu as well, but any deposits left up there have been destroyed by that volcano's many small eruptions since AD232. Large areas of forest were knocked over and carried away by the fast-moving ignimbrite flow. Enclosed by the hot ignimbrite deposit, and excluded from oxygen, the wood was baked to black charcoal. The ignimbrite deposit can be recognised by its Persil-whiteness, and the fact that it forms the layer directly below the present-day soil (unless there are some lake-beach deposits on top of it).

The Taupo Ignimbrite deposit presents two quite distinct forms: the hill veneer, and the valley pond. Wherever the fast-moving ignimbrite flow climbed a hill, the flow was compressed. This caused a strong shearing effect in the pumice near the bottom of the flow, and in every instance led to the deposition of a thin (1–2 m), fine-grained 'veneer deposit'. As the flow crested the hill and flowed over the valley below, the shearing pressure disappeared and the flow expanded. Consequently, the ignimbrite deposit found in valleys is typically several metres thick, contains large pumice blocks and has no internal structure. In part, this flat-topped valley pond deposit, as its name implies, is made of material that flowed off ridges and into valleys after the flow had stopped moving but while the deposit was still inflated by gas and acting like a liquid.

Box 9.6. The deposits of the Taupo eruption of c. 232 AD – an idealised sketch (photographed in Fig. 9.7).

There were at least five 'throat-clearing' explosive eruptions, over a period of days or weeks, followed by a particularly powerful sustained eruption column which reached a height of 50 km, and a volume of 30 cubic kilometres. The collapse of the column drove the ignimbrite flow radially outwards for 75 km at speeds of up to 300 metres/second (more than 1000 km/hr, faster than the speed of sound). A good place to see the complete sequence is in cuttings on SH5 between Taupo and Tarawera Springs.



Note:

1. the three ash layers passed through a wet cloud between eruption and falling to the ground. Note the evidence of small v-shaped channel erosion by surface water.
2. the two layers of pumice blocks travelled from the vent in a single ballistic arch, little affected by wind or clouds.
3. the explosive eruptions were influenced by a southwesterly wind, so are thicker to the northest.
4. see Boxes 9.3 A-C for the way in which an ignimbrite flow works.

Fig. 9.7. This road cutting on the Napier-Taupo road shows deposits erupted in sequence from Lake Taupo during its last eruption in 232 AD. Photographer Bruce Hayward.

The post-eruption flood

The Taupo eruption raised the lake outlet to the Waikato River by depositing a 30–40 m thickness of ignimbrite. The lake level rose rapidly and reached a new, higher level – at current rates of inflow, it would have taken about 15 years to fill the empty lake to its new level, or longer if significant amounts of water were absorbed by the pumice-filled caldera and vent. Nearshore and beach sediments deposited during this rapid rise can be seen at various places around the lake. The shoreline terrace that was built or eroded, depending on the slope, at the highest lake level, has been warped by earth movements accompanying earthquake swarms that occur from time to time. Consequently, its height ranges now from 28 m to 43 m above the current lake level, although it is estimated from the least deformed areas that the actual maximum level was 34 m above present (Fig. 9.8).

It is not known how long the lake stood at its maximum height, but it was long enough for a clear beach terrace to form right around it. Eventually, the overflow began to strip out the barrier of unwelded, loose ignimbrite, and a catastrophic outburst flood down the Waikato River valley began.

There have been many similar outburst floods in historic time, both from natural and man-made lakes, and they normally last a few hours at most. However, the Taupo flood was different in two ways. First, it involved a huge volume of water – about 20 cu km excess on top of the present volume of about 60 cu km, placing it in the top ranking of outburst floods in terms of volume of

Fig. 9.8. This smooth sloping ground rises gradually from the edge of Lake Taupo to a 3 m-high scarp 34 m above present lake level. The scarp marks the maximum height that the lake rose to after the 232 AD eruption and the slope was the former lake floor. Whakaipo Bay, 10 km west of Taupo City. Photographer Bruce Hayward.

water released. Larger-volume floods are recorded from the glacial lakes of Bonneville and Missoula in North America, which each released more than 1000 cu km of water, and did so many times.

Second, the outlet channel, which is downstream from where SH1/5 crosses the Waikato River on its way into Taupo, is carved into hard rock and is quite small in cross section. Thus the lake took a long time to empty to its present level, and the flood lasted for more than 40 days. It reached a peak discharge rate of around 20,000 cu m/sec, which is well down in the world outburst flood ranking. However, it was more than enough to cause dramatic changes along the Waikato River, which now has an average flow rate out of Lake Taupo of 130 cu m/sec.

Effects downstream of the lake

The first part of the Waikato River valley is narrow and twisty, and has steps at the Huka Falls and at the old Aratiatia Cascade (a rhyolite dome, now the site of a hydroelectric dam). It then opens out into the Reporoa Basin (through which SH5 passes). The flood carried a large quantity of sediment through the narrow valley, and, starting below the Aratiatia Dam, deposited it as a long, narrow delta/fan in the southern part of the Reporoa Basin, which functioned as a holding lake about 10 m deep. The flood eroded boulders of rhyolite from the Aratiatia Cascade rhyolite dome and deposited them downstream – they diminish in size from a maximum of 4 m just below Aratiatia to less than 1 m about 6 km away. Remember that the maximum boulder size is a function of two things: the power of the flood, and the biggest boulder available. These boulders can be seen on the river flats just below the Aratiatia Dam.

Downstream of the Reporoa Basin, the river passes through another long section of gorges, including Atiamuri, before opening out into the Hamilton Basin at Cambridge (Box 9.7). The outburst flood would have contributed some more sediment to the filling of the Hamilton Basin, although most of that filling dates from the earlier flood that occurred after the Oruanui eruption 25,400 years ago (see below). Some deposits of the post-Taupo eruption flood can be seen in cuttings on SH30, 5.4 km downstream from SH1 at the Waikato River crossing.

The Oruanui eruption

If you thought the post-Taupo eruption outburst flood was impressive, wait for this one. As mentioned above, the Oruanui eruption of the Taupo Volcano 25,400 years ago was enormous, producing more than 1000 cu km of pumiceous product, frothed up from about 530 cu km of molten lava. The erupted material was divided roughly equally between airfall, which went mostly to the southeast beyond the Chatham Islands; the ignimbrite collapse deposit, which covers an area measuring 160 km from northeast to southwest and 120 km from northwest to southeast, and reaches thicknesses of more than 200 m on the east side of Lake Taupo; and material that dropped back into the caldera (the northern half of Lake Taupo). The erupted material buried much of the existing topography, and in the cold, dry, glacial climate of the time created an actual sandy desert. The ignimbrite deposit (fine-grained, lacking internal structures and pinkish in colour) is overlain in many places by cross-bedded sand-dune deposits.

Lake Taupo as we know it now did not exist prior to the Oruanui eruption. Instead, a much bigger lake, Lake Huka, stretched about 70 km from the northern part of present Lake Taupo along the tectonic grain in a northeasterly direction through the Reporoa Basin. It was a long-lived lake whose sediment deposits are widely preserved both above and below ground – you can see them in the cliffs and cuttings around Huka Falls. The Waikato River flowed out of the northern part of the lake, along pretty much its present course.

The Oruanui eruption destroyed Lake Huka, and created a large caldera that now comprises the northern half of Lake Taupo. The southern half of the lake formed at about the same time through tectonic subsidence. The elevation of the new lake was 500 m above sea-level (compared with 367 m above sea-level today), and it would have taken between 100 and 200 years to fill up. The new lake area was about 700 sq km (compared with the present lake's 620 sq km) and its volume was about 175 cu km (the present lake measures 60 cu km). Initial overflow was from the northwest corner, down the Mangakino Stream to the Waikato River at Mangakino, cutting out the big loop followed by the present river. That overflow slowly wore down the sill to 480 m above sea-level, at which point the lake remained stable for a long time – hundreds or perhaps even thousand of years.

Eventually, erosion of the dam in the northeast corner, around the present township of Taupo, led to a catastrophic overflow, sometime before 22,500 years ago. Around 60 cu km of water – three times more than the post-Taupo eruption flood – passed down the Waikato River over an even shorter time period – the lake level fell 75–80 m in one go. Among the effects of the flood is a spectacular boulder deposit at Smythe's Quarry, Atiamuri, 80 km downstream from Taupo. Further slow erosion of the sill eventually lowered the lake to its present level.

Another difference between the two breakout floods was the quantity of sediment available. The post-Taupo eruption flood actually ran out of sediment to transport, but after the Oruanui eruption there was a vast quantity of loose volcanic sediment, the result being that the Waikato River was overwhelmed by sediment for thousands of years. In its present sediment-starved state, the Waikato runs in a single, well-entrenched channel, but for several thousand years following the Oruanui eruption the valley was filled by sediment and the river was wide, shallow and braided. Because of the geography around Hinuera, the braided river could flow to the left along its current course at the right-angle bend in Lake Karapiro, passing through the Hamilton Basin and entering the Tasman Sea at Port Waikato, or flow straight ahead through the Hinuera gap into the Hauraki Rift Valley, and thence into the Hauraki Gulf and the Pacific Ocean (Box 9.7). At this time the world sea-level was near its glacial minimum of –100 m or more, so that the Hauraki Gulf was dry land and the coastline was located seawards of Great Barrier Island. The river therefore could construct a very large braid-plain of volcanic sand. This can now be seen as the Hinuera Formation underlying the Hauraki Plains, Firth of Thames and Hauraki Gulf.

An interesting side-effect of the construction of the large braid-plain is that when sea-level recovered from its glacial minimum, gradually drowning the continental shelf and Hauraki Gulf by an average of 1–2 m/century, coastal construction reworked the Oruanui eruption sand into beaches. Some of these beaches were left behind as seafloor sand ridges, and these have been mined for sand, e.g. to replenish beaches in Auckland City. The current beaches around the Hauraki Gulf are also made largely of sand reworked from the Hinuera Formation, a fact that is important because, by virtue of the fact that sea-level has been constant for the last 7,000 years, the beaches are diminishing.

Going back to that right-angle bend in the Waikato River at Lake Karapiro, this point at which the river either flowed to the left (as now) or straight on to the Hauraki Gulf has been dubbed the Hinuera Disjunction. We don't have detailed records of when or how many times the river switched back and forth between the two courses, or indeed whether it split and flowed both ways at the same time. The river certainly delivered a large amount of Hinuera Formation sand into the Hamilton Basin (now used in industry) and built a large, low-angle alluvial fan through the basin.

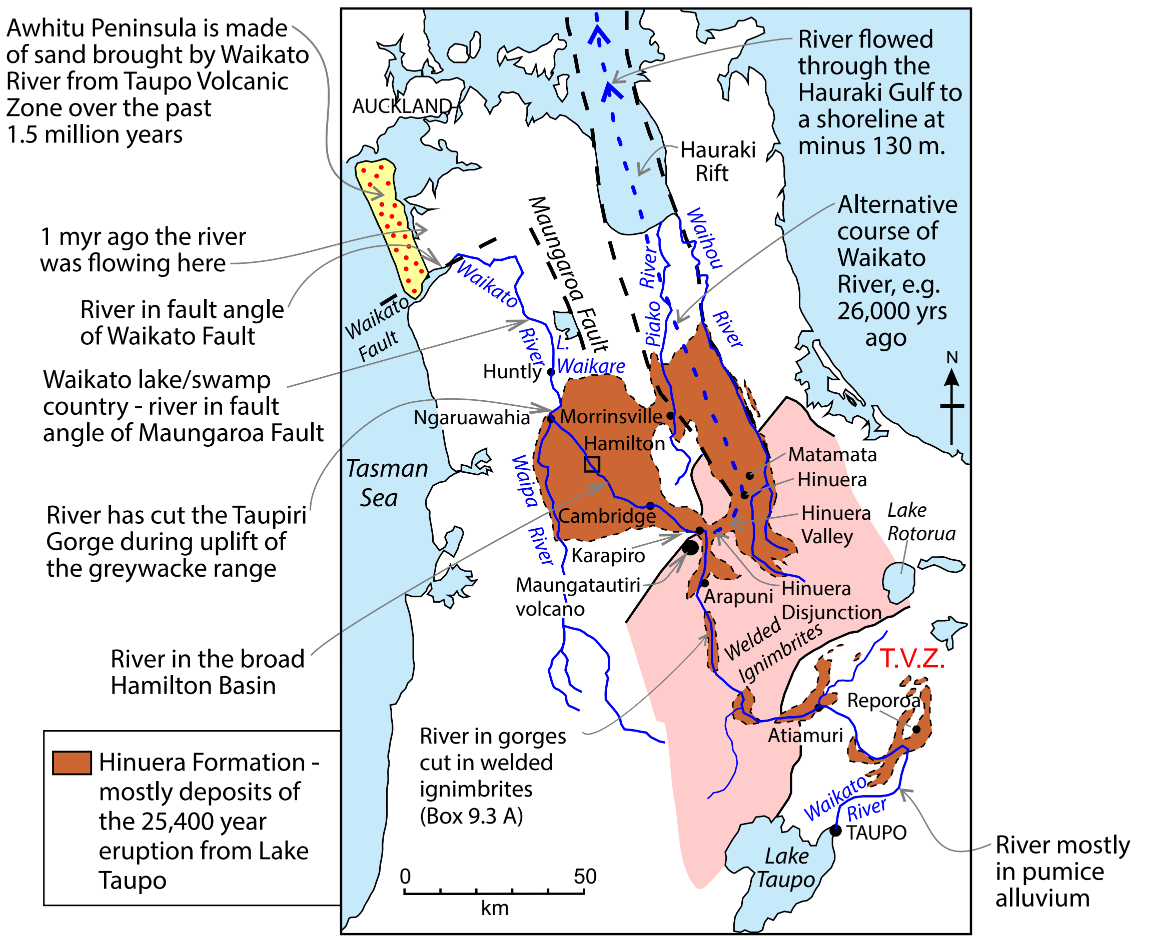
Eventually, after several thousand years, the oversupply of sand dwindled, the river changed to a single-channel flow that became somewhat entrenched, and the Pacific Ocean course was abandoned in favour of the present one.

The Frontier Volcanoes

Brooding over Taupo to the east is the cone of Mt Tauhara, one of the 'frontier' volcanoes that mark the volcanic front of the TVZ at its easternmost edge nearest the subduction trench (Fig. 9.2 A). The cone is a complex affair of domes, vents and pyroclastic flows that erupted around 120,000 years ago, and is dacitic in composition (has a silica content midway between andesite and rhyolite. Earlier fronts lie to the west, and date back 2 myr (Figs 9.2 A, B; 9.3) .

So far we've concentrated on the currently active, eastern sector of the TVZ. The earlier part of the 2 myr TVZ is located west of the active part, and several extinct calderas from this period have been recognised, e.g. Whakamaru (Fig. 9.2 A), although they are difficult to see on the ground owing

Box 9.7. Waikato River.



The Waikato River, New Zealand's longest, has a complex course. In places it is controlled by tectonic depressions, and in others it cuts gorges into hard rock. Most of the hydroelectricity dams are in the gorges between Atiamuri and Karapiro. Following the major eruption from Lake Taupo 25,400 years ago, the river was choked by large volumes of sediment. It raised its bed, became braided, and discharged either simultaneously or alternately through two courses - the existing course from Karapiro, and the Hauraki Rift via the Hinuera Valley. The Hinuera Disjunction, at the junction of SH1 and SH29, is where the river switched courses. Sea level was more than 100 metres below present level, and the river spread volcanic sand right across the Hauraki Gulf. That sand is now an important component of all the beaches around the Gulf. The river may have behaved similarly following earlier big eruptions, but no evidence has been preserved. The early history of the river is not known. However the existence of four major coastal sand barriers/tombolos (Box 6.10) north of the river mouth, all of them made of sand derived largely from the Taupo Volcanic Zone, and all of them probably going back at least 1 million years, suggests a long life for the river and a long term discharge point at the present mouth. There are old river deposits, about 1 myr old at Waiuku.

to modification by later rhyolite domes (Fig. 9.5) and erosion or sedimentation. The most notable difference between the older and newer TVZ is the prevalence of welded ignimbrite deposits in the older, and their absence in the younger. There is no known explanation for this difference, but the effect in landscape terms is substantial. Welded ignimbrites form elevated plateaux with bluffed edges showing vertical columnar jointing (Box 8.2). The middle reaches of the Waikato River comprise a series of gorges carved into a sequence of welded ignimbrite flows, many of them now occupied by hydroelectricity lakes.

Fig. 9.9. 120,000 yr old Mt Tauhara dacite dome, near Taupo, is one of the present-day frontier volcanoes on the easternmost edge of the Taupo Volcanic Zone. Photographer Bruce Hayward.

The Southern Sector

The cluster of big andesitic volcanic cones in the centre of the North Island comprises New Zealand's oldest national park, the Tongariro National Park, generously gifted to the nation by Te Heuheu Tukino IV in 1887. It is one of the world's iconic localities, where numerous features of geologic and scientific interest are superbly displayed in a setting of unsurpassed scenic grandeur.

Of the three main cones, Mt Tongariro is the oldest, and is somewhat frayed around the edges. Mt Ruapehu, the biggest, is middle-aged (around 300,000 years) and has a solid framework of lava flows. Mt Ngauruhoe is a young satellite volcano of Mt Tongariro, only a few thousand years old. All three are active (Fig. 9.3).

The basic features of andesitic volcanoes are set out in Boxes 9.2 A–C. The character that has the greatest influence on the style of activity and the architecture of the cones is the medium level of viscosity of the lava, combined with its high gas content. Lava flows do occur, and can be seen in cross section in cliffs around Mt Tongariro or as historic or immediately prehistoric surface features on the lower slopes of Mt Ngauruhoe. However, only 20–30% of the output is as lava flows. Most of the lava is erupted explosively and violently, as expanding gas bursts out of viscous blobs of lava.

Fig. 9.10. The cluster of three large andesite volcanoes in the centre of the North Island – Ruapehu (snow-covered), Tongariro (right) and Ngauruhoe (between them, partly obscured by the ash plume from Tongariro's Te Mari vent), viewed across Lake Taupo from Taupo City, 2012. Photographer Bruce Hayward.

Some of the explosive product travels great distances as fine ash clouds, while some remains locally as accumulations of blocks. The 1974–75 eruptions of Mt Ngauruhoe provided several examples of the partial collapse of a vertical eruption column, driving ground-hugging gravity flows of lava blocks down the flanks of the cone. These particular block and ash flows didn't travel beyond the steep slopes of the cone, but they are first cousin to the much bigger and further-travelled ignimbrite flows that were discussed earlier.

Most of the features described below can be seen in the Mangatepopo Valley between Mts Tongariro and Ngauruhoe.

The ring plain

An important feature of andesitic volcanoes is the smooth ring plain that surrounds the central peak of lava flows and fragmental deposits. Continual growth around the central peak leads to frequent oversteepening and gravitational instability. The resulting periodic collapses drive another variety of ground-hugging sediment gravity flow, the debris flow or lahar, to use the Indonesian word (Boxes 9.2 B, C). Lahars carry boulders in a matrix of mud, and can be very destructive.

Most of the damage and human casualties from andesitic eruptions are caused by lahars. The scale of lahars varies enormously, from flows of a few metres wide created when water is tossed out of the crater lake of Mt Ruapehu by small eruptions from time to time; to valley-confined lahars that follow a few years after eruptions, as when the crater lake of Ruapehu overtops and destroys the ash dam built by the eruption (e.g. as happened in 1953 at Tangiwai, following the 1945 eruption);

and finally to the sector collapses that see up to one-fifth or so of the cone collapse. A big collapse drives what is known as a debris avalanche, which carries large chunks of the mountain. The world's best-known debris avalanche occurred in May 1980 at Mt St Helens in Washington State, USA. In this case the sector collapse was the first event in a long and complex eruption sequence, but they do not have to accompany eruptions and can occur at any time.

The ring plain surrounding a central andesitic cone is built by lahars. In the case of Mts Ruapehu and Tongariro, and the smaller Pihanga to the north, the ring plains have coalesced to form an apron that slopes down to SH1 all the way from Rangipo to Waiouru. Part of this ring plain is known as the Rangipo Desert because of its sandy, forest-free nature.

Motorists driving across the ring plain along that 60 km stretch of SH1, the Desert Road, can see plenty of interest in the road cuttings. Look for coarse, angular breccias of dark-coloured andesite (lahar deposits), soil-parallel layers of fine brown ash (airfall ash from the volcanoes), lenses of white pumice (Taupo eruption pumice – sometimes with andesitic ash showers above it), and lenses of better sorted, more rounded gravel (stream deposits). The Desert Road lies downwind of the andesitic cones, and so has a good ash record of their eruptions (Fig. 9.11).

Fig. 9.11. This cutting on the Desert Road shows a sequence of andesite ash layers erupted from the nearby Ruapehu, Tongariro and Ngaruahoe volcanoes. Photographer Bruce Hayward.

The End of the Ring

Finally in this chapter, we ask the question why does the Pacific Ring of Fire – which is continuous for most of the distance anticlockwise around the Pacific Ocean from southern South America – finally end at Mt Ruapehu? If we continue the line of the active volcanoes southwards from Ruapehu, we find instead the deep sedimentary Whanganui Basin (Chapter 10). What's more, this is not a recent relationship – it has persisted for more than 20 myr. During this time, the arc–basin transition has moved southwards by about 300 km, and that trend is still going on. The gradual southwards march of the centre of subsidence in the Whanganui Basin is causing the north end of the South Island to subside, forming the Marlborough Sounds.

The answer to the question seems to be related to the length of time taken for the subducting slab to reach the critical magma-generating depth of 100 km. Bearing in mind that the movement of the Pacific Plate is an anticlockwise rotation about a point or pole at 60°S, 180°E (Box 2.1), it can be seen in Box 2.3 that the Pacific Plate approaches the northeast–southwest-oriented subduction zone at an angle that becomes increasingly oblique towards the south. In the South Island, the movement is nearly parallel to the boundary, resulting in the sideways-moving Alpine Fault.

At latitude 35°S, due east of North Cape, the Pacific Plate approaches the subduction trench orthogonally, or head-on at 90°, and is travelling at 60 mm/year. The active volcanic arc lies 115 km from the trench, measured along the line being taken by the downgoing plate, and 100 km above the zone of magma generation. This translates to an angle of subduction, measured from the horizontal, of 41°. At a plate movement rate of 60 mm/year, a given point on the subducting slab will travel 150 km and will take about 2.5 myr to reach 100 km depth.

In contrast, the distance from Mt Ruapehu, the southernmost volcano, to the trench, along the approach line of the Pacific Plate, is 300 km. With the 100 km slab depth lying beneath the volcano, that translates to a subduction angle of only 18°, a travel distance from the trench to the 100 km slab depth of 316 km, and, at a rate of 50 mm/year, a time of 6.3 myr for a given point on the slab to reach magma-generation depth. In other words, the slab beneath Ruapehu has had 2.5 times as long to warm up as the northern example.

One might have thought that for a given slab the subduction angle would stay the same, and that with increasingly oblique subduction the line of volcanoes would curl towards the trench. However, that does not happen. Remember that the other critical factor in the generation of magma is pressure, which is determined solely by depth.

Bearing in mind that those numbers are the simplest possible interpretation of the situation, we can propose that the explanation of the abrupt ending of the volcanic arc at Ruapehu volcano must lie in the warming of the slab reaching some critical level. We can be sure that pressure isn't the culprit, because the slab continues to reach 100 km depth for a further 300 km to the southwest.

We should note in passing that the subducting slab does not end at the depth where the earthquakes stop. It goes on downwards, steepening as it penetrates the earth's interior. The deepest earthquakes record the depth at which the slab has warmed sufficiently to become plastic, so that earthquake stresses no longer build up. Boxes 13.2 A–C shows the New Zealand subduction slab.

Chapter 10

Whanganui Basin

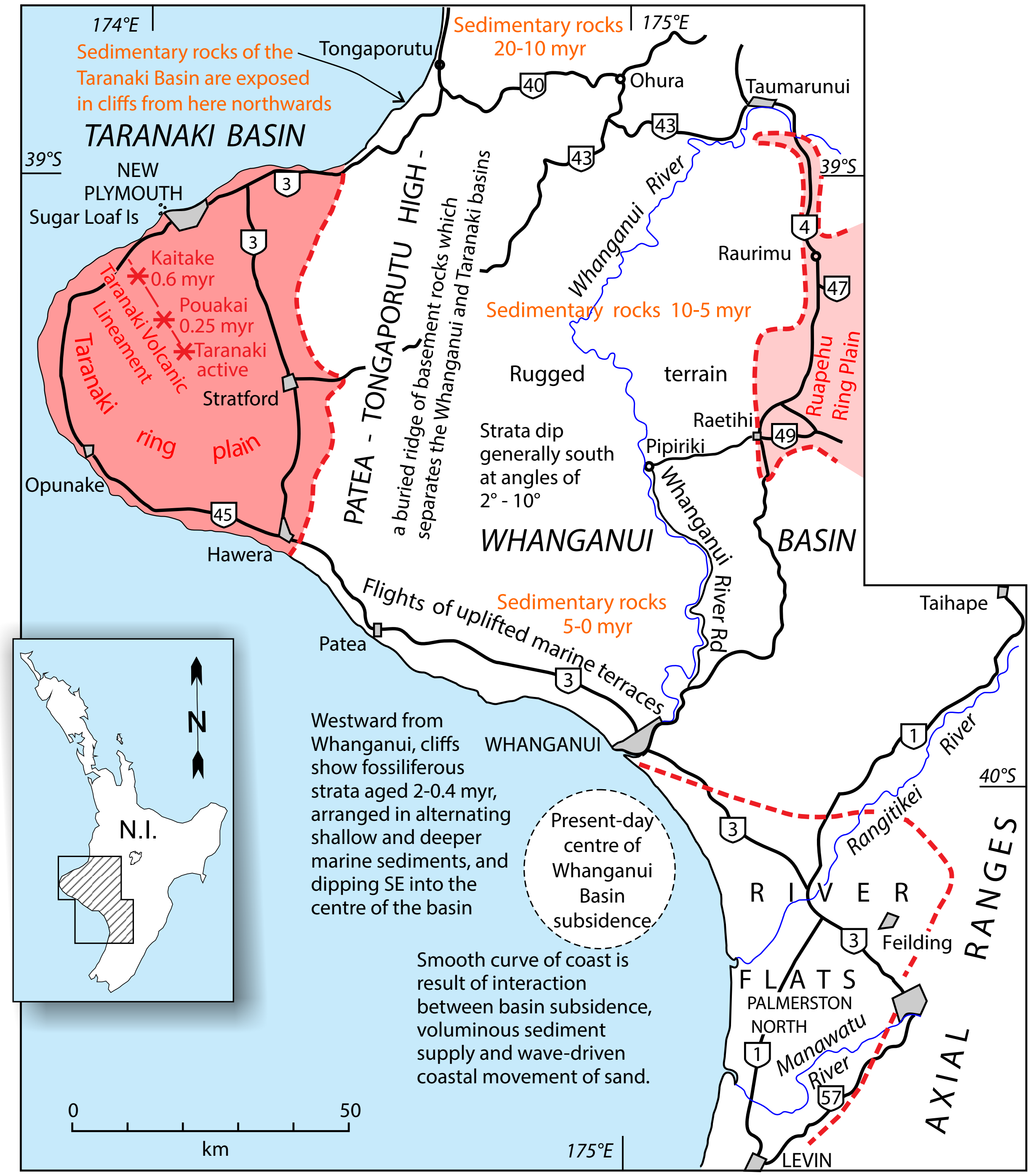


Fig. 10.1. Simplified geology of the Whanganui-Taranaki-northern Manawatu area.

Whanganui Basin is one of New Zealand's more remarkable geological phenomena. It is a sedimentary basin containing cover strata that are up to 5 km thick under the South Taranaki Bight – that at the present-day is the centre of basin subsidence. This centre, however, is only the latest in a series of basin centres, extending back 25 myr in time, and to the central King Country, 200 km to the north. The effect of the progressive southwards shift of the basin centre, as shown in Box 10.1, has been to stack strata in a shingled fashion, like a shelf of fallen-over books. If the thickness of all those shingled strata is measured at right angles to the bedding planes (the traditional and obvious way of measuring thickness of strata), it adds up to much more than 5 km. However, 5 km is the maximum thickness (vertical height) present at any one place – using the bookshelf analogy, the combined thickness of fallen-over books is greater than the height of the bookshelf.

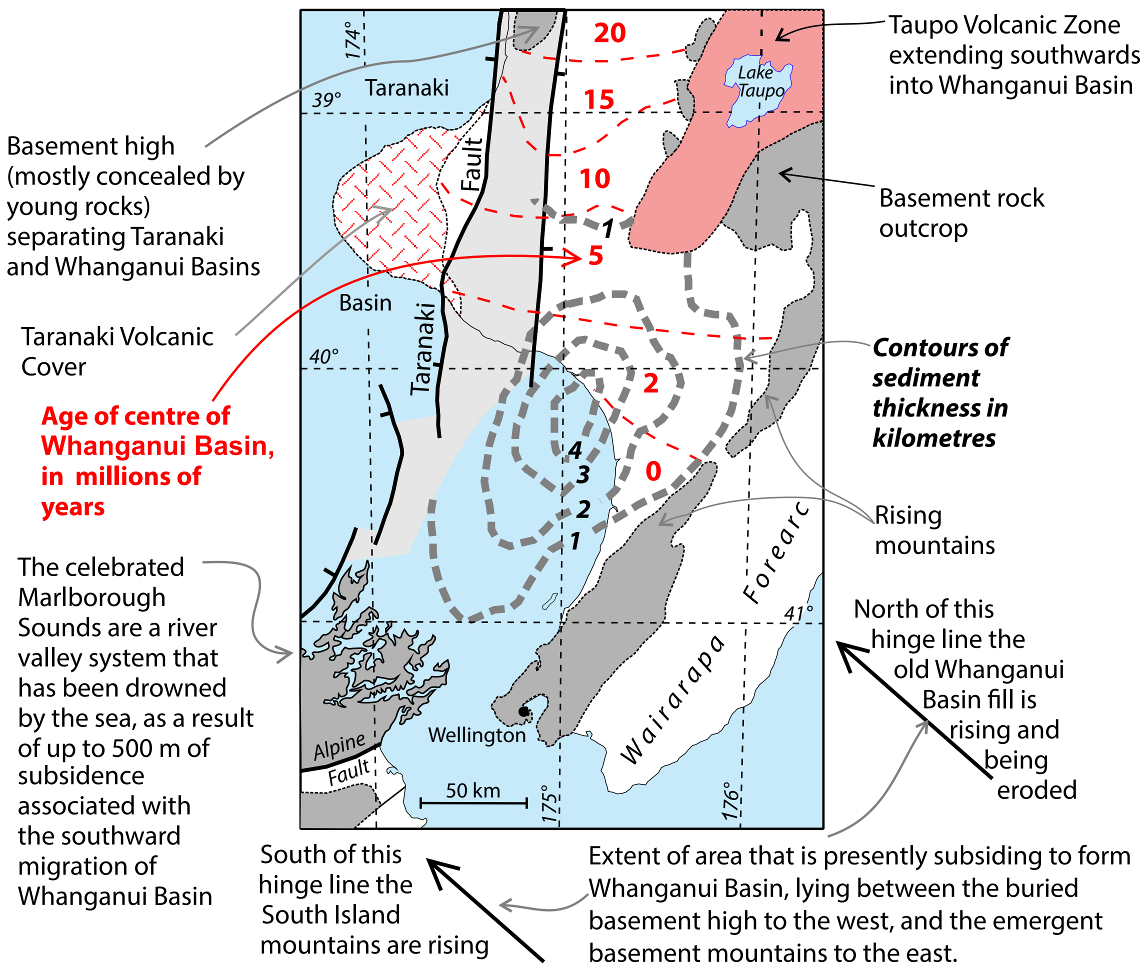
In the northern, older part of the basin, around Te Kuiti and Taumarunui, Whanganui Basin strata overlie the limestones of Te Kuiti and Waitomo (Chapter 7). The change from deposition of thin limestone strata 30 myr ago (continental shelf), to much thicker mudstones and flysch sediments 25 myr ago (deep sedimentary basin), was one of the countrywide responses to the great increase in activity on the Pacific–Australian plate boundary at that time. As noted in the previous chapter, the Whanganui Basin has always had a location to the south of the termination of the volcanic arc(s) of the time.

The Moving Basin

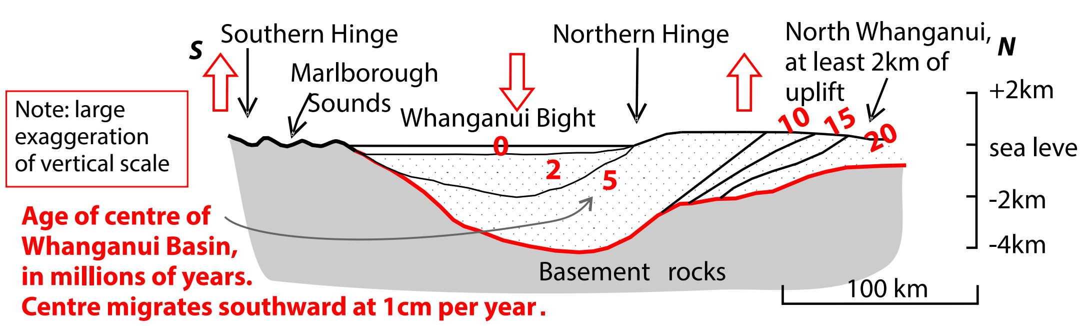
The first Whanganui Basin was centred on Taumarunui, and we see the flysch deposits exposed extensively around that area now. The question of why we can see them, and why they are not hidden underneath all the younger Whanganui Basin sediments, has to do with the way in which the centre of the basin has shifted. As the basin has moved southwards, building a shingled stack of strata, a tectonic hinge-line has moved with it, such that older strata north of the line are uplifted and eroded while new strata are deposited in the subsiding area south of the line. As shown in the cross section through the basin in Box 10.1, this has caused the north–south shingled arrangement of strata relative to the basement; to use geologists' jargon, the strata downlap to the south onto basement rocks. This arrangement also has the effect that virtually all the outcropping strata in the Whanganui Basin dip to the south, and that the tectonic hinge-line effectively determines the location of the coastline of the South Taranaki Bight. You can see examples of continuous south-dipping strata if you drive along SH1 between Waiouru and Bulls (or take the train between the same two places), or along SH4 between Te Kuiti and Whanganui, or if you take the Whanganui River trip between Taumarunui and Whanganui.

The most striking development of the basin has taken place over the past 5 myr, as the cross section in 10.1 shows. Somehow, this fact is related to developments in the active volcanic arc. For all of its history the basin has been situated in the centre of the North Island and just to the south of the southern termination of the active volcanic arcs (Box 2.3). As the arcs have moved, so has the basin. For the past 2 myr the basin centre has lain exactly in line with the Taupo Volcanic Zone (TVZ), and as the basin centre has moved towards the south-southwest, so the TVZ has advanced into the northern part of the basin at the same rate, as Box 10.1 shows. The extra uplift associated with

Box 10.1. Whanganui Basin.



Whanganui Basin is a long-lived, major sedimentary basin. It has existed for 25 million years, moving steadily southward. Its origin is linked, in some way that is poorly understood, with the active volcanic arcs situated to the north of it.



North - south profile through Whanganui Basin to show how the basin has migrated steadily southward for 20 million years. During that time it has maintained a position adjacent to the southern limit of the active volcanic arc.

Fig. 10.2. Thin-bedded flysch deposits that were deposited about 20 myrs ago in the northern part of the Whanganui Basin around Taumaranui. Herlihy Bluff, alongside the Whanganui River, 5 km southwest of Taumaranui. Photographer Bruce Hayward.

heating of rocks under the TVZ has accentuated the rise of the older part of the basin, north of the tectonic hinge, as seen spectacularly around Waiouru.

The deepest part of the basin is marked by the strongest negative gravity anomaly in New Zealand. That is, there is a marked deficiency of gravity – far too small for you or me to notice, but detectable instrumentally. This means that there is a large excess of lightweight rock beneath the centre of the basin. That rock is presumably continental crustal rock being pulled down into heavier mantle rocks (it has to be pulled down – it cannot just sink, because it is lighter in weight and should therefore float on the heavy rock).

Both the TVZ and the Whanganui Basin are driven by subduction of the Pacific Plate underneath the North Island. Box 13.2 A shows the relationships – the subducting Pacific lithospheric slab extends to a depth of at least 200 km (as indicated by earthquakes) as far south as the northern South Island, 300 km south of Mt Ruapehu, the southernmost volcano of the Tonga–Kermadec–New Zealand volcanic arc. It is not yet understood why the volcanic arc stops where it does, because conditions should favour magma generation well south of Ruapehu. As noted in Chapter 9, it may be to do with the increasing obliquity of subduction to the south, which gives the downgoing slab a longer travel time to the critical magma-generation depth of 100 km. But instead of a continuing arc we have the Whanganui Basin. There is clearly a first-order tectonic relationship between these two quite different, global-scale phenomena – arc and basin – but how it works is not yet known.

Two more consequences of the active nature of the Whanganui Basin and its southwards migration are frequent shallow earthquakes that occur underneath Whanganui, and the tectonic subsidence creating the South Island's Marlborough Sounds. The sounds are drowned river valleys, but while they were certainly affected by the glacially driven ups and downs of world sea-level like the rest of the New Zealand coastline (Boxes 10.2 A, B), up to 500 m of subsidence at the southern margin of the Whanganui Basin has drowned them much more profoundly. The sounds area will eventually disappear beneath the sea, but the subsidence is slow.

Cyclic Sedimentation

A characteristic feature of the deposition of sediment in the Whanganui Basin over the past 5 myr or so has been a regular cyclicity – an alternation between shallow marine sediments (sandstones, gravels and shell beds) and deeper marine, outer continental-shelf sediments (grey mudstones). The more resistant sandstones stand out as ledges in the hillsides and create rapids in the rivers. These sediment cycles were caused by scores of glacial–interglacial climatic cycles.

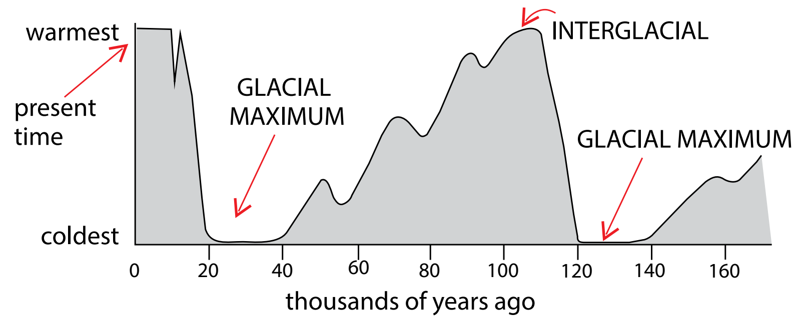
We are familiar with the notion of a few recent glaciations, but we need to grasp the fact that the three astronomical cycles (called Milankovitch Cycles) that drive glaciations have been in place for hundreds, if not thousands, of millions of years. They happen to be having their maximum effect at the present time because of the earth's current geography (including the presence of a circum-Antarctic seaway and lack of a circum-equatorial seaway), but they can be detected in strata from 100 myr ago, when there were no ice caps.

The important thing about glacial–interglacial cycles is that they make sea-level move up and down, by as much as 130 m, every 40,000–100,000 years (Boxes 10.2 A, B). It turns out that the Whanganui Basin has a very full record of these cycles for the past 5 myr, making it an important international reference basin for them. By extraordinary coincidence, the subsidence rate of the basin, in combination with the sedimentation rate, was exactly right to keep the water depth in the correct range to record the ups and downs to maximum effect (too far offshore and changing sea-level has little effect on the sediment; too near shore and much of the sediment record is lost to erosion). The shallow marine sandstones etc. record glaciations, when sea-level was low, while the intervening deeper-water grey mudstones record interglacial periods when the sea-level was high and the coastline retreated inland. There was a sufficiently high sediment input to record almost all the cycles, such that only one or two are missing.

As explained in Box 10.2 A, these climatic cycles have become an important part of the global geological timescale, because they are synchronous all over the world. They are known as 'oxygen isotope stages', the name referring to the fact that oxygen bound up in the lime shells of marine plankton contains a ratio of the isotopes oxygen-16 to oxygen-18 that is partly governed by sea-water temperature at the surface; shells from core samples from deep-sea oozes therefore contain a cyclic climate record that is complementary to the cyclic sedimentation record. The Whanganui Basin, as noted, occupies a central role in the overall story. By happy coincidence, the southwards migration of the hinge-line between subsidence and uplift has left all the older cyclic strata (5–2 myr) nicely uplifted and exposed in the Rangitikei and Whanganui valleys. Younger cycles are seen at the coast, between Hawera and Whanganui.

There is more. By themselves, glacial climatic cycles are simply on-off signals, each one exactly like the others. Independent time markers are needed to enable them to be correlated around the world, and in the Whanganui Basin we have two lots of these. One is occasional layers of volcanic ash – remember that there have always been active volcanoes to the north of the basin – that have been dated by radiometric means. The other is periodic reversals of the earth's magnetic field, which are recorded by faint magnetic signals in the mudstones. Fossil dating gives a broader indication of time that backs up the other two methods. When all three are combined, the record of glacial climatic cycles in the Whanganui Basin is very well tied down, and well correlated with other global records, making it of considerable international scientific importance.

Box 10.2 A. Glacial-interglacial changes of sea-level, and how they are recorded in subsiding basins.

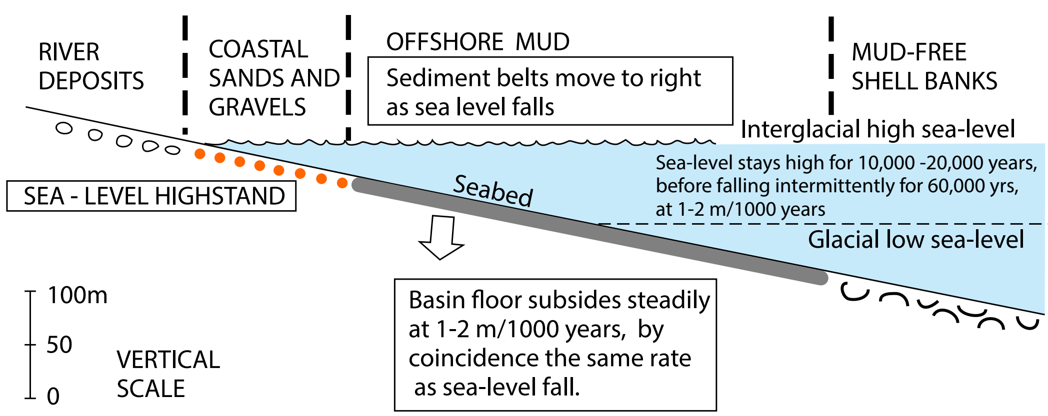


The graph shows a typical trace of temperature change and sea-level change during a glacial-interglacial-glacial cycle lasting around 100,000 years, based on the two most recent glaciations. The line can represent either global temperature or global sea-level. This is because change of sea-level mimics change of temperature very closely, as water is either locked up in ice caps or released from ice caps. The lopsided nature of the graph arises because global climate change is driven by variation in the input of heat from the sun, which is determined by three quite independent cyclic factors:

1) the earth's orbit round the sun varies over time from nearly circular to elliptical and back again, causing the distance from the sun to vary ('eccentricity of orbit', length of cycle about 100,000 years).
2) the inclination of earth's spin axis to the plane of orbit, which causes the seasons and is at present 23.5°, varies between 24.5° and 21.5° ('obliquity of the ecliptic') on a cycle of about 40,000 years.
3) the earth wobbles about its spin axis, like a spinning top, on a cycle of about 20,000 years, causing the 'precession of the equinoxes'.

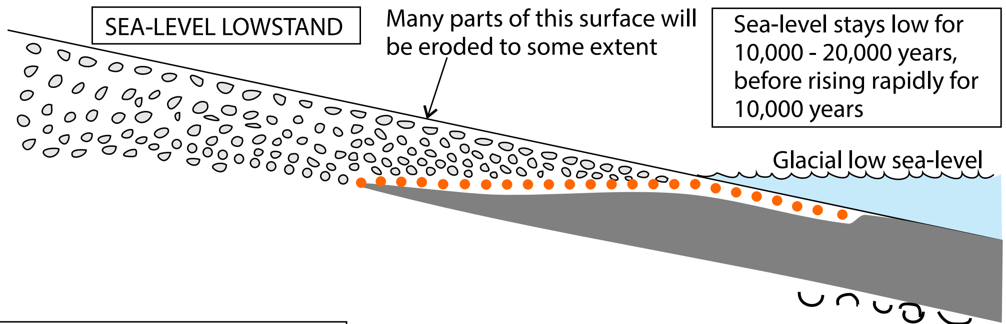
The three cycles have unequal effects on climate, (1) being strongest. Their combined effect is technically a Fourier series curve, with relatively brief glacial maxima and minima occurring on average every 100,000 years. The cycles are known as **Milankovitch Cycles**, after the Yugoslavian astronomer who first worked them out. At the present time we are probably nearing the end of an interglacial climate maximum, which has already lasted 10,000 years.

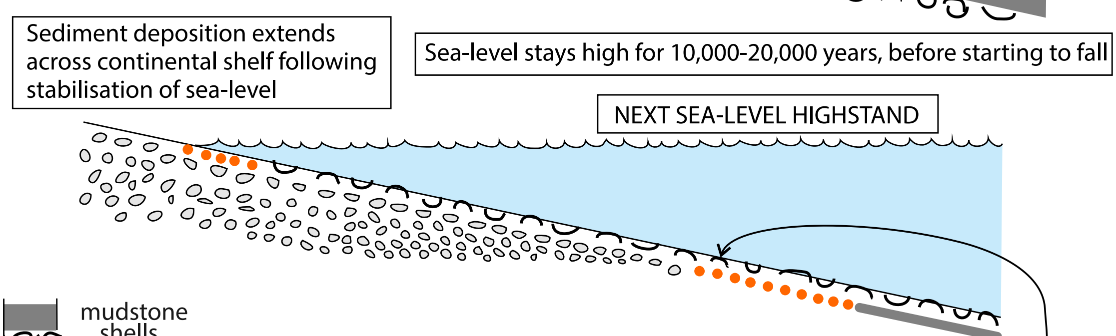
Diagrammatic cross section of a sedimentary basin, between the coast and a water depth of about 150 metres, showing the distribution of bottom sediment.
Vertical scale is greatly exaggerated.



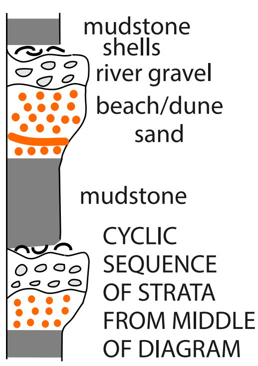
(continued in Box 10.2.B on next page)

Box 10.2 B. Glacial-interglacial changes of sea-level *(continued).*





Post-glacial sea-level rise averages 1 metre/century, leaving little accumulation of sediment — perhaps local patches of beach sand. Being sediment-starved the sea bed will accumulate mostly shells, until nearshore sand and offshore mud belts are re-established and extend seawards to cover the shells.



The pattern of sediment distribution in vertical section, as shown, is a result of the combination of sea-level fall and basin subsidence.

Vertical sequences of strata formed in a situation like this one will be repetitive (or cyclic). The precise nature of the repetition will depend on just where the particular section that we see was located across this diagram. At the far right it would be mostly mudstone with shell beds. At the far left it would be entirely river deposits, probably with erosion surfaces. The middle would contain the greatest variety of sediments, with cycles that become coarser grained upwards, from mud to sand/gravel, ending in erosion surfaces, and a return to mudstone. Cycles range up to 100 metres thick, depending on rate of subsidence and sediment supply, but are generally less than 50 m. Many variations occur from the basic model shown. In order for such sequences to form, and to be preserved from erosion, the rate of subsidence must be quite high - around 1 metre per thousand years or higher. The diagram shows only a fairly shallow basin.

Sea-level change and deep-water submarine fans

The effects of sea-level change are also felt in deep water, because during low sea-levels rivers take sediment to the edge of the continental shelf, from where it is taken by turbidity currents (Box 6.5 B) down the continental slope, to build submarine fans at the base of the slope. A method of analysis of strata in wide use in the oil exploration industry is based on the principles outlined above. It is called 'sequence stratigraphy'.

Long-term climatic change

Climatic changes caused by the Milankovitch Cycles have always been present, but they are most noticeable when global climate and global geography are delicately poised with respect to ice-sheet formation, as they are at present. Climate has been moving steadily further into "ice-house" mode (as opposed to the earlier, more equable, 'greenhouse' mode) during the past 50 million years. The change has been driven by changes in oceanic circulation that have been caused by continental drift (specifically, the closing of a global equatorial seaway and the gradual widening of a circum-Antarctic seaway). There have been hundreds of glaciations, and their effects can be found in sequences of strata all over the world, though they may be quite subtle. However, glaciations have been getting steadily more severe, and the last few have been the most severe.

Fig. 10.3. Cyclic south-dipping strata deposited during the climate and sea-level cycles about 3 myrs ago and now uplifted and eroding in Rangitkei Valley. Photographer Brent Alloway.

Slope Failures

There are interesting (and expensive) consequences of the see-saw down and up history of the Whanganui Basin. As noted earlier, the hinge-line separating subsidence to the south from uplift to the north is moving progressively southwards. The uplift to the north has maintained a strong flow of sediment into the basin to the south, so that strata here are quite thick. However, the see-saw tectonic action has meant that these thick strata, dominated by mudstones, have never been deeply buried and compacted. As a result, to use the jargon, they are undercompacted, and lack the strength that compaction provides.

The combination of low-strength muddy sedimentary rocks, and rapid uplift leading to deep river incision and steep slopes, is dynamite when it comes to slope stability, especially when the forest is cleared. Slope failures (slips in the common usage) are widespread in the region (Boxes 12.5 A–C). Part of Taihape itself is built on a large old landslip. Further south, the lumpy terrain a little way south of Utiku settlement is all slipped ground. The railway line here is cut into the steep west bank of the deeply incised Rangitikei River. The line had to be completely relocated in the late 1990s, because a portion of the hill is sliding towards the river on a layer of clay just a few millimetres thick.

There is a dramatic example of river downcutting as a result of rapid uplift 6 km south of Taihape. Turn into the road signed to Ohotu, Omatane and Taoroa Junction, and just a few metres east of SH1 stop at the bridge over the Hautapu River. The river here is in a slot approaching 50 m deep.

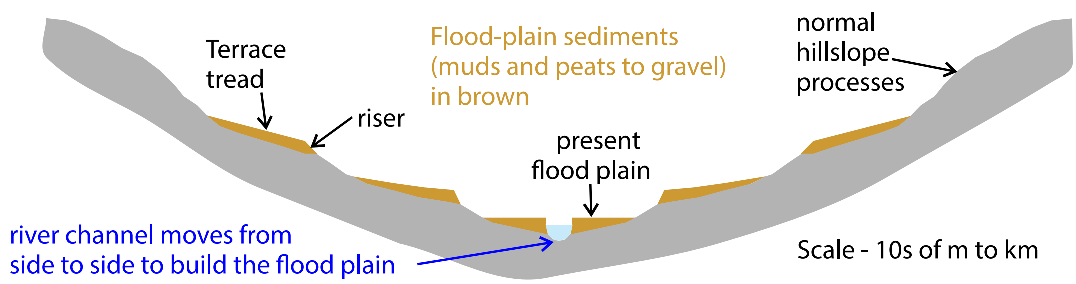
Another consequence of rapid uplift and deep incision of rivers is the widespread preservation of river terraces – uplifted portions of river floodplains (10.3, top). Flat terrace landforms occur at various heights above the river, and at the tops of many riverside cliffs you can see thin river gravel deposits underlying terrace surfaces (brown gravel resting on grey mudstone).

Box 10.3. Terrace landforms – steps on hillsides.

1. River terraces.

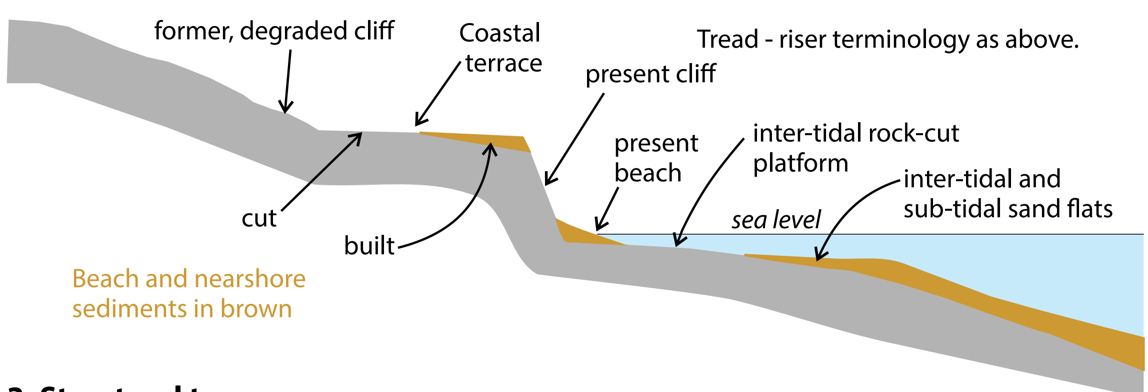
Each river terrace is a portion of a former flood-plain. Staircase terminology — treads and risers — is used for flights of terraces.
River terraces signify progressive downcutting by the river. This may have several causes -

a) lowering of base level (sea level, lake level, sill level)
b) uplift of the land
c) increased flow rate due to river capture in the headwaters.



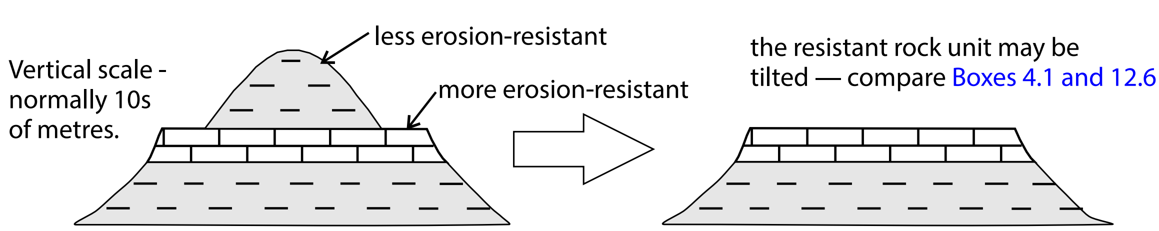
2. Coastal terraces

Each coastal terrace is a former inter-tidal platform. It can be wholly rock-cut, wholly sediment-built, or a combination of the two, as illustrated. A coastal terrace generally signifies uplift of the land. Coastal terraces are preserved from metres to 100s of metres above sea level.



3. Structural terraces

These landforms result from erosional removal of less-resistant rocks from above more-resistant ones.



Note that river and coastal terrace surfaces are not horizontal. There is always a slope towards the river or the coast.

Manawatu Country – a Multi-river Delta

Yet another consequence of the tectonic see-saw of the Whanganui Basin, combined with the unrelated uplift of axial ranges to the north and east, is a focusing of the lower reaches of five large rivers that have built the Manawatu – a large triangular delta deposited by the Whanganui, Whangaehu, Turakina, Rangitikei and Manawatu rivers. This is the area marked 'River Flats' on Fig. 10.1.

The delta has been built in post-glacial time (the last 15,000 years) by the five rivers, which have been sucked in, as it were, to the centre of basin subsidence. Together they drain a huge area of the North Island, but they all discharge into the sea in a 60 km stretch of coastline centred on the region of maximum subsidence.

The smoothly curved, aggrading coastline between Whanganui and Kapiti Island is the result of voluminous sediment supply, subsidence offshore, and strong wave and tidal activity in Cook Strait that is pushing sediment to the south. The recently aggraded sediments include sand-dune fields, now cut off from the coast, forming lumpy terrain covered with grass. The extensive wetlands that formed on the dune fields (fewer now than there used to be) are biologically important.

Later on in the Glacial Cycle Story

The coastal strip of the Whanganui Basin, between Hawera and Whanganui, contains the most complete section of very young (geologically speaking) marine sedimentary rocks in the country. It has long been the type area for our fossil-based time divisions for the past 3.5 myr, as shown by their names: the Whanganui Series is split into Waipipian, Mangapanian, Nukumaruan, Castlecliffian and Hawera stages, all called after local places. These names are used all over the country, and there is a parallel rock nomenclature for the local rock formations, again based on local place-names, e.g. Kai-iwi Siltstone.

These young strata are commonly richly fossiliferous, containing molluscan shells of shallow marine affiliation. Like the other young sediments of the Whanganui Basin, the strata are cyclic, alternating between near-shore sandy and pebbly sediments with fossils (glacial low sea-levels), and offshore grey mudstones and siltstones with fewer fossils (interglacial high sea-levels) – Box 10.2 B.

The fact that the cycles are both preserved and exposed to view is due to a neat combination of circumstances. As noted above, the centre of subsidence of the Whanganui Basin has been migrating, more recently in a clockwise curve. Thus subsidence has moved southeastwards (i.e. parallel to the coast), so that the areas around Hawera, then Patea, and finally Whanganui were affected first by subsidence, allowing preservation of the sediment cycles while sea-level moved up and down in response to glaciations, and then by the uplift that has affected everywhere north of the present coastline (the tectonic hinge-line).

As a result, along the coast from Hawera to Whanganui we see more than 50 layers of conglomerate (old gravel), sandstone, shell-beds and intervening siltstones. They dip very gently to the southeast, taking successively older layers below sea-level and allowing successively younger beds to occupy the cliff in a southeasterly direction towards Whanganui. There are fault zones that move the strata up and down somewhat, although not by large amounts, at Waipipi Point, Waverley Beach and Nukumaru Beach. The age range of the dipping strata is from approximately 2 myr to 450,000 years.

The coast along here is accessible at many places, but as with any coastal trips, take care and visit only at low tide. The fossiliferous sandy beds normally form intertidal reefs as well as being exposed in the cliff. The deeper-water grey mudstones generally recede in the shore platform, but are also visible in the cliff. You should be able to get a good impression of this important stratigraphic record more or less anywhere along here.

Ties to the Taupo Volcanic Zone, and Beyond

Large amounts of volcanic ash and pumice, ejected during big ignimbrite eruptions in the TVZ, have been transported into the Whanganui Basin down the various rivers that flow out of the TVZ. Concentrated ash and pumice layers are generally white, and they can be dated radiometrically (Chapter 1). They provide an important link between the fossil-based subdivisions of the strata, and absolute time. As noted earlier, this link can be carried further, to the so-called oxygen isotope timescale that records the global glacial–interglacial cycles in deep-sea sediments, and further still to the global timescale of magnetic reversals. Thus the Hawera–Whanganui coastal section is very important for the global geological timescale for the past 2 myr, complementing the Rangitikei Valley (SH1), where the record for the 2–5 myr period is well displayed.

Coastal Terraces

While the strata in the cliffs are one component of the young Whanganui Basin story, the coastal terrace landforms are an even younger component (Box 10.3: 2.). They give information for the last 400,000 years or so, to complete a remarkable record. Gradual uplift of the area moves places from below sea-level to above it. While passing through the intertidal region, an intertidal platform is both eroded from the pre-existing rock and built from sands and muds – exactly as is happening at present. The platform cut during a high interglacial sea-level becomes a coastal terrace when raised above sea-level during the following glaciation.

Given what was said earlier about the southeasterly tilting of strata along the Hawera–Whanganui coast, you should expect to see a flat top to the cliffs here now, representing an uplifted intertidal platform. The flat top may be cut across the dipping strata directly (an old rock platform), or it may be underlain by a few metres of sand (an old sandflat) that rests on a surface cut across the older dipping strata. This is known as an unconformity (Box 4.1). These thin sediment sequences are another important component of the geological history of this region, the youngest component. They, too, may be fossiliferous. Higher terraces, located inland, are older than those at the coast.

Terrace surfaces always slope gently towards the sea, but are usually planar. Locally they may be lumpy if sand dunes were present. They are dissected into separate areas by stream valleys that are deepening in response to the uplift.

Altogether, the Whanganui Basin, past and present, covers an area of around 30,000 sq km. It contains a huge volume of sediment, despite the fact that a lot of it in the north has been eroded to feed the hungry monster to the south. It's a cannibalistic system in part. With its unique tectonic position, and unique record of glacial sea-level control, there isn't another sedimentary basin in the world like it.

Fig. 10.4. Sea cliffs west of Whanganui composed of young marine sedimentary rocks capped by an uplifted flat coastal terrace that was eroded intertidally over the top during the Last Interglacial period about 120,000 years ago. Photographer Brent Alloway.

Fig. 10.5. A series of higher and higher terraces extend inland for many kilometres from the Whanganui coast. Each formed as an intertidal platform during a high interglacial sea level and has been progressively uplifted and cut into by stream erosion. Kai Iwi Beach, west of Whanganui City. Photographer Lloyd Homer, GNS Science.

Chapter 11

Taranaki: the Basin and the Volcano

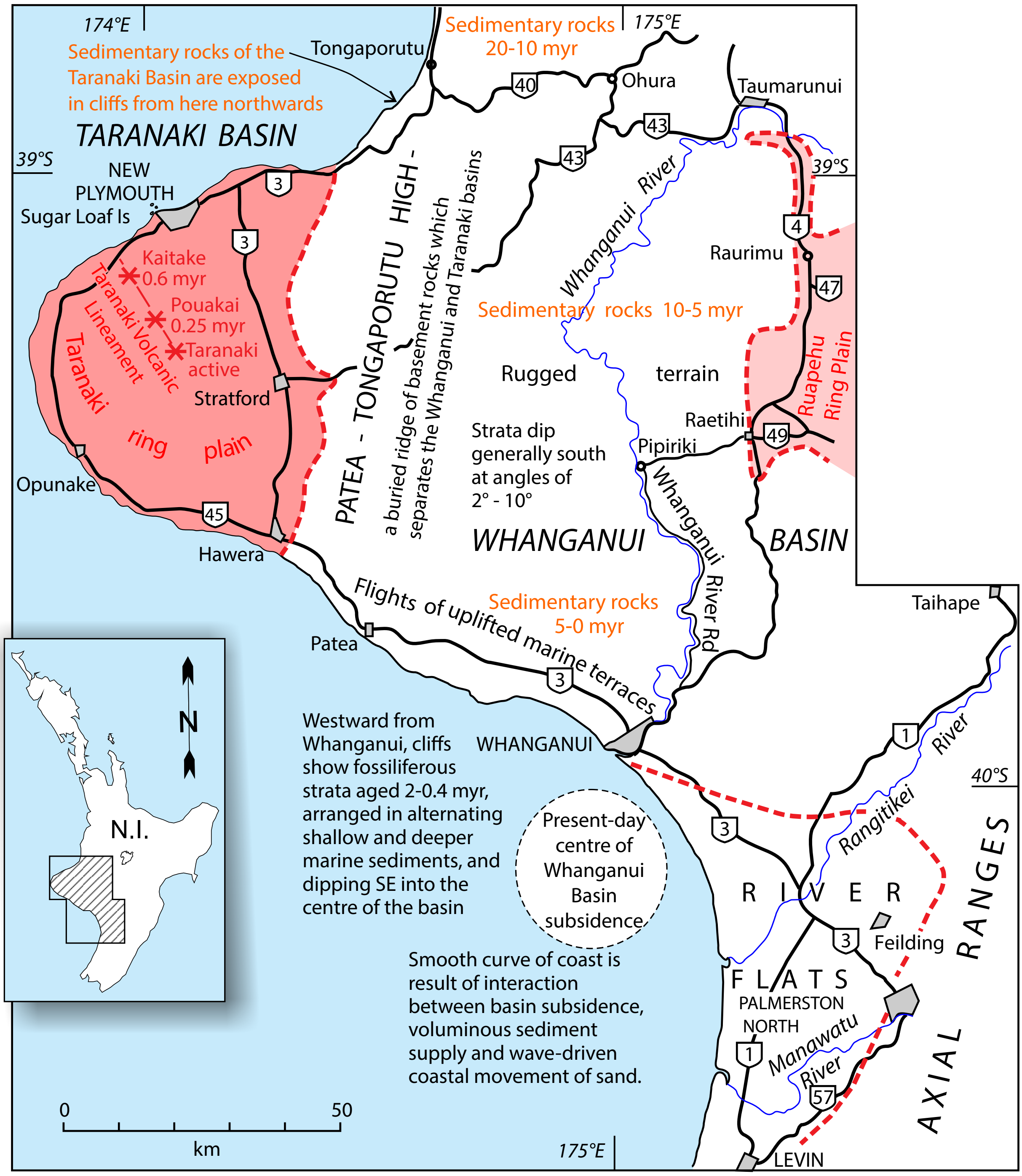


Fig. 11.1. Simplified geology of the Taranaki and Whanganui basins *(Fig. 10.1 repeated).*

The Taranaki region has two unique geological features: Mt Taranaki volcano on top; and the Taranaki Basin, New Zealand's only producing hydrocarbon basin (at the time of writing) underneath. The basin is largely hidden, either offshore or beneath the deposits of Mt Taranaki, which is New Zealand's only active 'behind-arc' andesitic volcano. All communities in the region, including the gas and oil town of New Plymouth, lie within the hazard zone of Mt Taranaki. The volcano has been asleep for 250 years, but it could erupt at any time.

Taranaki Basin

Taranaki Basin is a close neighbour to the Whanganui Basin (Chapter 10). These two very large sedimentary basins are separated by just a ridge of basement rocks, called the Patea–Tongaporutu High. This ridge is mostly buried underneath the region's youngest rocks, but it comes to the surface as the Herangi Range north of Awakino. However, despite their close proximity and the fact that they were both being formed and filled over much the same time period, the two basins have absolutely nothing in common. Their tectonic origins and detailed history are completely different.

Taranaki Basin is a particularly long-lived sedimentary basin (a large crustal depression filled with sedimentary rocks). It has had two quite distinct periods in its development, which correspond with two of the three most recent major phases of New Zealand's development. The basin originated during phase two of our development – the break-up of 'our' part of Gondwana. At this stage (100–80 myr ago) it was a rift valley, an elongate, fault-bounded depression caused by stretching of the crust. During continental stretching and break-up, several rifts are commonly initially active, but only one or two go on to develop into new centres of seafloor spreading. In our case, the Taranaki Rift became inactive, and another rift developed to become the focus of seafloor spreading, ultimately creating the Tasman Sea (Boxes 5.5 A, B).

Failed spreading rifts (called aulacogens) have considerable economic importance, because the sediment fill of the rift valley commonly contains coal measures and/or hydrocarbon (oil and gas) source rocks. The North Sea and Nigerian oil fields are two examples. In our case, Taranaki Basin contains thick coal measure deposits, and it is these that became the main source for oil and gas.

To make hydrocarbons, the source rocks (either coals or carbon-rich marine muddy sediments) have to be pressure-cooked to the correct recipe. This involves being heated to 100–120°C and held at that temperature for some time. The only way to achieve that is by deep burial – to around 5 km at a geothermal gradient averaging 20°C/km. This happened in the Taranaki Basin during the second stage of basin development, which began 25 myr ago when the Pacific–Australian plate boundary was energised, and coincided with phase four of New Zealand's geological history.

In addition to source materials and deep burial, critical factors in the formation of hydrocarbons are permeable rock bodies to allow the oil and gas to migrate, and large, simple traps, capped by impermeable rocks like mudstone/shale, to hold them. Taranaki has the permeable sandstone and limestone layers that reservoir (store) the oil and gas, as well as trap layers. The final critical factor is timing – if the traps are not in place when the oil migrates, the oil moves up to the surface and is lost instead of being reservoired.

The location of the Taranaki Basin, 300–400 km west of (behind) the plate boundary subduction zone is a vital part of the story. That location ensured that the region received mild compression from the east, but not so strong that the fold structures became overly complex.

The kick-starting of the plate boundary 25 myr ago reactivated the old rift faults. These faults, being rift-bounding, were 'normal' extensional faults. However, they were now responding to the opposite process – compression. Thus they switched to a reverse (shortening) mode, in contrast to their original normal (stretching) mode (Box 6.6 C).

A major overthrust of basement rocks took place along the eastern side of the basin. These rocks were pushed west by the compression. This was very important, because at this time the coal measure hydrocarbon source rocks had only a thin cover of sediment. The weight of the basement rocks caused further depression of the original rift, and allowed several more kilometres of sediment to accumulate. In this regard it was also important that Taranaki was close enough to the newly rising New Zealand landmass to be supplied with plenty of sediment. A sedimentary basin of this type, depressed ahead of an advancing mass of heavy basement rocks, is called a foreland basin. There is one such basin presently forming in New Zealand, namely the Moutere Depression/ Tasman Bay Basin in the northern South Island.

Deep burial of the coal measures to 5 km was accomplished around 10–15 myr ago, when their temperature was finally raised into the optimum cooking range to form hydrocarbons. As long as source rocks stay buried and out of reach of oxygen (which would oxidise the essential carbon), there can be long delays between the original deposition and their eventual cooking; in our case, this delay was about 85 myr.

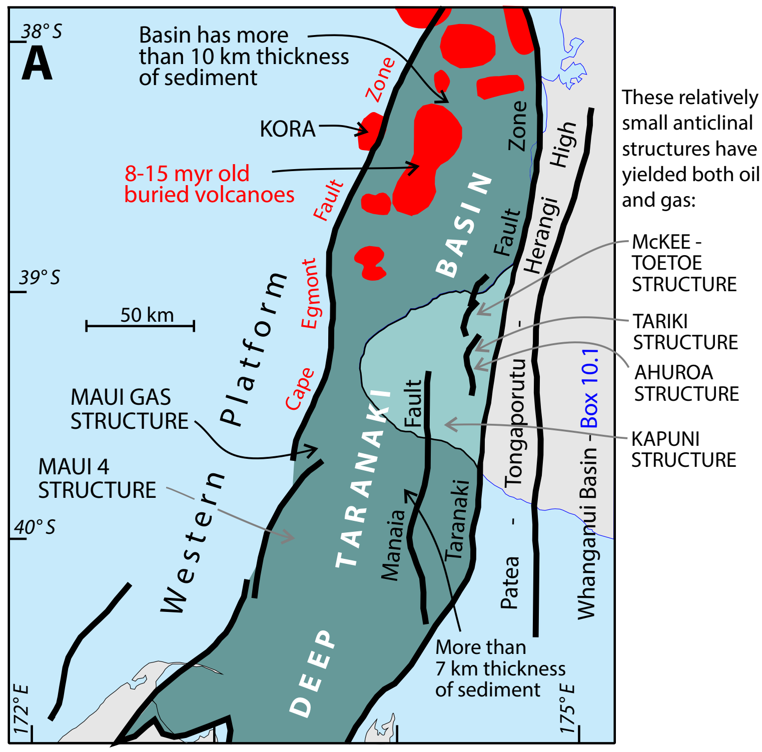
The next essential step in the development of a hydrocarbon field is that the traps, sealed by impermeable mudstones, are in place to catch the migrating hydrocarbons. In Taranaki, some oil and gas was trapped, but not all – some escaped to the seabed and was lost.

So you see, a considerable number of geological events have to be got just right, and occur in the correct order, before a productive hydrocarbon field eventuates. While New Zealand has other prospective hydrocarbon basins, none of them has been quite as favourable a sequence as Taranaki Basin, and to date (2017) Taranaki is our only producing basin. Because nearly all of the basin is either buried underneath Mt Taranaki or located offshore, our huge knowledge of it is derived mostly from work by hydrocarbon exploration and production companies.

Box 11.1 A. Taranaki Basin.

Taranaki Basin is New Zealand's only commercially producing oil and gas basin. A potentially commercial gas field was discovered in Hawkes Bay at the close of the 20th century but to date has not gone into production.
Taranaki Basin extends from the northern tip of the South Island, underneath the Taranaki Peninsula, and as far north as Cape Reinga, a distance of about 800 km. However, it has not been extensively explored north of offshore Raglan Harbour. The first wells drilled offshore of Northland, in 2000, were unsuccessful.

A remarkable feature of Taranaki Basin is that it lies almost entirely offshore, except for the Taranaki Peninsula, where it is concealed by the young volcanic deposits from Mt Taranaki. In fact, the only places where the rocks filling Taranaki Basin are exposed to view are a narrow coastal strip between Kawhia Harbour and Urenui, and northernmost South Island around Cape Farewell and Whanganui Inlet. Our knowledge of this interesting and important sedimentary basin, which was first brought to attention by oil seeps and early oil wells (1865) in New Plymouth, is due almost entirely to seismic profiles, geophysical surveys and oil wells.



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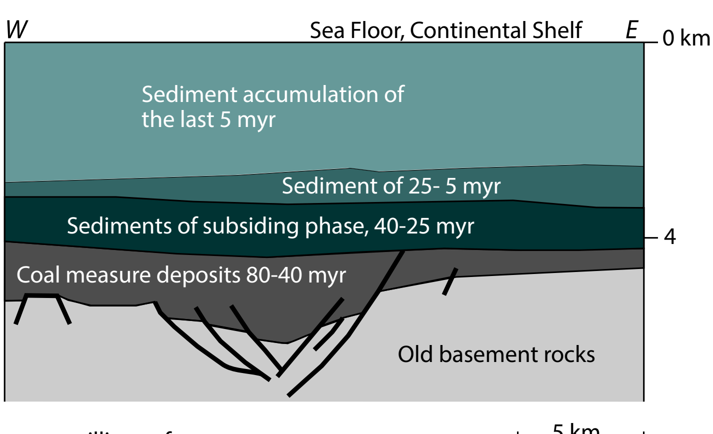

myr = millions of years

Cross section of the Western Platform part of Taranaki Basin, showing one of the 80 myr old rift valleys dating from the opening of the Tasman Sea, now buried beneath several kilometres of younger sediment. From an oil industry seismic reflection profile. Note the great sediment thickness of the past 5 myr - this reflects the erosion of the rising Southern Alps (South Island) when they were further north than they are now, and before Cook Strait was formed.

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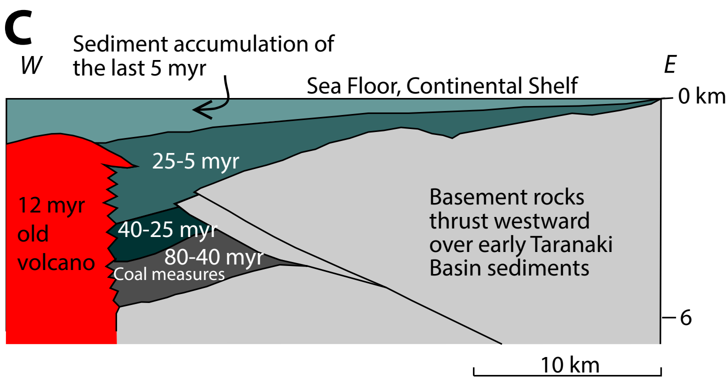
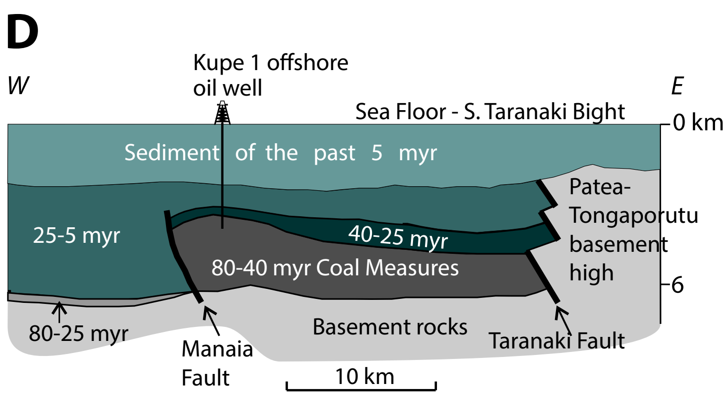
Box 11.1 B. Taranaki Basin *(continued from previous page).*



Cross section across the eastern margin of Taranaki Basin, north of Taranaki Peninsula, showing basement rocks thrust westward over early Taranaki Basin sediments, in response to compression exerted by the North Island subduction zone, starting at 25 myr. Also shows one of the 12 myr old buried volcanoes that are present in the northern part of Taranaki Basin.

Note that the volcano was not finally buried by sediment until less than 5 myr ago. This has important economic implications, because oil and gas that was generated in the adjacent buried coal measure sediments, about 10 myr ago, was able to migrate upwards through the fractured and porous volcanic structure of Kora Volcano, and escape into the sea because of the absence of a sediment seal. Other volcanoes have still to be drilled. Cross section from an oil industry seismic reflection profile.



Cross section across the eastern margin of Taranaki Basin, south of Taranaki Peninsula, showing the two main thrust faults - Taranaki and Manaia - which formed in response to the compression from the east that was initiated 25 myr ago.
The arch structure produced above the Manaia Fault was the target of Kupe 1 offshore oil well. It was hoped that oil and gas generated within the coal measures were held in porous sedimentary rock and trapped under the crest of the arch. The well was unsuccessful.

Note that if the post-25 myr thrust movement on Manaia Fault is restored so that the top of the 80-25 myr sediments becomes a continuous line, as it would have been 25 myr ago, then the fault is still required in order to explain the great thickening of the 80-25 myr sediments to the east of it. In other words, Manaia Fault already existed 80 myr ago as a 'normal' or extensional fault, bounding one side of a rift valley in which coal measure sediments (Box 7.3 A) accumulated.
Taranaki Fault formed the other side of the rift valley. It was one of a whole family of rift structures that formed around New Zealand during the crustal stretching phase that preceded the opening of the Tasman Sea and the separation of New Zealand from Gondwana (Boxes 5.5 A, B).
Both faults were re-activated during the crustal shortening phase that began 25 myr ago. Compression was due to plate convergence across the North Island subduction zone, and in Taranaki was directed westwards. Both faults changed style from 'normal' to 'high-angle thrust' (Box 6.6 C), and the old rift valley was everted (turned inside out).

Box 11.1 C. Why the Taranaki Basin is producing oil and gas.

Before a sedimentary basin — region of long-term sediment accumulation — can become a producing oil and gas field, several exacting requirements must be met.

1. Source rocks -
   These must be sedimentary rocks containing more than 1% of organic carbon (i.e. fossilised life tissues, animal and/or plant). In this case the coal measure sediments (Box 7.3 A) that filled the rift valleys between 80-40 myr ago are the main source rocks.

2. Cooking -
   The source rocks must be baked at between 100°C and 125°C for some millions of years. Given average crustal geothermal gradients of 20° to 25°C per kilometre of depth, cooking normally requires burial to depths of 4-6 km. In this case, burial was brought about by the change to crustal compression and rapid influx of sediment that began 25 myr ago, and was deep enough in the deeper parts of the basin.

3. Expulsion and Migration -
   Once created by the cooking process, oil and gas must be expelled from the source rocks. This normally accompanies the release of water which had been bound in the molecular structure of clay minerals. The oil and gas then migrate slowly through permeable or fractured rocks towards regions of lower pressure - ultimately the earth's surface. In the Taranaki Basin cooking and expulsion did not occur until around 10 myr ago.

4. Sealing and Trapping -
   To prevent loss of oil and gas to the surface, suitable seals and traps must be in place before migration takes place. Seals are layers of impermeable rock, and traps are geological structures which guide oil and gas into regions of lesser pressure from which there is no escape. The commonest trap is an arch-shaped structure (anticline) - see **D**, the Kupe structure.
   The Kora prospect in the northern Taranaki Basin was explored and drilled by ARCO in the late 1980's, and as mentioned in **C**, is a good illustration of the criticality of correct timing. In this case the 12 myr old Kora volcano was already extinct and in position adjacent to the 'kitchen' when expulsion and migration occurred about 10 myr ago. The highly permeable volcanic edifice provided a pathway for hydrocarbons leaving the 'kitchen'. Unfortunately, however, the top of the volcano still stood proud in the sea, and both oil and gas escaped into the sea. Burial and sealing of the trap (in this case a volcano) occurred a few million years too late.
   Another important factor is the nature of the structural traps. Ideally they should be broad and gentle, many kilometres across, allowing large volumes of oil and gas to accumulate, for example the Maui structures. The Taranaki Basin was just far enough from the subduction zone that most of the compressive arch structures are indeed broad and unbroken.
   Thus the Taranaki Basin by chance evolved with almost ideal circumstances for an oil and gas field. It is the only major basin in New Zealand with that ideal combination. However, most of the exploration wells drilled to date have been dry, which illustrates both the variation and complexity within the Taranaki Basin, and the inherently 'hit and miss' nature of oil and gas exploration.

For a discussion of the North Island's other potential oil and gas province, the East Coast forearc, see the introduction to Chapter 12.

The exposed sedimentary rocks of the Taranaki Basin

The basin's sedimentary rocks are clearly seen along the coast between Awakino and Waitara, accessible from SH3. There is typically a gentle dip to the south, so that progressively younger strata occupy the coastal zone towards the south.

Fig. 11.2. Thin-bedded sandstone and mudstones (about 10 myrs old) dip gently to the south (right) on the north Taranaki coastline, south of Tongaporutu.
Photographer Bruce Hayward.

Buried volcanoes and oil

The cliff south of Mokau River mouth exposes marine sandstones about 15 myr old, which contain some volcanic minerals and are darkish in colour. They are called Mohakatino Formation, and the dark volcanic grains were derived from an andesitic volcano – or perhaps more than one. This was one of a chain of old andesitic volcanoes that underlie the continental shelf and are now completely buried by marine sediments. The volcanoes were active between 15 myr and 8 myr ago, and formed part of the Colville–Taranaki Arc, as shown in stage 6 in Box 5.5 D. Extinct volcanoes are full of holes, and as such can make excellent oil and gas reservoirs provided there is a seal over them. Boxes 11.1 B, C give an interesting example of the place that old volcanoes can hold in oil exploration – provided they are sealed in good time.

Volcanic sandstones of the Mohakatino Formation are well known for their trace fossils – tracks and, particularly, burrows left by infaunal worms and other animals. Box 6.5 B illustrates some types of trace fossil, although there is a huge variety. Look also for features of the flysch type of sediment, also shown in Boxes 6.5 A, B.

Mt Messenger Sandstone

From Mokau south to Urenui, all coastal and roadside exposures are in the Mt Messenger Sandstone. This formation is typically gently dipping, stratified in beds 2 m or more thick, and made of fine- to medium-grained sand with a component of glistening mica grains. In places, the bedding is thinner and the mudstone layers are more prominent. All the sandstones were deposited on a moderately deep sea floor, by bottom-hugging sediment gravity flows such as turbidity currents and debris flows (Boxes 6.5 A, B). They formed a series of submarine fans, the underwater equivalent of alluvial fans, built at the mouths of submarine valleys cut into the continental slope. These valleys and fans were fed from the south. At this time, around 10 myr ago, Cook Strait did not exist, and the Mt Messenger sand was derived largely from erosion of the granites of northwest Nelson, hence its mica component.

The coastal cliff exposures can be accessed at many places, but you need to be there at low tide, and do remember that this coast is dangerous – it is easy to be cut off by the tide, and the cliffs are not readily climbable. Use Box 6.5 A as your guide to the strata. There is some variation in the sandstones from place to place. Look for old channels, measuring tens of metres deep and hundreds of metres wide, which were cut into sandstones already deposited, and then filled by coarser-grained shelly, pebbly sands. Look also for beds folded by slumping on the sea floor, and then covered by non-folded strata (Box 6.6 B).

Roadside exposures in the Mt Messenger Sandstone are common on SH3 between Tongaporutu and Waitoetoe. As the road climbs steeply over Mt Messenger, tunnels are cut through sandstones. However, the small scale of roadside exposures, compared with sea cliff exposures, makes it difficult to see the large-scale channel features.

From the parking area at the Mt Messenger summit there is a short walk along an old road, which gives access to some sandstone exposures, and also to views over wild and rugged inland Taranaki (Chapter 7). Much of the character of the view derives from the combination of rapid uplift (a few millimetres per year), causing deep, steep-sided valleys, and the gently dipping, thick sandstone beds that form cliffs. Most of what you are looking at to the east is the inland portion of the adjacent Whanganui Basin (Chapter 10).

Taranaki Volcano

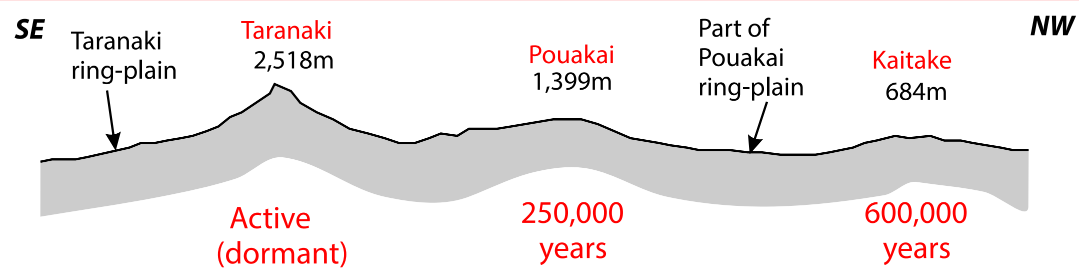
Taranaki region is dominated visually (on a clear day) by Mt Taranaki and its ring plain. Taranaki is an andesitic stratocone volcano (Boxes 9.2 A–C) from the same stable as Mts Ruapehu and Tongariro in the Taupo Volcanic Zone (TVZ; Chapter 9). However, it is different in two ways.

First, it occupies a unique 'behind-arc' position in New Zealand, being our only subduction-related andesitic volcano so far from the Hikurangi subduction trench (300 km west of it). By definition, therefore, since the subducting slab slopes steeply towards the west, it is the only slab-sourced volcano in New Zealand located above the deeper part of the subducting slab of Pacific Plate lithosphere (more than 200 km deep). The 'normal' volcanoes of the TVZ are typical of active arcs in general, being situated 90–100 km above the subducting slab and therefore lying closer to the trench. They are arc volcanoes. Mt Taranaki, in contrast, is a behind-arc, or back-arc, volcano.

Second, Taranaki lava differs from typical volcanic arc andesitic lava (e.g. that of Mt Ruapehu) by containing higher levels of potassium. Aside from these two differences, the lava and the volcanoes are very similar.

Mt Taranaki is younger than 250,000 years and has ancestors. They include two immediate precursors, Pouakai and Kaitake, which are lower, eroded cones located to the northwest of Taranaki. The three cones lie on a straight line trending from northwest to southeast – the Taranaki Volcanic Lineament. Kaitake is the oldest at 600,000 years, while Pouakai lies between it and Taranaki in age at 360,000–250,000 years (Fig. 11.1 and Box 11.2). The two older cones are not visible from the eastern and southern sides of Mt Taranaki's ring plain.

Box 11.2. Three Taranaki volcanoes – same type, similar size originally, same climate.



Very little of the Kaitake ring plain is preserved. Height reduces by approximately one half every 300,000 yrs once volcanic activity ceases. An initial lowering rate of more than 4 mm per year reduces to about 2 mm per year.

CONCLUSION — Rates of weathering, soil formation and erosion, and tectonic change, throughout the North Island, are all measured in millimetres per year, and are fast enough to cause big changes to the landscape in geologically short times.

Volcano migration

The lineament shows that the Taranaki volcano is marching southeastwards, at an average rate of 30 mm/year (18 km in 600,000 years). The TVZ is also migrating to the southeast at a similar rate (20–30 mm/year). Because both sets of volcanoes are tied to processes taking place in the subducting slab of Pacific Plate lithosphere, at particular depths (Box 2.2), their migration tells us that the slab is moving. The process causing migration is trench rollback, as described in Chapter 2. Rollback chains of volcanoes similar to the Taranaki Lineament are visible in Fig. 1.3 behind the Kermadec volcanic ridge. The rate of rollback determines the rate of advance of the lineaments.

The Taranaki ring plain

Volcanoes that erupt andesitic lava are explosive. As explained in Chapter 9, they typically have extensive ring plains that are built by lahar flows and debris avalanches originating from the central cone. The ring plains are much larger in area and volume than the central cone, which is made of lava flows interlayered with fragmentals (Fig. 11.3). Mt Taranaki has a particularly extensive ring plain, a fact that is directly associated with the slender, pointed profile of the mountain. That profile is determined more by frequent collapse events than it is by construction during eruptions. For comparison, Mt Ruapehu has more lava flows, fewer collapse events and a much broader, blocky profile.

Thus, most of peninsula Taranaki consists of the ring plains of Mt Taranaki and its two predecessors. They form a relatively thin cover, hiding the sedimentary rocks of the Taranaki Basin underneath. Most of the land-based petroleum drilling in Taranaki starts by drilling through the ring plain.

Fig. 11.3. Mt Taranaki's extensive ring plain, deposited by passing lahars, spreads out southwards from the foot of the cone all the way to the coast near Hawera. Photographer Lloyd Homer, GNS Science.

Mound fields – collapses of sectors of the cone

The Taranaki ring plain is famous for its mound fields. A typical mound is 5–10 m high and 50 m or so across (Fig. 11.4). Each mound is the weathered remains of a chunk of the volcano, and a mound field marks the deposit of a catastrophic debris avalanche event, caused by the collapse of a sector of the volcanic cone. A recent example occurred in Washington state, USA, in May 1980, when the whole northern flank of Mt St Helens collapsed. Taranaki is particularly prone to such sector collapses, as shown by the large number of mound fields all around the ring plain, and by the volcano's slender conical shape.

Lahar and debris avalanche deposits

The best place to see debris flow (lahar) and debris avalanche deposits is in the region's coastal cliffs, which can be accessed at many places (see below). Most of the material exposed will be bouldery gravels (technically, a boulder is more than 256 mm across). Virtually all the boulders and smaller gravel pieces are andesite lava, although there are rare blocks of fossiliferous sandstone brought up from the deposits of the Taranaki Basin that underlie the volcanoes. Pieces of wood are also present. Note the degree of rounding acquired by the lava blocks during weathering on the mountain, and during transport in a fast-moving debris flow.

Taranaki volcanoes are generously supplied with rainfall, and act as a giant sponge of water, and therefore lahars and debris avalanches here are typically very wet – i.e. they are well lubricated. In some parts of the world they are drier and less well lubricated, and so do not travel as far.

Fig. 11.4. These 5–10 m high mounds on the western ring plain of Mt Taranaki were deposited by a catastrophic debris avalanche, caused by the collapse of a sector of the volcanic cone. Photographer Lloyd Homer, GNS Science.

Fig. 11.5. This coastal cliff near Oakura, south of New Plmouth, is composed of bouldery gravel of andesite lava that was deposited by a fast-moving debris avalanche that flowed down from Pouakai Volcano's cone. Photographer V.E. Neall.

When looking at a cliff exposure, try to distinguish individual depositional events. Relationships can be quite chaotic where flows have carried large chunks of the volcano, or have ploughed into earlier deposits, ripping chunks off them. Layers can dip quite steeply as a result of rip-up and ploughing relationships, and their thickness can range from less than 1 m to more than 10 m. Layers of bouldery lahar material may be separated by fine-grained layers of airfall volcanic ash.

Boulders can range up to 30 m in size, but note that a boulder of that size will be a chunk of the collapsing volcanic cone, and therefore may itself be made either of jointed lava or of bouldery material. It may also form a mound projecting above the general upper level of the deposit.

The proportion of boulders to sandy matrix varies greatly, depending on exactly where in the lahar flow or debris avalanche the deposit is situated. Debris flows and avalanches sort the material they are carrying, and a lobe-shaped lahar deposit 1 m thick and 50 m across can end up, for example, with an edging of larger boulders (known as a levée) and a central zone of smaller boulders and more sand.

By contrast, the bigger and more energetic debris avalanches on the Taranaki ring plain can spread across several kilometres. They tend to concentrate the volcanic chunks in the mid-region of the deposit (where each chunk forms a mound), while the finer matrix (sand and mud) and smaller boulders are concentrated around the edges of the deposit.

The coastline from New Plymouth to Kaihihi Stream, a distance of about 25 km, is made entirely of older ring plain materials derived from the Pouakai Volcano. They are easily accessed around Oakura River mouth, where cliff exposures show chaotic debris avalanche deposits. Interestingly, radiometric dating of lava boulders here includes ages of 3.6 myr and 0.6 myr. The older date does not tie to a known volcano in Taranaki, and so may indicate a previously unknown part of the region's volcanic history. The younger rock is presumably derived from Kaitake Volcano.

Further along SH45, past Okato, the volcanic deposits are from Mt Taranaki and have therefore travelled further. There is an extensive mound field around Pungarehu, where the Cape Egmont lighthouse is located on a mound.

Shore platforms here may be armoured by a layer of boulders eroded out of the underlying lahar deposits. Beach sands around Taranaki tend to be dominated by black ironsand derived from the lava materials. The black mineral in the sand is titanomagnetite, a magnetic iron ore containing quite high levels of titanium that originates as an accessory mineral (less than 1% by volume) in Taranaki lavas. Ironsand was mined for a time at Waverley, near Patea, from old dune deposits, and exported to Japan. Waves push the ironsand, and other dark volcanic minerals such as pyroxene, a long way eastwards, beyond Whanganui beach.

Airfall ash deposits and loess

Box 9.4, describes airfall ash deposits, while Box 15.1, describes associated dust deposits (loess, pronounced 'lerss'). Because of the prevailing westerly winds, ash and loess layers are seen mostly on the eastern side of Taranaki's ring plain. Loess deposition is associated particularly with glacial periods, when vegetation was sparse and winds strong.

Geology of New Plymouth

New Plymouth is located on the ring plains of Kaitake and Pouakai. These older cones and their ring plains have the effect of shielding the town from lahar and debris avalanche events on the currently active ring plain of Mt Taranaki.

In New Plymouth itself, Paritutu and the Sugar Loaf Islands are remnants of even older (around 2 myr) andesitic volcanoes of the Taranaki Volcanic Lineament. They are constructed of volcanic breccias.

Andesitic volcanic rocks weather to form a fertile soil. In association with high rainfall generated by Mt Taranaki, the region's fertile soil has contributed to its flourishing dairy industry. It also supports a vigorous rainforest cover in the Egmont National Park, as well as some nationally renowned gardens like the rhododendron garden at Pukeiti.

Other Behind-arc Volcanoes

While Mt Taranaki is unique in being the only New Zealand andesitic volcano in a behind-arc position, there are other behind-arc volcanoes north of Taranaki, as noted in earlier chapters. One type comprises the Auckland volcanic field, of basaltic composition (Chapter 6); other similar fields (10 of them altogether) formed in the last 11 myr (Box 6.8). The origin of these fields is quite different from the andesitic lavas, as the basaltic lava forms in the wedge of mantle rocks, below the crust and above the subducting slab. In the case of Auckland, the slab is more than 300 km deep beneath the city.

Chapter 12

East Cape to Wellington

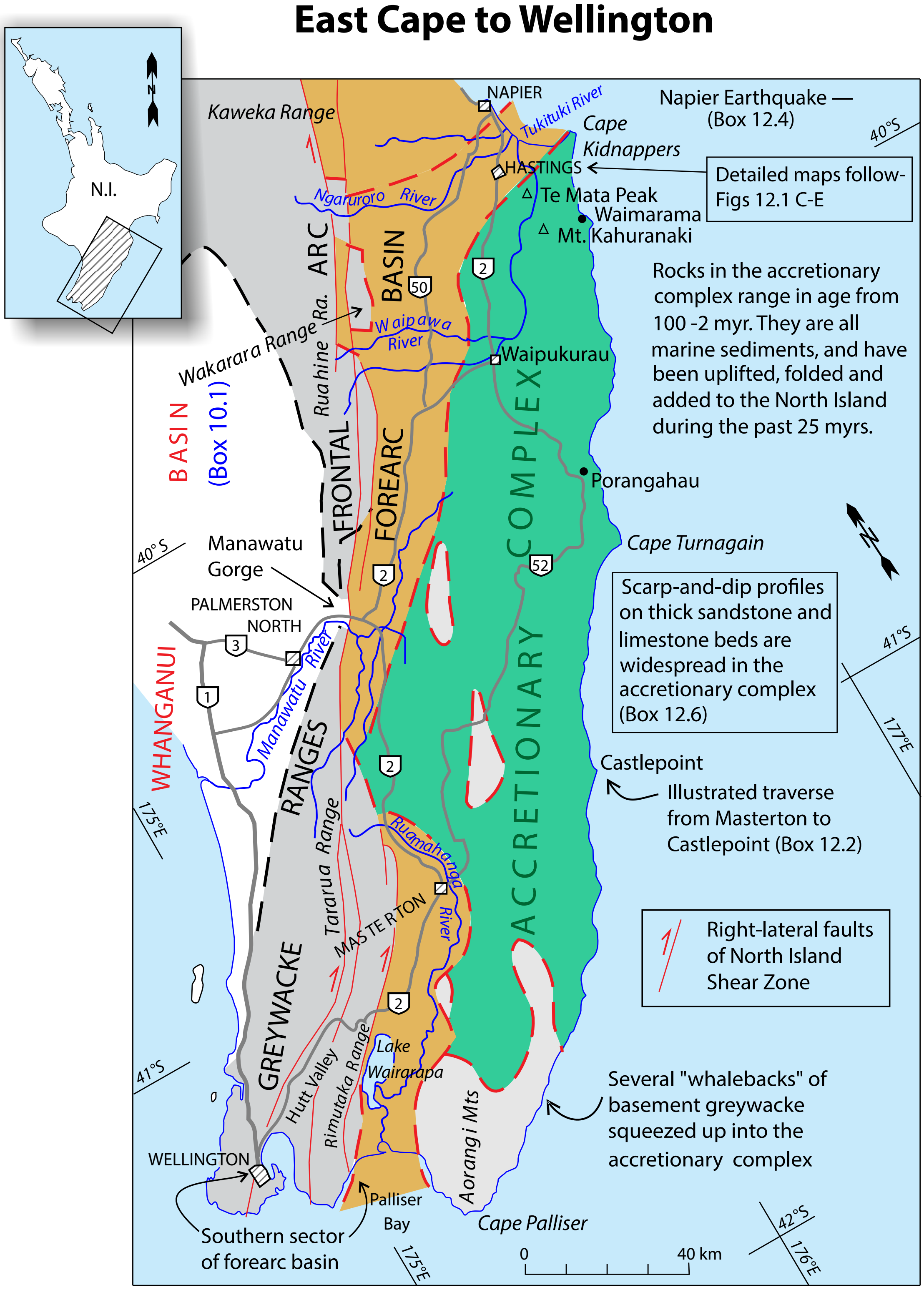


Fig. 12.1 A. Simplified geology of the southern east coast region from Napier to Wellington.

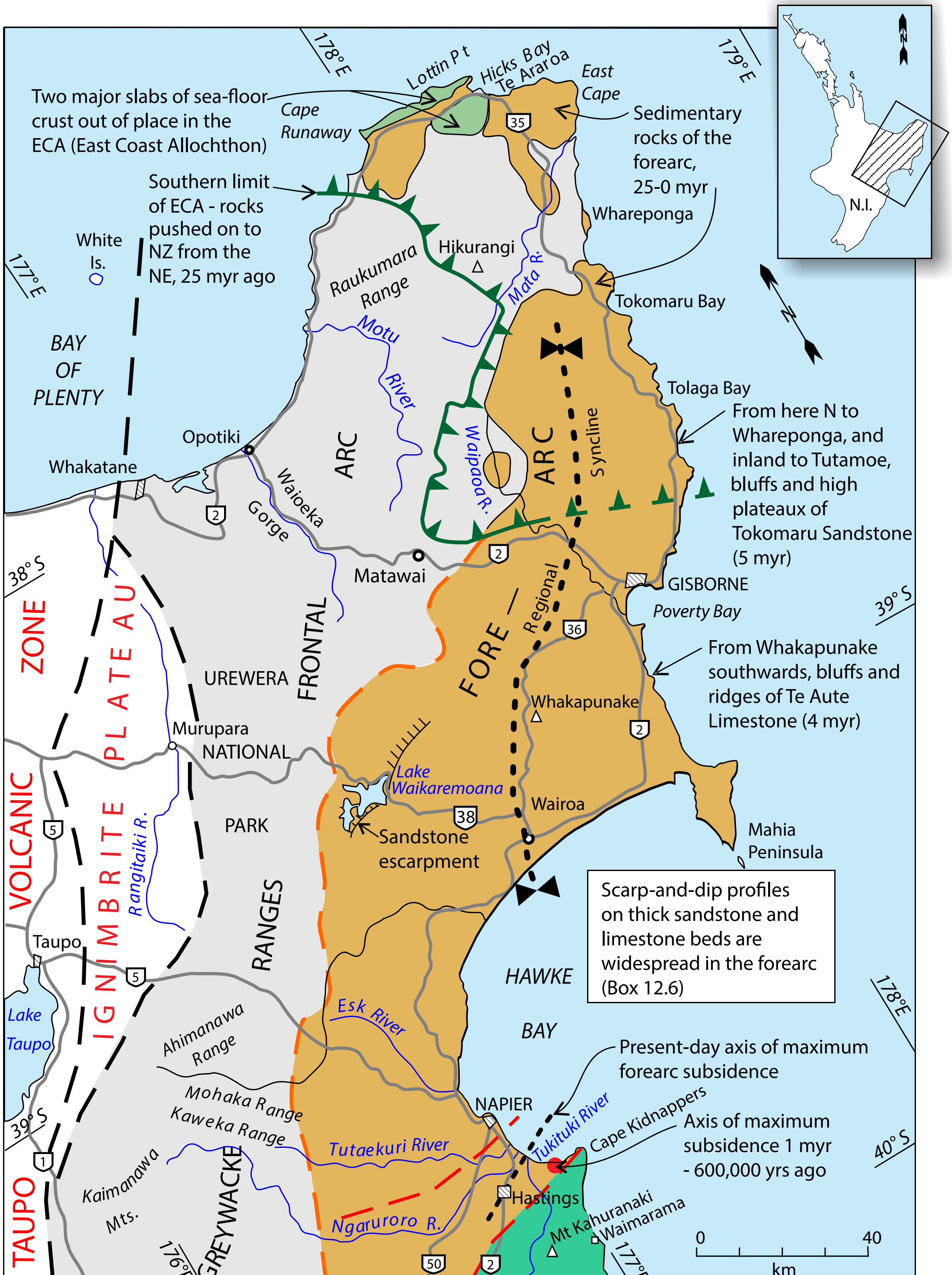


Fig. 12.1 B. Simplified geology of the northern east coast region from East Cape to Napier.

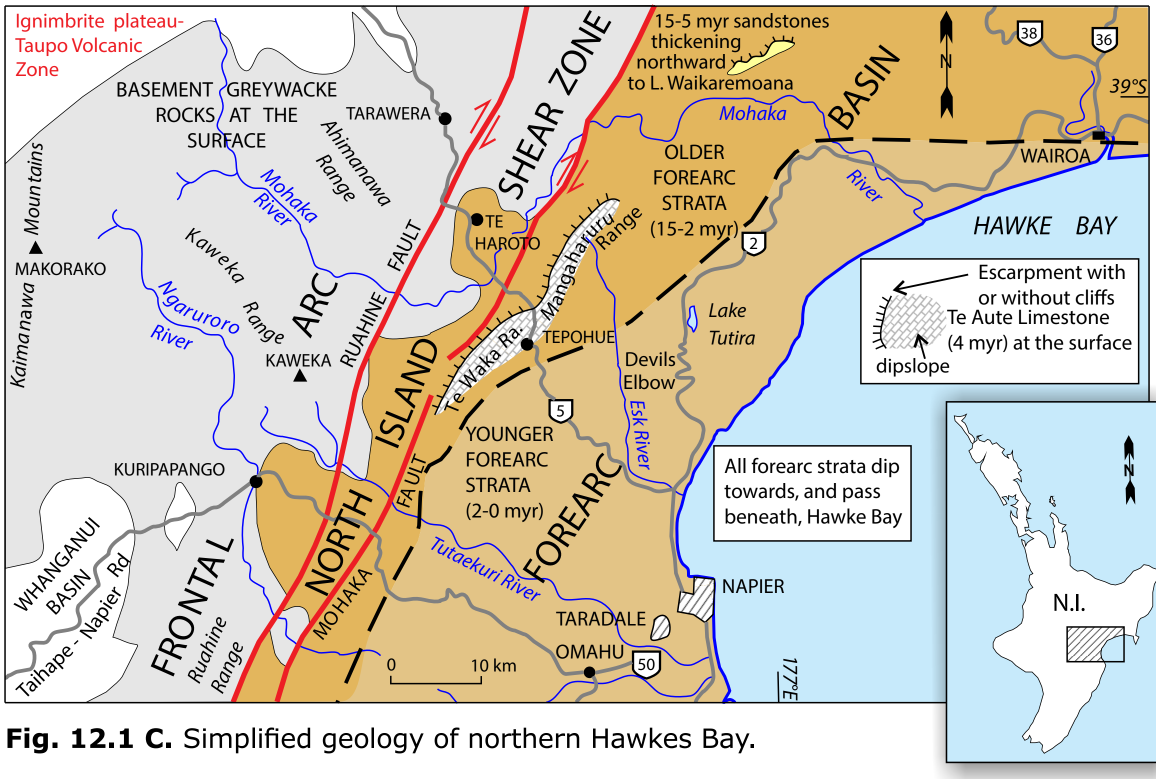


Fig. 12.1 C. Simplified geology of northern Hawkes Bay.

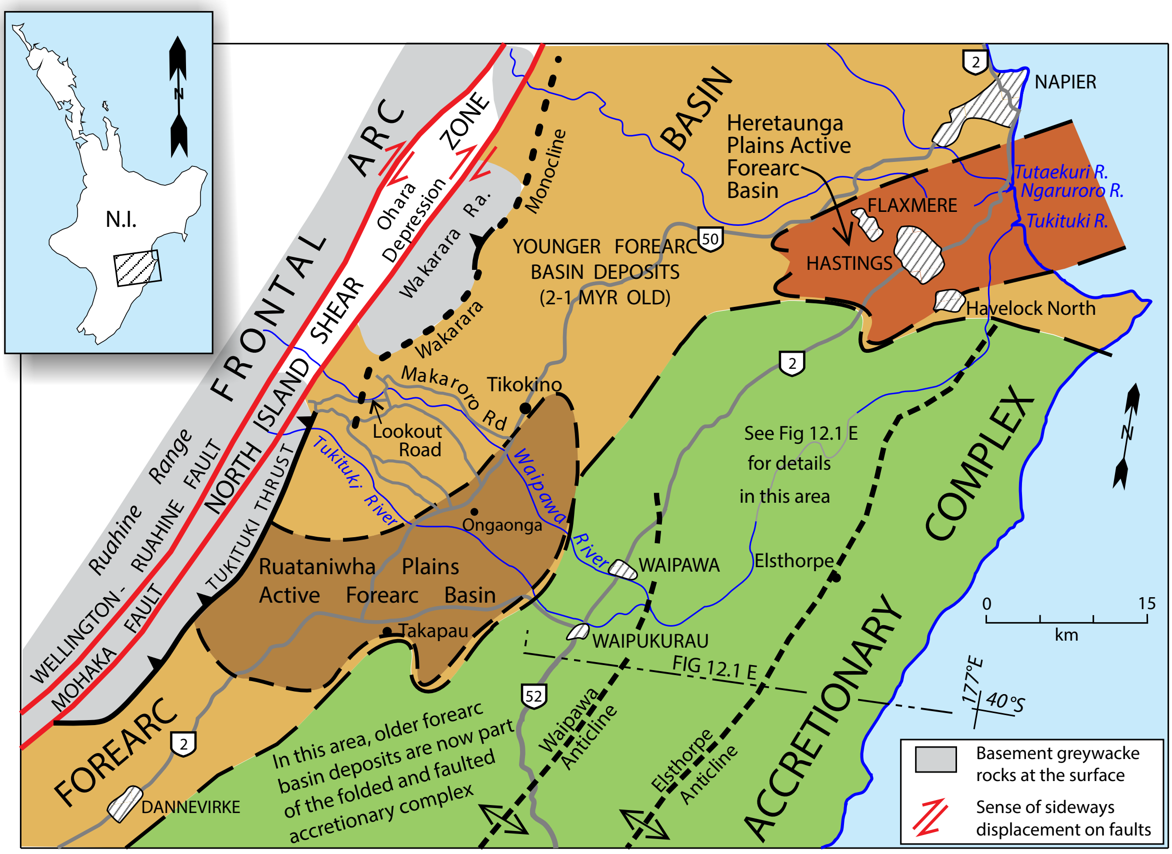


Fig. 12.1 D. Simplified geology of southern Hawkes Bay.

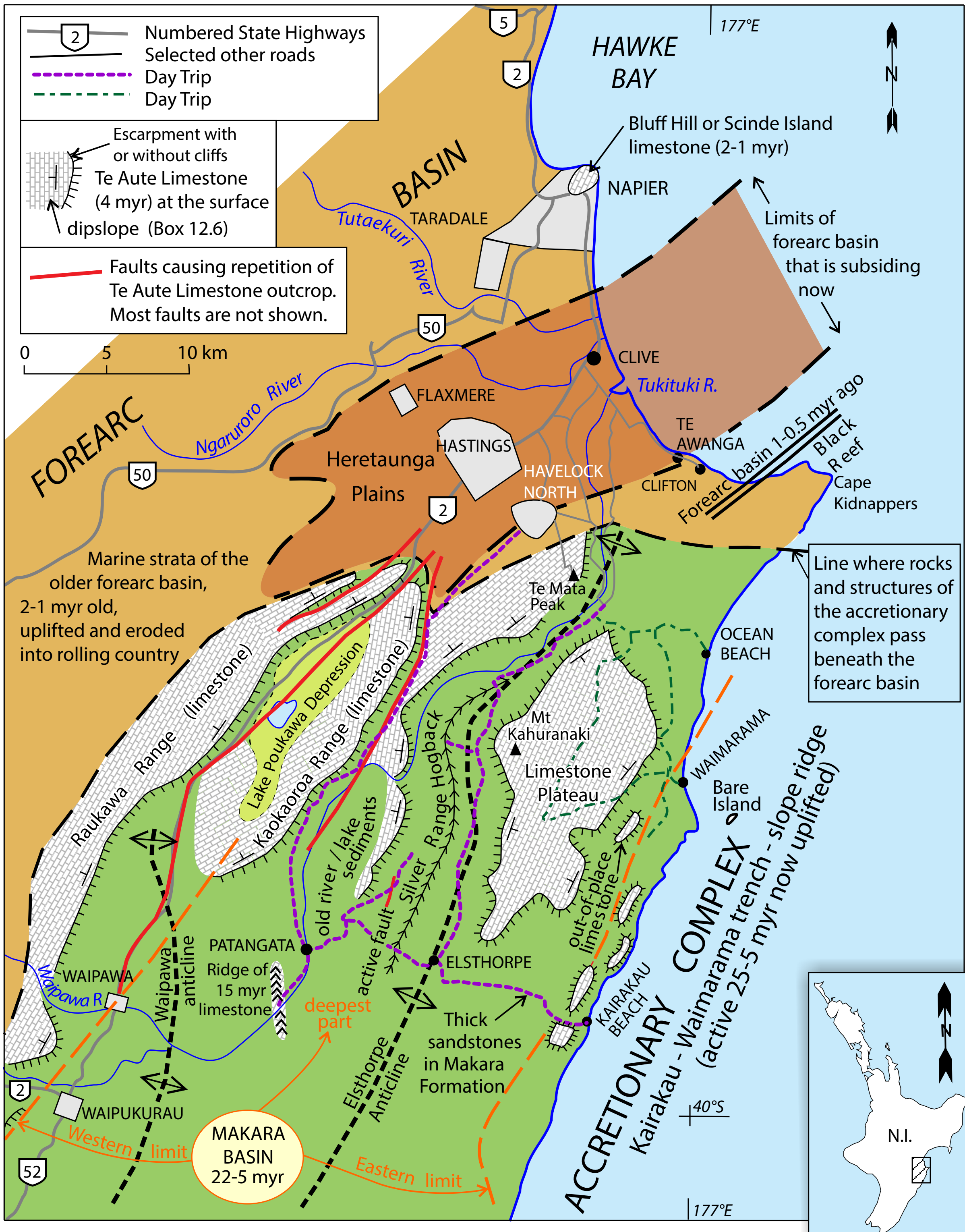


Fig. 12.1 E. Simplified geology map illustrating the relationship of the forearc basin strata and the accretionary complex, southern Hawkes Bay.

The Forearc of the North Island Convergent Plate

A strong geological theme runs through the whole of eastern North Island, from East Cape to Cook Strait and including the axial mountain ranges. The region is next door to the subduction trench along the Pacific–Australian plate boundary, and is where we see most of the direct effects of interaction between these plates. In geological terms, it is the forearc of the North Island convergent plate margin. Boxes 2.1, 2.2 and 2.3 show the plate layout and a generalised interpretation of what happens as the Pacific Plate passes underneath the Australian Plate at our part of it (i.e. the North Island).

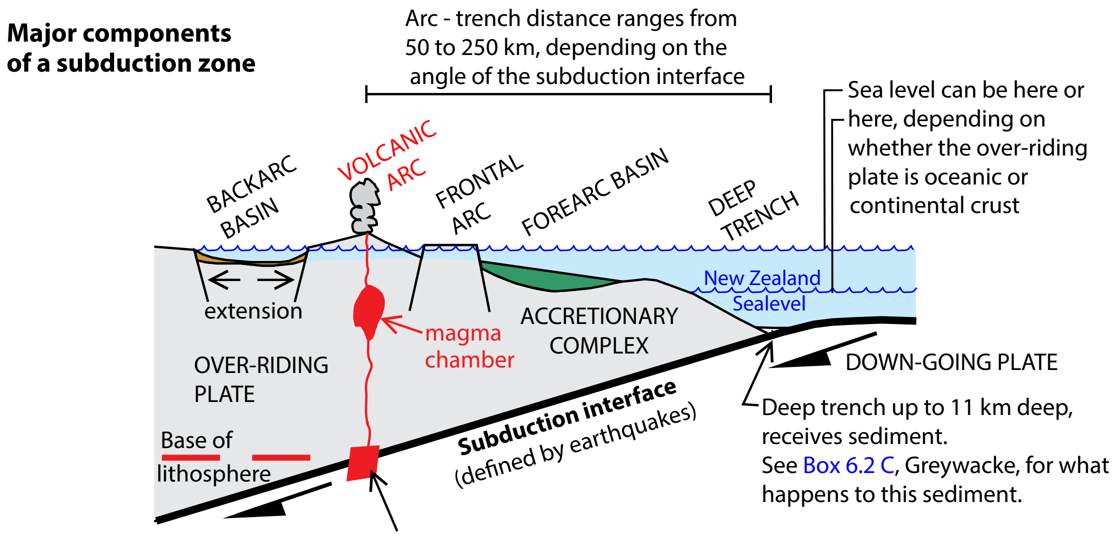
A forearc is typically 100–200 km wide. It includes everything between the active volcanic arc (in this case the Taupo Volcanic Zone, or TVZ, in central North Island – Chapter 9) and the deep-sea subduction trench (in this case the Hikurangi Trough). The forearc is divided into three parallel parts: a frontal arc adjoining the volcanic arc (this comprises the North Island axial mountains); a forearc sedimentary basin next to the frontal arc (Hawke Bay is part of this); and an accretionary complex between the forearc basin and the subduction trench. This last part is so called because it is accreting (adding) new rocks to the North Island, and after being squeezed, the rocks are complex. Figs 12.1 A–C explain these three parts.

Eastern North Island has been partitioned in this way for the past 25 myr. Many of the rocks involved are actually much older than that, and have simply been incorporated into the forearc by accident, as it were. However, all rocks younger than 25 myr have formed in the geological environment created by the forearc, and as such contain the historical record of the forearc. Box 12.2 shows a transect across almost the entire forearc, from Masterton to Castlepoint.

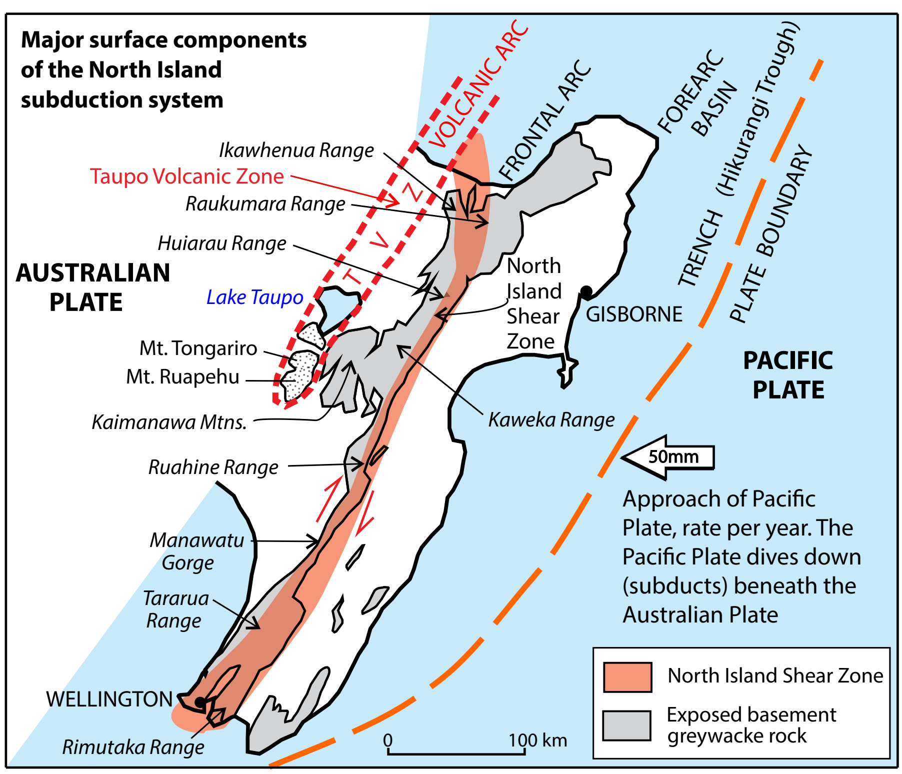
Fig. 12.2. Oblique view north over the onland portion of the North Island between the active Taupo Volcanic Zone arc volcanoes to the west (left) and the Hikurangi Trough subduction trench offshore to the east (right). Photograph courtesy of Google Earth.

Box 12.1 A. Eastern North Island: the surface expression of a subduction zone.

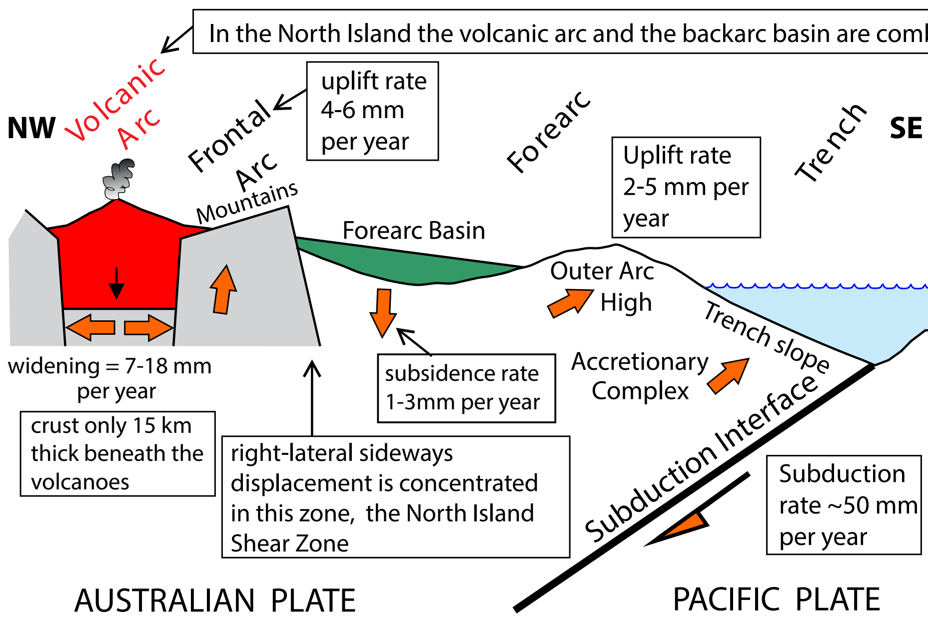
A subduction zone occurs where two tectonic plates converge, and one passes beneath the other. All subduction zones give rise to a characteristic series of tectonic features:



Magma generation typically occurs where the subduction interface is at a depth of about 100 km. Magma melts and stopes its way to the surface.
How it works in the North Island —



Box 12.1 B. Eastern North Island: schematic cross section.



This schematic cross section shows some of the movement directions and rates in the near-surface rocks above the subduction zone. Those not shown are those going into and out of the page. These arise because the Pacific Plate approaches the North Island obliquely rather than directly — the angle between the 50 mm/year arrow and the line of the mountains is not a right angle.

The angle of the approach drives the forearc southwards (out of the page) relative to the frontal arc, causing a right-lateral sideways displacement between them. (Right lateral = the opposite side of the fault moves to the right). This displacement is concentrated on a few major faults, which tend to form the eastern boundary of the frontal arc. They are known as the **North Island Shear Zone**. Right-lateral displacement becomes more pronounced in southern North Island, because the oblique component of movement increases from north to south (Box 2.3). During the large 1855 Wellington earthquake, rupture on the Wairarapa Fault near Cook Strait amounted to uplift of about 3 metres on the west side, (the mountainous side) and right-lateral sideways movement (i.e. west side moving northward relative to the eastern side) of about 18 metres. See Chapter 13 .

The **Volcanic Arc** in the North Island is the Taupo Volcanic Zone (TVZ). It is described in Chapter 6.

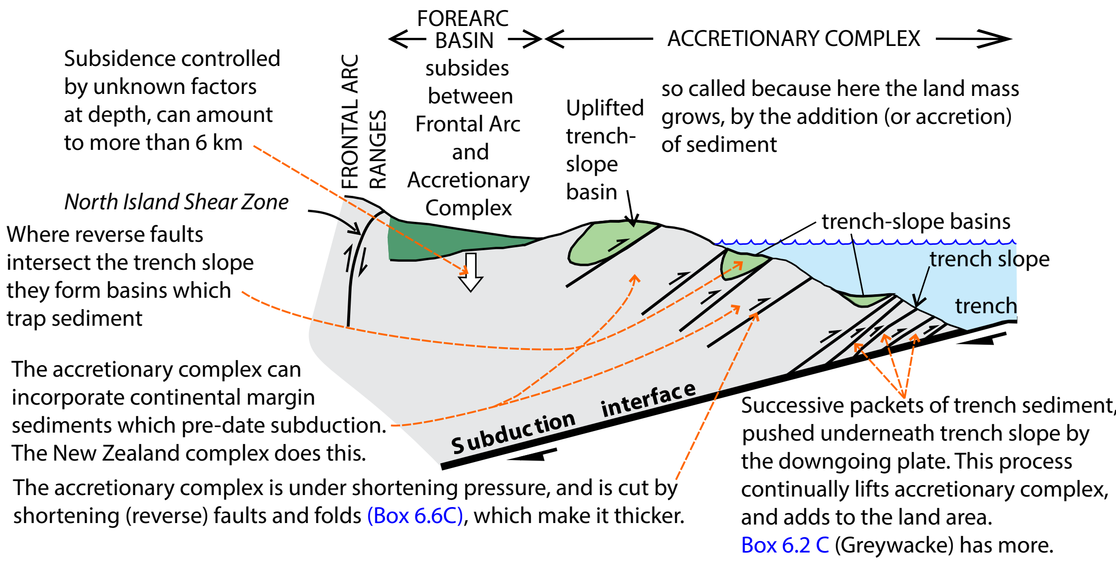
The **Frontal Arc** in the North Island comprises the axial mountain ranges, which extend from Cook Strait to East Cape. The ranges are made of uplifted basement greywacke rock. Uplift has taken place only within the last few million years. Uplift is caused by compression created by westwards movement of the Pacific Plate. It became stronger following a change in the movement pattern of the Pacific Plate 5 myr ago.

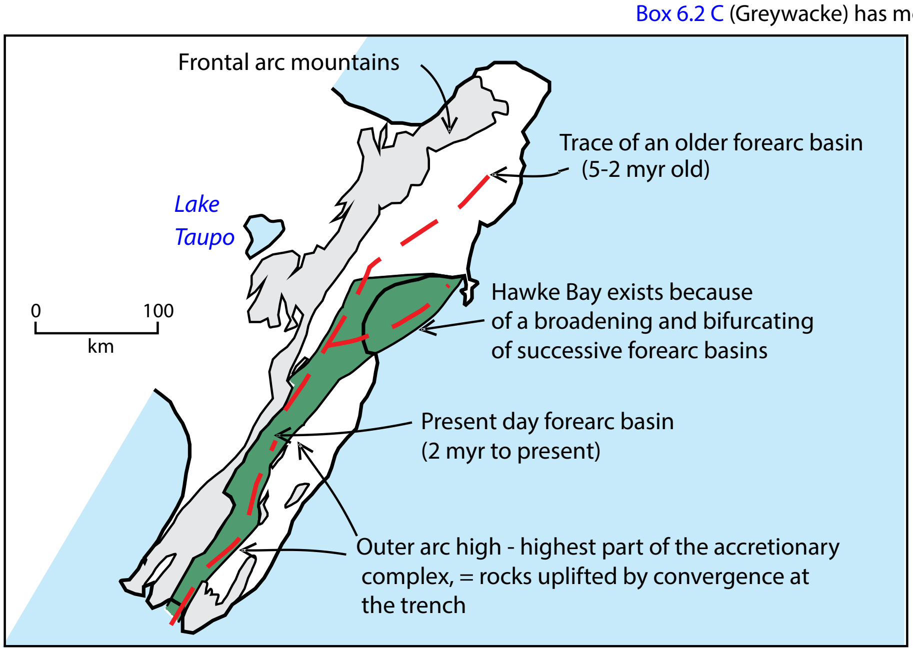
The axial mountains are sliced up by actively moving faults (Box 13.3 B) of the North Island Shear Zone. The faults help to delineate the named ranges, each of which, along with various minor ranges, is basically a large fault-bounded block of greywacke rock.

Despite the prevalence of sideways fault displacement, there has been doubling-up of mountain ranges only in two places: in central North Island adjacent to Mt Ruapehu, where the Kaimanawa and Kaweka Ranges lie side by side; and south of Bay of Plenty, where the North Island Shear Zone bends to the north and leaves the axial mountains, stacking up several minor north-trending ranges between the Ikawhenua Range and the Raukumara Range.

Box 12.1 C. Eastern North Island forearc region of the New Zealand subduction zone.

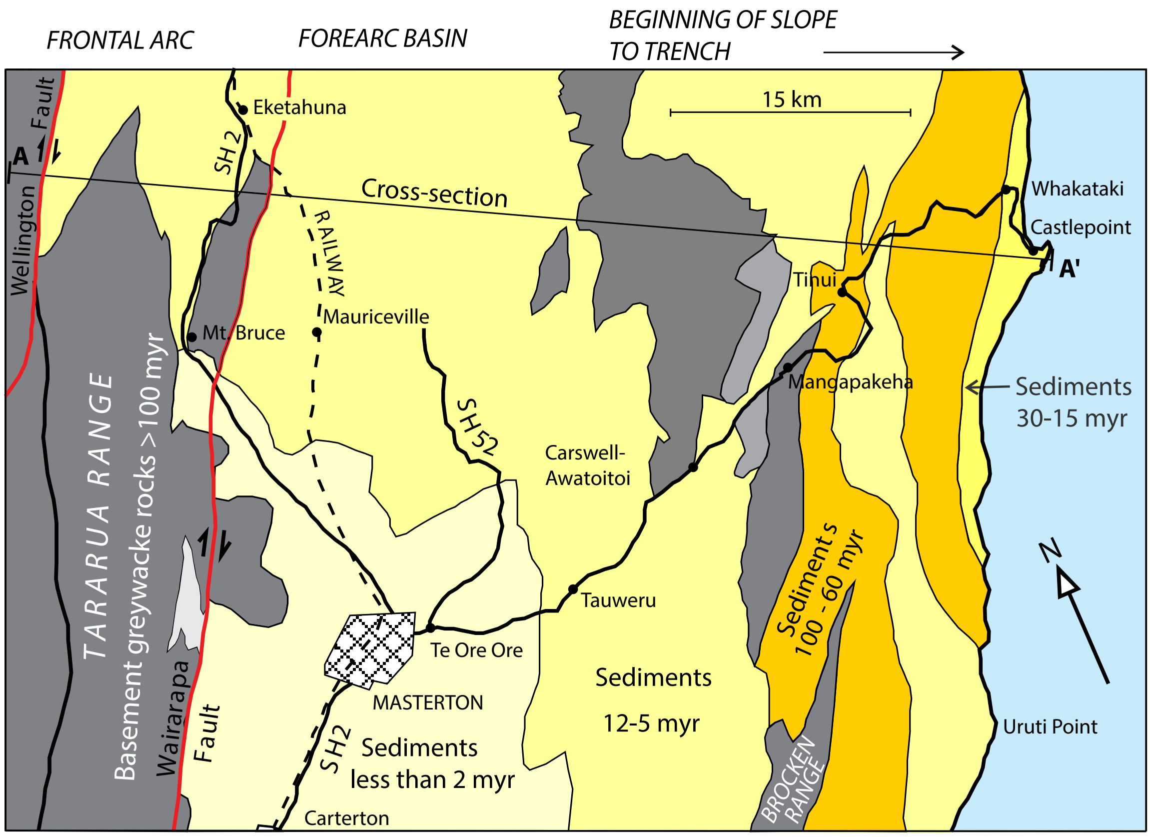
The entire eastern region of the North Island, between the mountains and the sea, and then beneath the sea to the deep trench off the coast, constitutes the **Forearc region** of the New Zealand subduction zone.
The Forearc region has two major components - **forearc basin**, and **accretionary complex**.

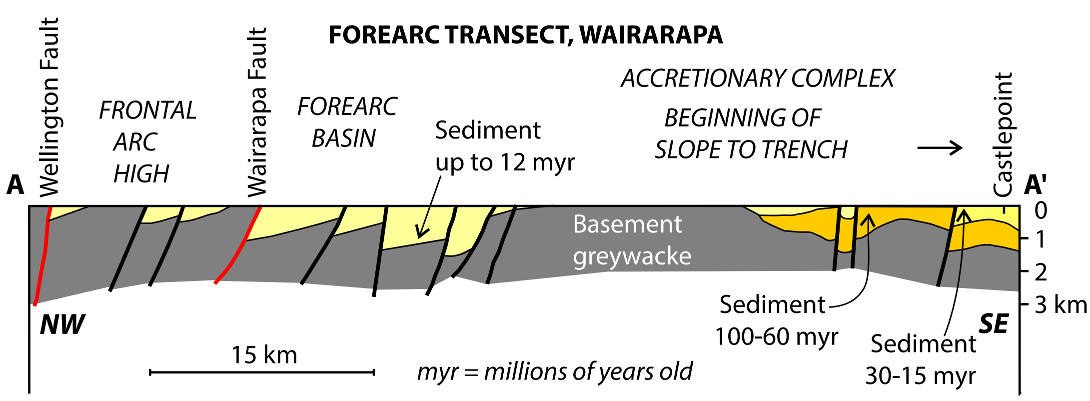




There is more information on all sectors of the Frontal Arc and Forearc in Chapter 13.
The New Zealand subduction zone has been active for 45 myr. In one way or another it has either determined or influenced just about every topographic and geologic feature that we see in the North Island.

Box 12.2. Eastern North Island forearc region of the New Zealand subduction zone.





Note that there are many faults - more than shown here. Most of them are reverse (i.e. shortening) faults; some, e.g. Wellington Fault, have a component of right-lateral movement as well. Only the Wellington and Wairarapa Faults are shown on the map.

The Plate Tectonic Vice

Convergence between the plates is taking place along the East Coast at a rate of about 50 mm/year. It generates huge compressive forces, like tightening a vice. However, this vice is different from a carpenter's vice – the guides bringing in the outer (Pacific) plate are not at right angles to the inner (Australian) plate. Instead, they are at an angle of about 50° (as indicated by the thick red arrows on Box 2.3). Consequently, rocks in the forearc zone are pushed to the left as well as being squeezed as the Pacific Plate approaches the Australian Plate. This is known as oblique plate convergence, and it becomes progressively more pronounced to the south, so that at Cook Strait the plate boundary changes from being predominantly compressive to predominantly sideways.

North Island Shear Zone

As it works out, the sideways displacement between the Pacific and Australian plates is not distributed uniformly across the boundary zone. Instead, it is concentrated inboard, along a narrow strip called the North Island Shear Zone, which is the site of several major active faults (Fig. 12.1 A and Box 12.1 C). The shear zone is partly located within the frontal arc mountains, and partly forms the boundary between the mountains and the forearc basin. The faults, which include the Wellington Fault, have components of both up and down and sideways displacement (Box 6.6 C, D), the latter right lateral (the opposite side of the fault moves to the right). Being in a zone undergoing compression, the up and down movement on some of the faults is reverse (shortening).

Tectonic squeezing

Squeezing, though, is distributed uniformly across the boundary zone. It normally translates into folding of strata, thickening of the rock stack, and hence uplift of the crust. In the forearc as a whole there is one exception to that, where one strip of crust – the forearc basin – is subsiding. This is because of things happening at depth as the lithosphere of the subducting plate dives deeper into the earth's interior and is subjected to increasing temperatures and pressures. The process is not fully understood, but may have to do with the compaction of sediments contained in rift valleys on the downgoing plate surface (Box 2.2). This is another difference between a carpenter's vice and the plate boundary – the guides on which the Pacific Plate rides are infinitely long, and the plate passes underneath the Australian Plate (equivalent to the back plate of the vice) and dives deep into the earth's interior. It has been doing this for 45 myr, although there were few recognisable effects for the first 20 myr.

Earthquakes

Active subsidence and uplift, and active fault movement, both in the North Island Shear Zone and generally throughout the forearc, mainly occurs in jumps rather than gradually, as built-up stress is released periodically. Each jump is associated with an earthquake, so here these phenomena are frequent – in geological terms, that is. Any one area of a few thousand square kilometres can expect a damaging earthquake (7 or higher on the Richter scale) every few hundred years (also Boxes 13.2 A–C and 13.3 A, B).

Recent Fault Traces

The forearc region, as you will have gathered, contains many faults. Recent movement on a fault can be shown as a displacement of the land surface, making a step a few metres high that runs in a straight line and cuts and displaces all other landforms (Box 13.3 A). Note that the fault plane itself may not be vertical. Fault traces are easy to see on aerial photographs, which is why they appear on maps, but as it happens they are much less easy to recognise on the ground. This is because there are so many other kinds of step in the landscape with which they can be confused, like the edge of a river terrace.

To be sure you are looking at a recent fault trace, you need to be on foot and satisfied that the step is displacing all other landforms. If the downthrown side of the fault is opposite to the downwards slope of the land, streams may pond behind the fault trace.

The Frontal Arc – the Axial Mountain Ranges

Geologically, North Island mountains are only distantly connected to the South Island mountains, and they are distinctly younger. We are talking here about the mountain landforms, not the rocks that make up the mountains – these are the old basement greywacke rocks that are pretty much the same the length and breadth of the country (Chapter 5). The best evidence for growth of mountains is gravel in the surrounding basins. North Island mountains have been shedding gravel for no more than 1 myr.

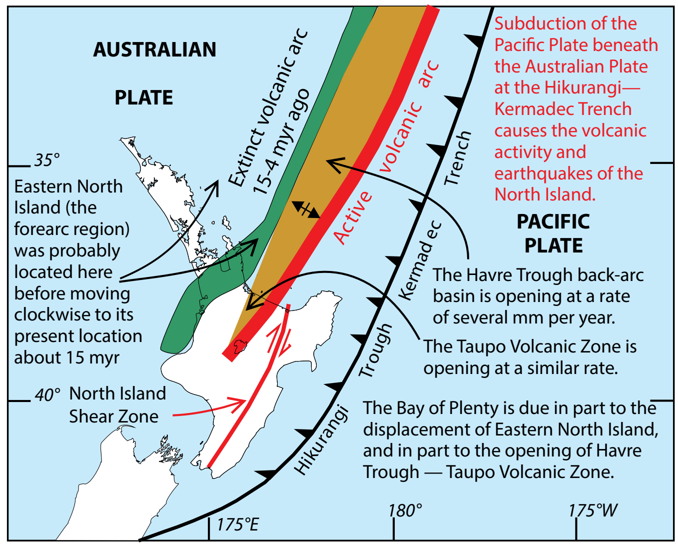
The basic reason for the growth of mountains in both islands was the same: a slight change in the direction of approach of the Pacific Plate about 5 myr ago, which increased the compression between the two plates (Box 5.5 D). South Island mountains started growing more or less immediately, because there the compression is between adjacent slabs of rigid continental crust. In the North Island, the compression is between oceanic crust and continental crust, mediated through the accretionary complex, and as a result there was a long delay in the response – hence the younger age of the mountains here.

The North Island Shear Zone maintains a constant southwest–northeast heading until, north of Hawke Bay, it splays into several branch faults that separate a number of smaller ranges, and bends to a northerly trend. And then, in the vicinity of Whakatane and Opotiki on the Bay of Plenty coast, just where the shear zone faults begin to intersect the volcanic arc, all the ranges dive down under the sea. Only the slice of greywacke on which Whakatane is located actually reaches the coast, which leads us on to the next section.

Why is the Bay of Plenty so wide?

We have a good explanation for the formation of the central one-third of the Bay of Plenty. As discussed in Chapter 9, it is the offshore extension of the TVZ, forming a link between the TVZ and its offshore equivalent, the Havre Trough back-arc basin. Both of these are widening by several millimetres per year, and the central Bay of Plenty itself widened by 1 m during the 1987 Edgecumbe earthquake (Fig. 12.3). Similarly, the western third of the Bay of Plenty is explained by the Tauranga Basin and its onshore extension (Chapter 8).

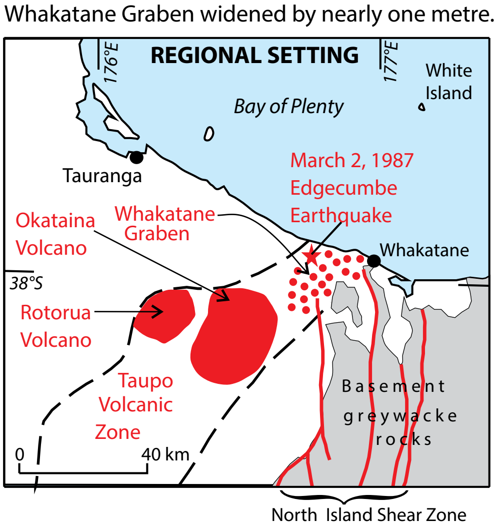
Box 12.3. Origin of the Bay of Plenty.

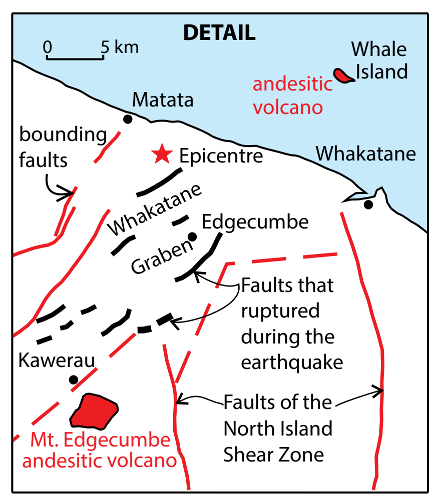


The North Island is 'unzipping' along the line of the Havre Trough-central Bay of Plenty-Taupo Vocanic Zone (TVZ), to as far south as Mt Ruapehu. This is effectively the extension of the backarc basin (Havre Trough) into the North Island. It and the accompanying volcanoes have been extending to the southwest (southern termination of the TVZ) and to the southeast (eastern limit of TVZ) over the past 2 million years. This process will presumably continue. The southeastward movement is in response to trench rollback. The centre of the Bay of Plenty is thus a backarc basin, However, the bay is much wider than that, so how does the rest of it originate ? The western part is the expression of the Tauranga Basin, which is possibly an older part of the Havre Trough. The eastern part is less easily explained.

THE MARCH 2, 1987, EDGECUMBE, MAGNITUDE 6.3 EARTHQUAKE

The Edgecumbe Earthquake of March 2, 1987, magnitude 6.3, gave a perfect illustration of the way in which the Taupo Volcanic Zone — Bay of Plenty — Havre Trough are opening. Although of a moderate magnitude, the earthquake caused extensive damage because the origin was shallow. During it the Whakatane Graben widened by nearly one metre.





Whakatane Graben - an elongate depression filled with up to 2 km thickness of volcanic rocks, marine sediments and alluvium.

Subsidence rate - estimated at up to 2-3 mm per year (long-term average rate). The shoulders of the depression are rising at about 1mm per year.

Faults that ruptured the ground surface during the earthquake are shown. All had normal separation (Boxes 6.6 C, D), mostly down to the NW, without any sideways displacement.

Long-term (average) rate of widening - about 7 mm per year.

What needs explaining is the eastern one-third of the curve of the Bay of Plenty, and why it truncates the frontal arc of greywacke rocks. It is not known what rocks underlie the seabed between the TVZ at White Island and the Raukumara Peninsula. Nor do we know why the Raukumara Range of greywackes drops so steeply into the Bay of Plenty. At the moment we simply do not have a good explanation for this section of the Bay of Plenty.

Greywacke slice landforms

The rocks of the greywacke association can be examined at many places along the frontal arc between the Raukumara Peninsula and Wellington (e.g. there are exposures in Whakatane town), and the northwards nosedive taken by several greywacke slices at the Bay of Plenty has given rise to several coastal fault-angle depressions here. These are occupied by Ohiwa Harbour, by the coastal flats to either side of Opotiki, and by flat-bottomed, sediment-filled, fault-angle valleys that extend up to 30 km inland along the fault traces, such as those occupied by the Whakatane River, and Waimana River and Nukuhou Stream.

The 1931 Napier Earthquake

Although usually called the Napier Earthquake, the magnitude 7.8 earthquake that hit Hawke's Bay on 3 February 1931 in fact affected the whole region, and the greatest geological effects occurred away from Napier city. These effects are summarised in Box 12.4.

The earthquake resulted from sudden release of the squeezing stresses that had built up over a few centuries from the progressive downwards motion of the Pacific Plate underneath Hawkes Bay. The energy it released caused much damage and loss of life, as well as permanent strain in the rocks. The latter took the form of nearly 4 m of vertical displacement and a smaller amount of west–east shortening, producing one step (of many) in the growth of an anticline (upfold)-syncline (downfold) pair of folds (Box 6.6 A).

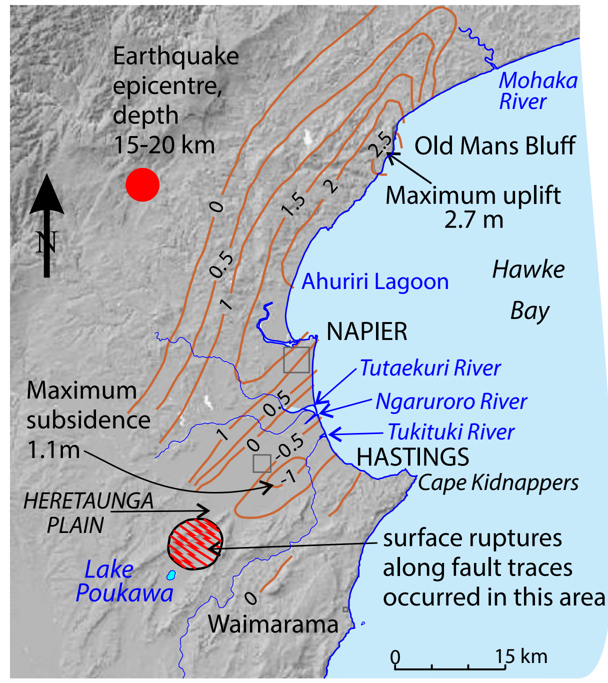
The small folds produced in 1931 trend northeast–southwest, as shown by the contours of uplift and subsidence in Box 12.4. This is parallel to other folds in the region and at right angles to the direction of shortening. They illustrate how the much bigger folds that we see around the region have grown in stages with each successive earthquake over a total period of 25 myr. No individual fold has been growing for all of that time.

The Forearc Basin

In general terms, we would expect to see the forearc basin as a long, narrow, subsiding strip on the trench side of the frontal arc axial mountains. And this is how we do find it northwards from Cook Strait, where it forms Palliser Bay and carries on through Lake Wairarapa and Masterton as far as Hastings. It takes a sidestep to the left north of Masterton, in company with the eastern front of the axial ranges, but is otherwise continuous. SH2 follows the basin from the Manawatu Gorge to Hastings. Rivers, on the other hand, do not follow the basin slavishly, but cut across it and enter the hill country to the east (the accretionary complex). This is a reflection of the long-term history of rivers compared with the short-term history of tectonic uplift and subsidence in this highly mobile area of New Zealand. The river courses are superimposed on the younger geology.

Box 12.4. The effects of the February 3rd, 1931 Napier Earthquake, magnitude 7.8.

Contours (in metres) of the uplift and subsidence that accompanied the Napier Earthquake. They define an upfold-downfold (anticline - syncline) pair trending NE-SW, with a maximum known vertical separation of 3.8 m.

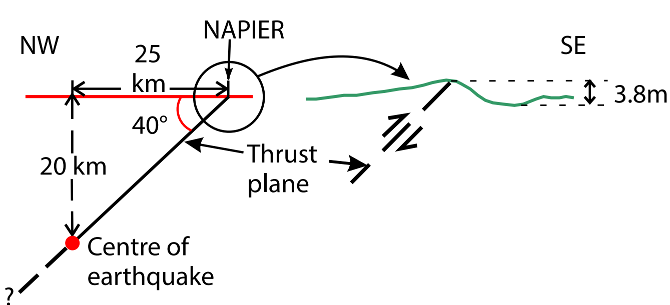


NOTES:

1. The area uplifted measured 125 km from NE to SW. This elongation is parallel to the regional tectonic grain, which is a direct consequence of the shortening caused by underthrusting of the Pacific Plate beneath Hawkes Bay (Boxes 12.1 A-C).

2. The location of the epicentre relative to the fold pair defines a thrust plane (Box 6.6 C) dipping about 40° to the NW and extending at least 125 km laterally (NE to SW) and to a depth of at least 20 km.

Schematic vertical sections to show the thrust plane and its relationship to the surface folds.



The thrust plane is inferred to have been already in existence, and to have moved in many previous earthquakes. A few more metres of displacement on the thrust plane occurred during the 1931 earthquake.

3. The downfold (minus 1.1 m) is located on the Heretaunga Plains. It is simply the latest in a long series of downfolds which have given rise to the Heretaunga Plains forearc basin. It is no coincidence that three major rivers (Tutaekuri, Ngaruroro, Tukituki) discharge into Hawke Bay within a distance of 5 km, all within the area of maximum subsidence in 1931.

4. Despite the great magnitude of the earthquake, surface ruptures along fault traces were confined to a small area between Lake Poukawa and Hastings. Up to 4.6 m of vertical separation and up to 1.8 m of right-lateral horizontal separation occurred, but very few of those ruptures are now visible.

The contours of subsidence formed by the 1931 earthquake (Box 12.4) outline the present-day forearc basin perfectly. That area subsided 1.1 m. Napier-strength earthquakes have an estimated return time of around 600 years, giving a short-term subsidence rate approaching 2 m/1000 years – geologically very rapid. For comparison, the long-term rate of subsidence for the analogous Clifton–Cape Kidnappers cliff sequence (see below), when the basin was in a different location, was about 1 m/1000 years – i.e. lower, but averaged out over a much longer time period and still high on a global scale.

The present-day forearc basin forms the coast of Hawke Bay between Napier and Clifton. Remarkably, the concentrated subsidence here has pulled in three major rivers – the Tutaekuri, Ngaruroro and Tukituki – which together drain a length of more than 100 km of the axial greywacke ranges but discharge into Hawke Bay along just 5 km of coastline (Box 12.4). These rivers carry a great quantity of greywacke gravel, which is why the Heretaunga Plains extend inland through Hastings, why a large gravel-aggregate business can be based here, and why the beaches in Napier are gravel rather than sand. Sand is moving around not far offshore, and on some beaches, while mud is being deposited further out in the bay.

From the Heretaunga Plains, which indicate the area of concentrated subsidence at the present time, subsidence extends inland to the Ruataniwha Plains and onwards to the south (Fig. 12.1 A). We are building a picture here whereby the forearc basin is more or less stable over the longer term (1 myr-plus), but where at the same time areas of concentrated subsidence move around within the overall basin, on a timeframe of hundreds of thousands of years.

The Clifton–Cape Kidnappers forearc basin sequence

Strata forming the cliffs between Clifton and Cape Kidnappers are seen by thousands of tourists each year as they take the tractor ride along the beach to the famous gannet colony. These strata are the record of the forearc basin between 1 myr and 0.6 myr ago, when it was in a slightly different place from today. They are now uplifted for us to see, but dip to the northwest as they are pulled down into the present-day centre of subsidence. The black layers are greywacke gravel, deposited by the three rivers that are now discharging a bit further along the coast (see above), while the intervening lighter-coloured layers comprise grey mudstone deposited in the deeper parts of the ancestral Hawke Bay. As mentioned above, the average rate of subsidence at this location between 1 myr and 0.6 myr ago was 1 m/1000 years, which means that these cliffs are made up of a 400 m thickness of strata. To make sense of the sequence of strata, we need to consider the fact that glacial–interglacial ups and downs of sea-level were superimposed on the steady subsidence of the basin.

Glacial–interglacial sea-level changes

In Chapter 10, describing the Whanganui Basin, we looked at how glacially driven changes of sea-level can be recorded in great detail, provided the sedimentary basin subsides at the right rate and is supplied with plenty of sediment. Those conditions have been satisfied in the Whanganui Basin for more than 5 myr, and at other locations for shorter periods – including in the forearc basin in and around Hawke Bay. This basin is strongly affected by the changes of sea-level (amounting to more than 100 m) that are driven by the glacial–interglacial cycle every 100,000 years or so.

Fig. 12.3. The high sea cliffs between Clifton and Cape Kidnappers are conglomerate (dark layers) and mudstone that were deposited between 1 and 0.6 myr ago during a time of alternating interglacial high (mudstone layers) and glacial low (conglomerate layers) sea levels. The strata have subsequently been uplifted and are now eroding rapidly. Photographer Brent Alloway.

We are presently in an interglacial period of high sea-level. During a glacial low sea-level, the river plains of the forearc basin extend out over the muds of the outer Hawke Bay, and all the sediment bands move offshore. In a subsiding basin, under a depth of water that is sensitive to these changes, the cycles of sea-level change give rise to cycles of alternating shallow-water and deeper-water sediments (Boxes 10.2 A, B). Cyclic strata originating in this way and dated at 1–2 myr old are seen on SH2 between Wairoa and Napier, around Devils Elbow.

In the case of the Clifton–Kidnappers sequence, the depositional environment was exactly like that of the present day but offset a short way to the southeast, with the Heretaunga River gravel plains passing seawards via beach deposits into offshore muds. Each alternation of river gravel and offshore mudstones represents one climatic cycle. Just how you hang ages on the cycles is discussed in Chapter 10, and the same methods apply here.

Walking along the beach from Clifton, going downwards in the tilted strata, an 80 m-thick river gravel (the Clifton Conglomerate) is the first layer encountered at beach level (Fig. 12.3). It forms an impressive cliff. At its base is a 15 cm-thick pumice layer, an airfall deposit from the TVZ. Further along the base of the cliff you find older and older strata. Beneath the pumice layer are sands with shallow marine fossil shellfish, including large specimens of the present-day cockle.

Even further along, and a little lower in the sequence, is a prominent grey mudstone with notable high-angle, intersecting joint planes; it contains mussel-like fossils, and originates in deeper water. Underneath that is a unit of inter-fingering sands and gravels containing some fossil shells and some tree stumps (now black) in growth position; this is a beach/coastal dune sequence. Thus, in 1 km or so of walking you have traversed one 100,000-year climatic cycle dating from around 600,000 years ago: river gravel = glacial low sea-level; shelly sands = beach/nearshore zone during rising sea-level; muds = interglacial high sea-level; beach/dune sands = glacial falling sea-level.

The forearc basin north of Napier

As it approaches Hastings from the south, the active forearc basin for some unknown reason swings to the right and away from the mountains. It enters Hawke Bay, as noted, between Napier and Clifton, but then becomes difficult to trace across Hawke Bay. In a general sense, Hawke Bay itself is part of the longer-term forearc basin, and all the uplifted strata surrounding it and dipping into it are forearc basin deposits, as in the 1–2 myr-old cyclic sediments mentioned above at Devils Elbow, north of Napier.

North of Hawke Bay, however, there is no sign of a forearc basin, at least as defined here and seen everywhere south of this point. It may be that the thick sequences of strata aged between 15 myr and 3 myr that are present north of the Mohaka River and Wairoa should be regarded as the record of the older forearc basin. There is no known explanation for the large difference in overall nature of the forearc north and south of Hawke Bay.

The Accretionary Complex

The accretionary complex lies seawards of the forearc basin and is the zone where most of the squeezing resulting from plate convergence is taken up. The rocks affected by this squeezing comprise all those at the continental margin, regardless of age or origins. Thus we have to begin our deliberation of the East Coast accretionary complex (an active continental margin) by talking about the earlier passive continental margin sequence.

The passive-margin sediment wedge

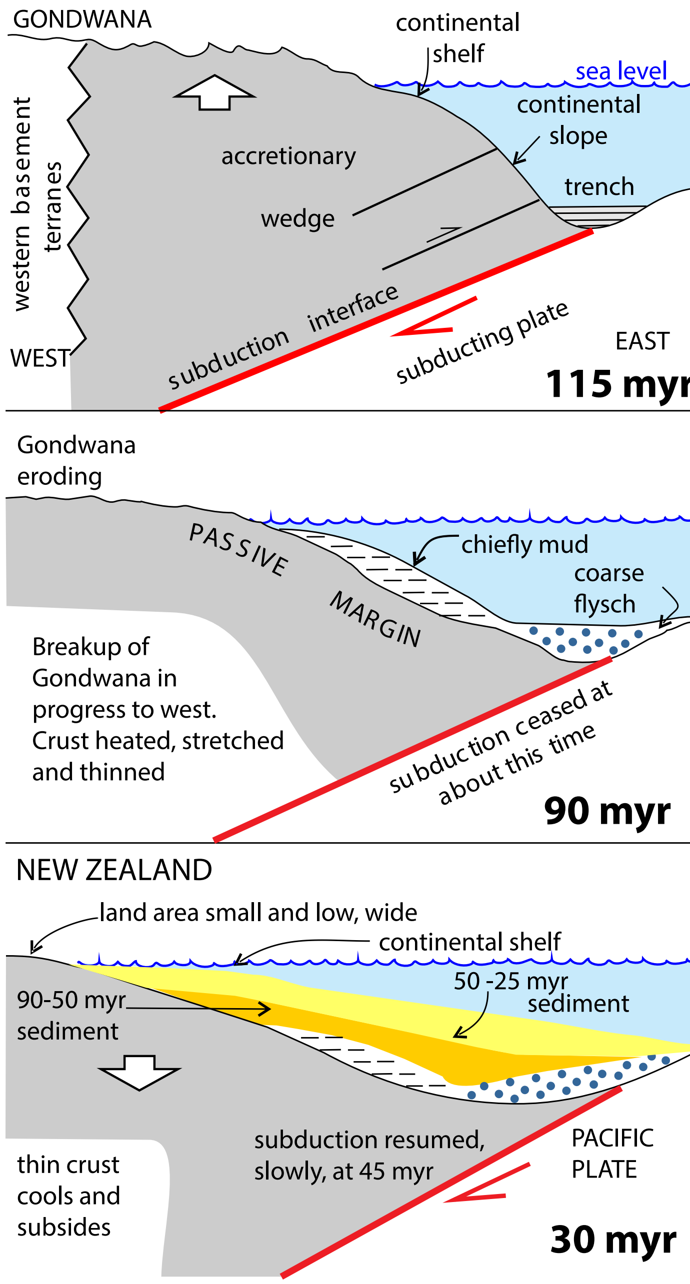
While the actual forearc construction has occupied the last 25 myr, sedimentary rocks formed over the previous 75 myr play an important part in the story. Around 100 myr ago, in phase two of New Zealand's geological history, the old, long-lived regime of plate convergence and subduction at the active margin of Gondwana, which had created the basement rocks of the country, finally ceased (Chapter 5). At this stage, the part of Gondwana that later separated to become the Zealandia mini-continent 80–55 myr ago was part of a passive continental margin. That is, it was a continental margin like the present Atlantic margin of North and South America, where no subduction is taking place, and where sediment delivered from the continent is left alone to build a simple continental shelf and continental slope, a seawards-tapering passive-margin sediment wedge. Eastern North Island was part of this original passive margin, which was simply the Gondwana margin as before, but without subduction. After the opening of the first stages of the Tasman Sea, the passive margin then extended right around Zealandia. Most of this continental margin is still passive today (Fig. 12.4).

The East Coast passive-margin wedge sediments were initially coarse-grained, because the landmass was still mountainous. They were transported by turbidity currents and formed flysch deposits that filled up the remnants of the earlier subduction trench. As the landmass wore down, the sediment became finer-grained, and the amount of lime increased (from marine organisms). Hence, the sediments become finer higher up the wedge. In parts, the wedge is rich in bentonite (swelling clay). This clay expands greatly when it takes up water, and has had a significant effect on tectonic processes, on landscape-forming processes, and on oil exploration, as discussed below.

The East Coast passive-margin wedge is important because it became involved in all the later tectonic processes. It was located directly above the edge of continental crust (the edge of greywacke), which became the subduction interface between the Pacific and Australian plates 45 myr ago (an active continental margin). Thereafter, especially in the last 25 myr, the wedge began to be shortened and thickened ('kneaded') by the compressive effect of the Pacific Plate passing underneath it (Fig. 12.5). This process is still going on today. The primary mechanism behind the kneading is the production of low-angle reverse faults (thrusts) that dip inwards towards the North Island. Rocks underneath the faults are pushed downwards towards the North Island, while rocks above the faults ride upwards and outwards, becoming rucked and folded in the process. The continental crust of the North Island acts as a backstop for the upper levels of the wedge, while lower down, sediment being deposited in the Hikurangi Trough is being pushed underneath everything else to form an accretionary wedge (Box 6.2 C), which is not exposed to view.

During kneading, folding of passive-margin wedge rocks forms sedimentary basins on the trench slope, which then trap sediment coming off the active margin. Along the East Coast, the folds and the basins all tend to be elongated in a northeast–southwest direction, parallel to the backstop and at high angles to the direction of compression. Further folding then involves both younger and older sets of strata. Thus, where we see the uplifted end product of this process (the accretionary complex) on land (where it has been pushed up out of the way and abandoned, as it were, while the process continues further down the continental margin), things tend to be organised into northeast–southwest-trending parallel strips of folded active-margin sedimentary rocks younger than 25 myr, separated by up-thrust strips of highly deformed passive-margin rocks. Most of the basin-fill sediments are turbidity current deposits, or flysch (Boxes 6.5 A, B).

As with the forearc basin, the accretionary complex pattern is clear and obvious on geological maps (e.g. the 1:1,000,000 map of the North Island) south of Hawke Bay, but less so north of Hawke Bay because of the huge volume of active-margin sedimentary rocks here. Nevertheless, small windows of passive-margin rocks seen here and there (e.g. on Mahia Peninsula and around Gisborne) show that the relationships are the same. The greater volume of active-margin sediments north of Hawke Bay is probably due to the fact that during the past 25 myr that area has been closer to the major sources of sediment – the Northland and East Coast allochthons (see below) and the active volcanic arcs. For example, white pumice is a prominent component of the basin turbidite deposits on the northern coast of Mahia Peninsula and around East Cape.



A - Near the end of subduction underneath Gondwana. Trench sediment continues to be stuffed into the accretionary wedge of greywackes (Boxes 6.2 A-D).
Older, higher parts of the wedge are being eroded, and sediment deposited on it (continental shelf, e.g. Koranga Sandstone, and slope, e.g. Karekare Formation).

B - Subduction ceases.
The old trench fills up with coarse-grained flysch (Boxes 6.5 A, B) sourced from still-mountainous Gondwana. The continental slope is blanketed by mud. Gondwana is breaking up further west.

C - 90 to 25 myr ago, New Zealand separates from Gondwana, is worn down (peneplained - Box 4.1), and subsides as the crust cools, giving widespread coastal coal measures (Box 7.3 A). A wide continental shelf and little land sediment gives Te Kuiti shell limestones (Box 7.2 A,B). On the continental slope (passive margin) a wedge-shaped body of sediment buries the old trench, becoming finer grained and more calcareous (limy) upwards. This reflects peneplanation, and drift of the continent from near-polar to temperate-subtropical latitudes.

Fig. 12.4. Formation of the passive margin wedge of fining-upwards continental-slope sediment, 110 to 25 myr, information box. Shown with particular reference to the North Island east coast. However, following final break-up of Gondwana (80 myr) a similar margin wedge, without a basal trench-fill, accumulated all round New Zealand. Vertical scale greatly exaggerated. Fig. 12.5 shows later involvement of the wedge in the East Coast accretionary complex.

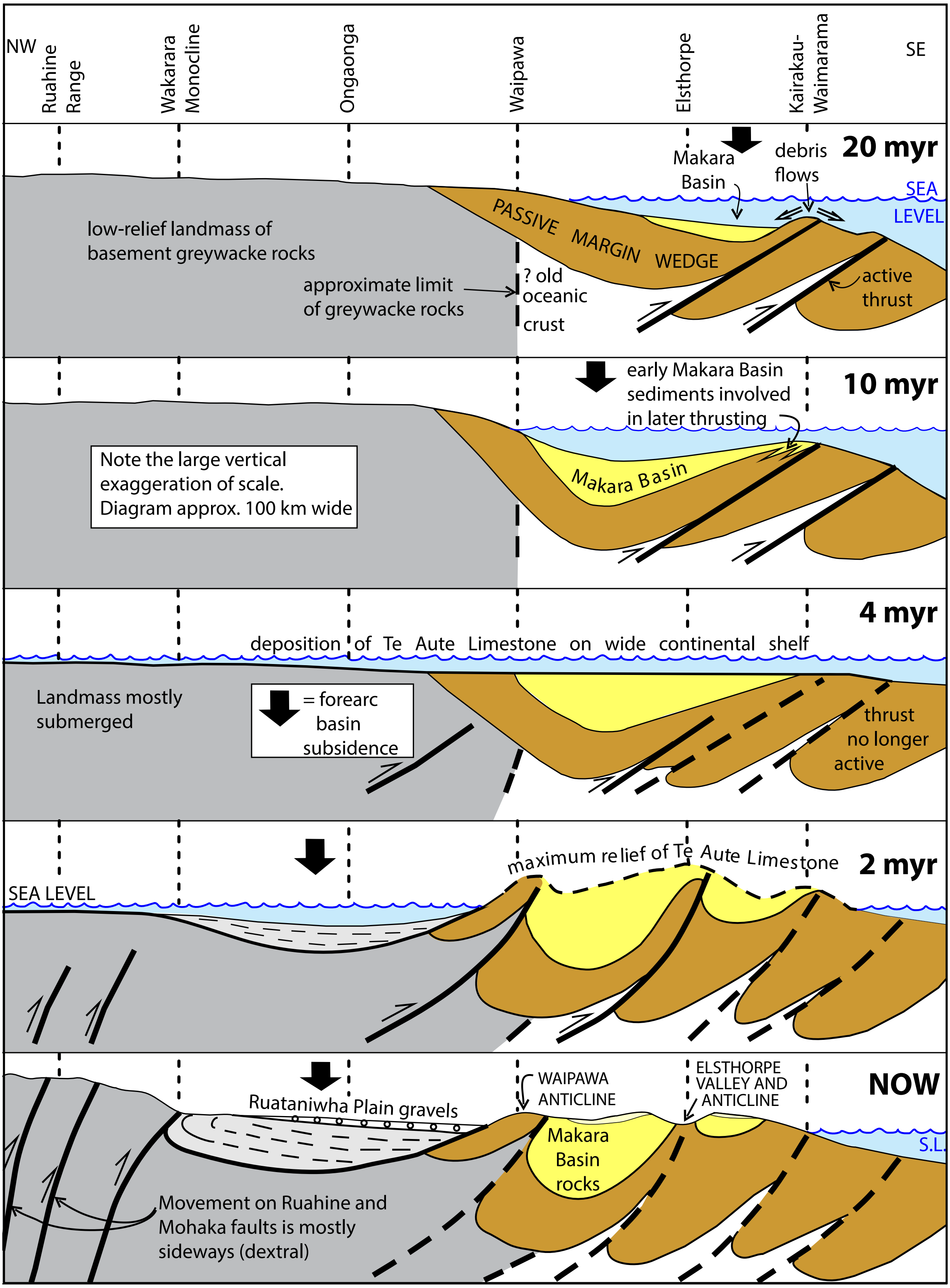


Fig. 12.5. Folding and faulting of the passive-margin wedge.

The East Coast Allochthon

One of the earliest events in the forearc was also the most dramatic: the emplacement of the East Coast Allochthon, 25–20 myr ago (Boxes 6.3 A, B and Fig. 12.1 B). As discussed in Chapter 6, the allochthon (and its former neighbour, the Northland Allochthon) comprises a large northern sector of the passive-margin wedge that was strongly compressed by the initial invigorated plate convergence. It was uplifted above the level of the adjacent continent, allowing it to slide southwestwards onto what is now the North Island. Thus, all of the wedge sediments that we see in the Raukumara Peninsula north of Gisborne (Fig. 12.1 B) are part of the East Coast Allochthon (Box 12.3).

The two allochthons also include ophiolites (slices of basaltic-volcanic ocean-floor crust), made mostly of seafloor lavas that were somehow obducted, or pushed up onto the continental margin, instead of being subducted. The entire northern coastline of Raukumara Peninsula between Cape Runaway and Matakaoa Point is made of these lavas. There are also smaller ophiolite fragments in the accretionary complex further south.

Fig. 12.6. Deformed passive-margin rocks of Cretaceous Whangai Formation at Reporua, on the coast near Ruatoria, East Cape. Photographer Bruce Hayward.

Ruatoria and the five big sandstone mountains

The ranges to the west of Ruatoria are made of Cretaceous-age rocks of the East Coast Allochthon. They include the bodies of sandstone that form a cluster of five big mountains – Hikurangi, Honekawa, Aorangi, Wharekia and Taitai (Fig. 12.7). These can be seen from SH35, from a marked viewing point about 3 km south of the turn-off to Ruatoria. However, there are much better views from the Tapuaeroa Valley Road, a turn-off just north of Ruatoria and the Waipu River bridge. On a clear day the peaks are stunning scenic objects, not least Hikurangi, which is the highest non-volcanic mountain in the North Island.

The bodies of coarse, pebbly sandstone that form these mountains are merely the largest of hundreds of sandstone blocks that 'float' in the East Coast Allochthon like plums in a pudding. The age of the sandstone is early Cretaceous (around 100 myr old) and it is called the Taitai Sandstone. It was originally a continuous, coherent set of strata, before being broken up – stretched and dismembered – and mixed around during emplacement of the allochthon. The sandstone blocks, which are of all sizes, are more resistant to erosion than the surrounding mudstones and flysch, and form numerous knobs and hills.

Fig. 12.7. Four of the five big sandstone mountains, viewed from Ruatoria turnoff on East Cape highway. From left Hikurangi (partly obscured), Aorangi, Wharekia (partly obscured) and Taitai. Photographer Bruce Hayward.

Serious erosion

Older (Cretaceous) members of the passive-margin sequence are mostly flysch deposits, and have been severely deformed. They are very weak rocks. Thus where they, and some of the younger basin sediments, have been well uplifted and dissected by river erosion along the inner margin of the accretionary complex, spectacular erosional features and pervasive slope failures abound (Fig. 12.8). The instability of much of this country is sobering to behold. It is best seen from the inland back roads, such as around Ihungia in the valley of the Mata River.

The large quantity of sediment delivered to rivers via this erosion also causes many problems downstream. Gisborne's Waipaoa River carries the heaviest sediment load in the country. Even away from the extreme examples, slope failure and soil erosion are major problems throughout the forearc (Boxes 12.5 A–C).

Fig. 12.8. The Blue Slip in the lower Ihungia Valley is typical of the erosional features in the weak Cretaceous rocks of the passive-margin sequence in the East Cape area. Photo from Jill Kenny.

Box 12.5 A. Erosion and the production of hillslopes.

Rivers transport sediment to the sea.
Some sediment they derive by eroding their beds and banks, some is delivered to them by landslips.
Therefore:

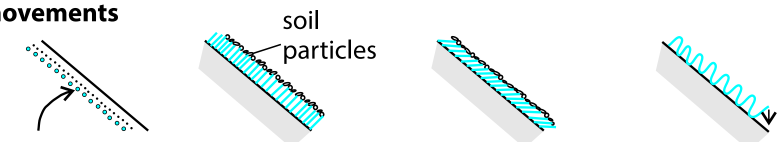
- landscapes are sculptured by a combination of processes, most of which actually relate to the production and evolution of hillslopes.
- landscapes evolve with time, and, if no other forces intervene, relief will eventually be reduced.

The ubiquitous New Zealand "slip"
In New Zealand any visible mass movement on a slope is termed a "slip". This is sloppy nomenclature, because it conceals the fact that there are several processes involved. Some are so rapid they are virtually instantaneous, while others act with glacial slowness — but they are all slips!

BREAKDOWN OF SLOPE-FORMING PROCESSES

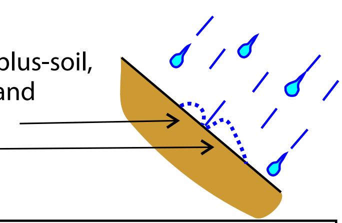
1. Individual particle movements

Freeze and thaw

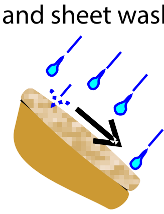


Rain splash on bare soil

Each rain drop impact creates an outwards explosion of water-plus-soil, which is symmetrical. However, gravity slows the uphill splash and accelerates the downhill, therefore uphill splash is less than downhill splash.

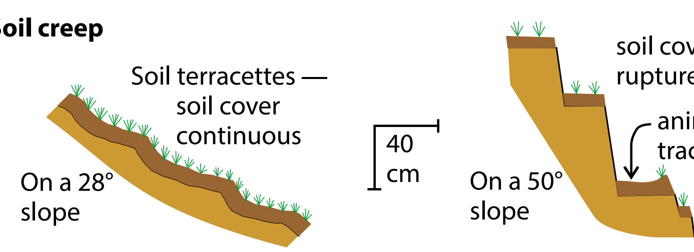


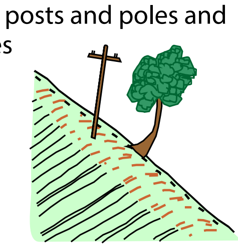
Splashes accumulate in heavy rain to form rill and sheet wash



These three processes all move soil particles downhill. They are hindered — but not entirely prevented — by a cover of vegetation. Forest and forest litter absorb much more water than grass does, and also insulate against frost.

2. Soil creep

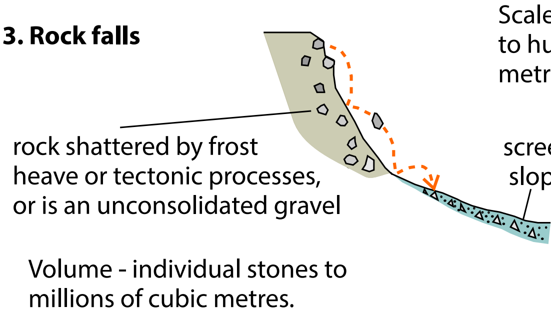


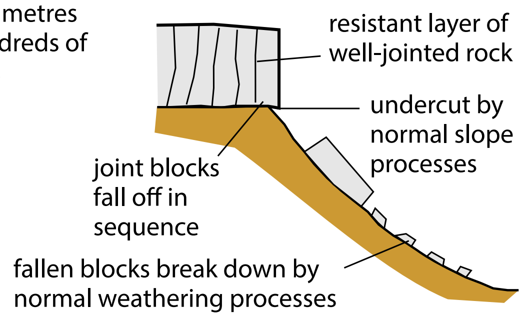


Rates of soil creep are in the range 1 mm to several cm per year.
Soil terracettes do not occur on gentle slopes. Although they are commonly called "sheep tracks", in fact sheep and cattle do not cause them; they use and accentuate them.

Box 12.5 B. Erosion and the production of hillslopes – *continued*.

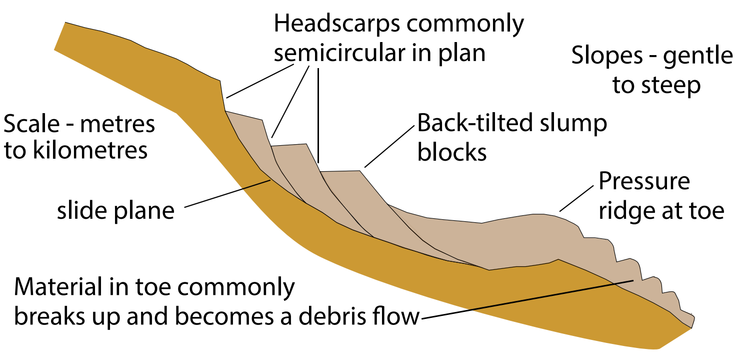
3. Rock falls





4. Rock slides - the general case.

Material that slides is not deformed internally. It requires a basal slide plane.



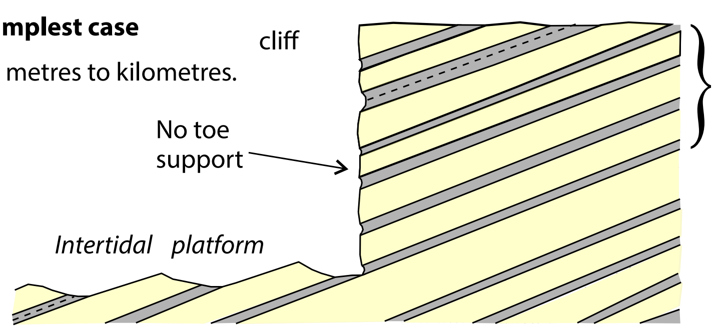
In the general case the rock is essentially homogenous, and the depth to the slide plane (from the original surface) is determined by physical factors like downwards decrease in weathering (hence increase in rock strength), and degree of water saturation. The semicircular headscarps are commonly seen on road surfaces, where the road edge is unstable.

5. Rock slides - the special case.

The special case is where the slide plane is a geologically pre-determined plane of weakness. This is usually a thin layer of clay, which may be an original sedimentary layer, or a later-developed shear plane. Almost always, the slide plane is inclined in the same direction as the overlying surface slope. The slide process is normally triggered by removal of support from the toe region. Speed of process - very slow (mm per year) to fast (metres per second).

The simplest case

Scale - metres to kilometres.



Any of these mudstone layers could become a slide plane, e.g. if a heavy structure were placed above, or if pore-water pressures were raised.

Alternating layers of sandstone and mudstone.
Diagram for a coastal cliff, but could equally be a river cliff, road cut or quarry.

Box 12.5 C. Erosion and the production of hillslopes – *continued*.

6. Rock flows

A flow is a moving mass of material (air, water, sediment, rocks) which deforms internally. Many rock falls and slides are transformed into flows at their front ends, e.g. the large rock fall from the summit of Mt. Cook in December 1991. The most common kind of rock flow is the debris flow - see 'Andesite Volcanoes' (Box 9.2 B) for details.
Volume of flows - up to millions of cubic metres
Rate - mm per year (rock glaciers) to metres/tens of metres per second.

7. Mud and earth flows.

These are similar to rocky debris flows in terms of flow mechanism, volumes and rates, just different in being mud-dominated. Their importance is high in much of the North Island steep hill country, where they are the dominant slope-forming and slope-eroding mechanism.

LONG-TERM DEVELOPMENT OF SLOPES

The long-term development of slopes is not fully understood. Because they are losing material all the time they must retreat. In many places the ash of the Taupo 232AD eruption (Box 9.3 A-C) has been completely removed from hill slopes, despite the forest cover which was present for most of the last 800 years.
In some parts of the country, slopes are steep and straight, and retreat without the slope changing, so that adjacent valley slopes meet in razor-back ridges. In other places they are rounded. They are influenced by varying resistance to erosion of the underlying rocks (Boxes 7.2 B, 12.6). They are also influenced by having material deposited on them (Boxes 9.4, 10.3).
New Zealand is a good place to study the relationship between slopes and tectonic processes of uplift and subsidence. Subsidence generally causes sedimentation and wide river plains - e.g. the Heretaunga Plains of Hastings (Box 12.4). Where uplift prevails there is a direct relationship between rate of uplift and height of the country. The highest uplift rates in the North Island, 6 to 8 mm per year, occur along the axial ranges from Wellington to East Cape.

Bentonite and hydrocarbons in the accretionary complex

The East Coast is a hydrocarbon province, with many natural-gas seeps and a few oil seeps, nearly all of which are associated with exposed areas of passive-margin wedge sedimentary rocks. These rocks include the Waipawa Black Shale, a carbon-rich marine mudstone that would be an ideal hydrocarbon source rock if it were put through the proper cooking process – baking at 100–120°C for a while. As it happens, the carbon in the Waipawa Black Shale tells us that it has not been cooked properly. Therefore, the hydrocarbons must be coming from some other passive-margin sedimentary formation, although at the moment it is something of a mystery as to which one it is.

In addition to the source rocks in the older passive-margin sequence, the overlying forearc sediments contain many porous and permeable sandstone and limestone bodies that could make good reservoir rocks, while the rise of bodies of passive-margin sediments here creates anticlines, or arch folds. These make good traps for oil and gas, provided there is a suitable seal (usually impermeable muddy sediment) on top. See Chapter 11 for a more extensive discussion of the geology of hydrocarbons.

In light of the geology, it is therefore not surprising that there has been extensive oil exploration on the East Coast – chiefly onshore, but also offshore. The first oil seeps were reported in

Poverty Bay, Gisborne, in 1865, and the first wells were sunk there at Waitangi Hill in 1874 by Poverty Bay Oil. For comparison, the world's first oil well was sunk at Titusville, Pennsylvania, in 1859, and the first Taranaki Basin well was drilled in 1866.

However, despite continuing exploration, no commercial discovery was made until the very end of the twentieth century, representing 125 years of fruitless endeavour and expense. Why was this? There are several reasons. One is the passive-margin sediments themselves, which are technically difficult to drill through.

The passive-margin sediments contain smectite clays, which swell when they absorb water. In parts of the sequence these are so abundant as to form bentonite, a white clay deposit almost entirely composed of swelling clay. Ironically, the oil industry uses bentonite as drilling mud. When mixed with water, it has thixotropic properties that are ideal for drilling (thixotropic materials set like jelly when not disturbed and liquefy when disturbed, and alternate between the two states fast and often). When drilling is in progress, the mud liquefies, cooling the rotating drill bit and carrying rock chips back to the surface; when drilling stops, the mud gels, and the rock chips cannot sink back down the hole.

However, having dry bentonite in the rocks you are drilling through is another thing altogether, as it soaks up water from the drilling mud, swells and closes the hole. Bentonite has been mined in the past at Porangahau, where its prevalence at the surface also gives rise to marked slope instability. It is the ultimate weak rock – fenceposts sunk into it have a tendency to lie down!

Another reason for the lack of success in oil drilling on the East Coast is the nature of the anticlines here, which tend to be steep-sided, narrow and heavily faulted. Faults let hydrocarbons escape, while the narrowness makes it difficult to position the drill to best advantage and limits the volume of the potential trap. To date, drilling attempts on the visible anticlines have all failed. Drilling away from the anticlines has encountered the twin problems of much thicker rock sequences to drill through, and the general absence of a suitable cap rock or seal. In some cases, the target reservoir rock – the highly porous and permeable Te Aute Limestone (see below) – changed in character when tracked underground, from limestone to a useless mudstone. In other cases, the limestone was present underground, and had good seal rocks on top of it, but hydrocarbons were absent because there were no associated passive-margin source rocks underneath.

Mud volcanoes

The combination of tectonic compression and swelling clays gives rise to the phenomenon of the mud volcano. The passive-margin sediments that contain swelling clays tend to be pushed or injected upwards in the form of narrow uprisings, or diapirs, in response to compression. The ultimate expression of this tendency is the forcible ejection of liquid mud at the surface, in periodic mud eruptions. There are several of these mud volcanoes in the Raukumara Peninsula, e.g. at Waimata Valley. Mud is squirted quite high in the air, and falls back with much splashing to build a very low-angle mound of grey mud. The mud is usually associated with salt water and natural-gas seeps.

Fig. 12.9. Small mud volcanoes in a farm paddock at Waimata, inland from Gisborne. Photographer Bruce Hayward.

Hot springs

Forearcs are generally areas of low heat flow, so hot springs are rare. There are, however, two hot springs in the North Island forearc, both north of Hawke Bay: one at Te Puia Springs near Ruatoria, and one at Morere, north of Mahia Peninsula. The source of the heat for these springs is not known.

Southern portion of the forearc

At the southern end of the accretionary complex, south of Cape Turnagain, the old continental margin – the edge of old greywacke rocks – swings to the east. Here, outboard of the forearc basin, the accretionary complex contains fault-bounded slices of basement greywacke rocks. These form spectacularly rugged country, culminating in the Aorangi Range of greywacke, which extends to the southern tip of the North Island, Cape Palliser. The Brocken Range, lying between Stronvar and Te Wharau, 27 km southeast of Masterton, is made of the youngest greywacke rocks (around 100 myr old) and has a striking topography, with many upstanding sandstone ridges formed on vertical strata. These landforms are known locally as taipos (Fig. 12.10) and are found beyond the Brocken Range, e.g. the Mangapakeha Taipos, north of Carswell. In this part of the accretionary complex, close to the Cook Strait transition from a subducting plate boundary to a transcurrent boundary, subduction is so oblique that sideways motion tends to dominate over compression.

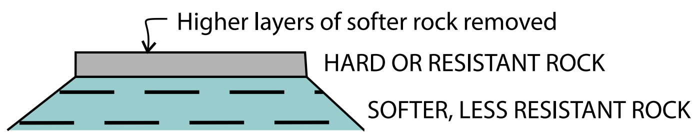
Fig. 12.10. The Tinui taipos are formed from upstanding vertical beds of sandstone within the Cretaceous greywacke sequence that forms the Brocken Range. Photographer Bruce Hayward.

Prominent landscape-forming rock units

In addition to those already mentioned, there are some rock units of the forearc that form prominent landforms. The general type of landform is the escarpment and dip-slope profile formed on an erosion-resistant stratum that is dipping at any angle between horizontal and vertical. Box 12.6 illustrates the generalised profiles.

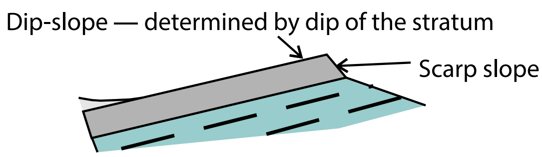
Box 12.6. Landforms related directly to resistant strata: schematic vertical sections.

(1)



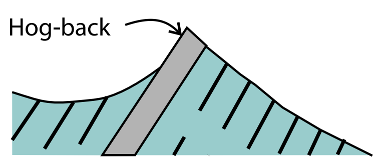
Strata horizontal — "table mountain"

(2)



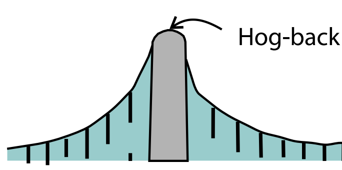
Gentle dip — "Scarp-and-dip-slope profile"

(3)



Steep dip — "Hog-back" or "Razor-back"

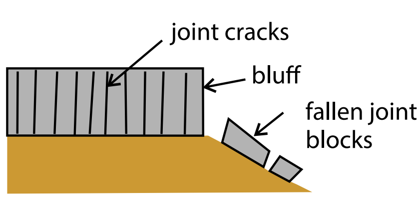
(4)



Vertical dip — "Hog-back"

Identical landforms can be caused by resistant igneous intrusions, either dikes or sills.

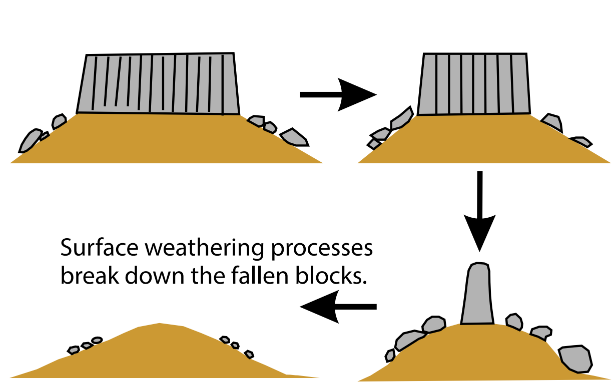
(5) Influence of joints



Scale - commonly tens of metres

Resistant rocks are commonly broken up by joint cracks. The bluff retreats in steps, as joint blocks fall off, whereas the less resistant rocks beneath, retreat steadily by normal slope erosion processes (Box 12.5) e.g. most commonly limestone, welded ignimbrite, sandstone, volcanic conglomerate.

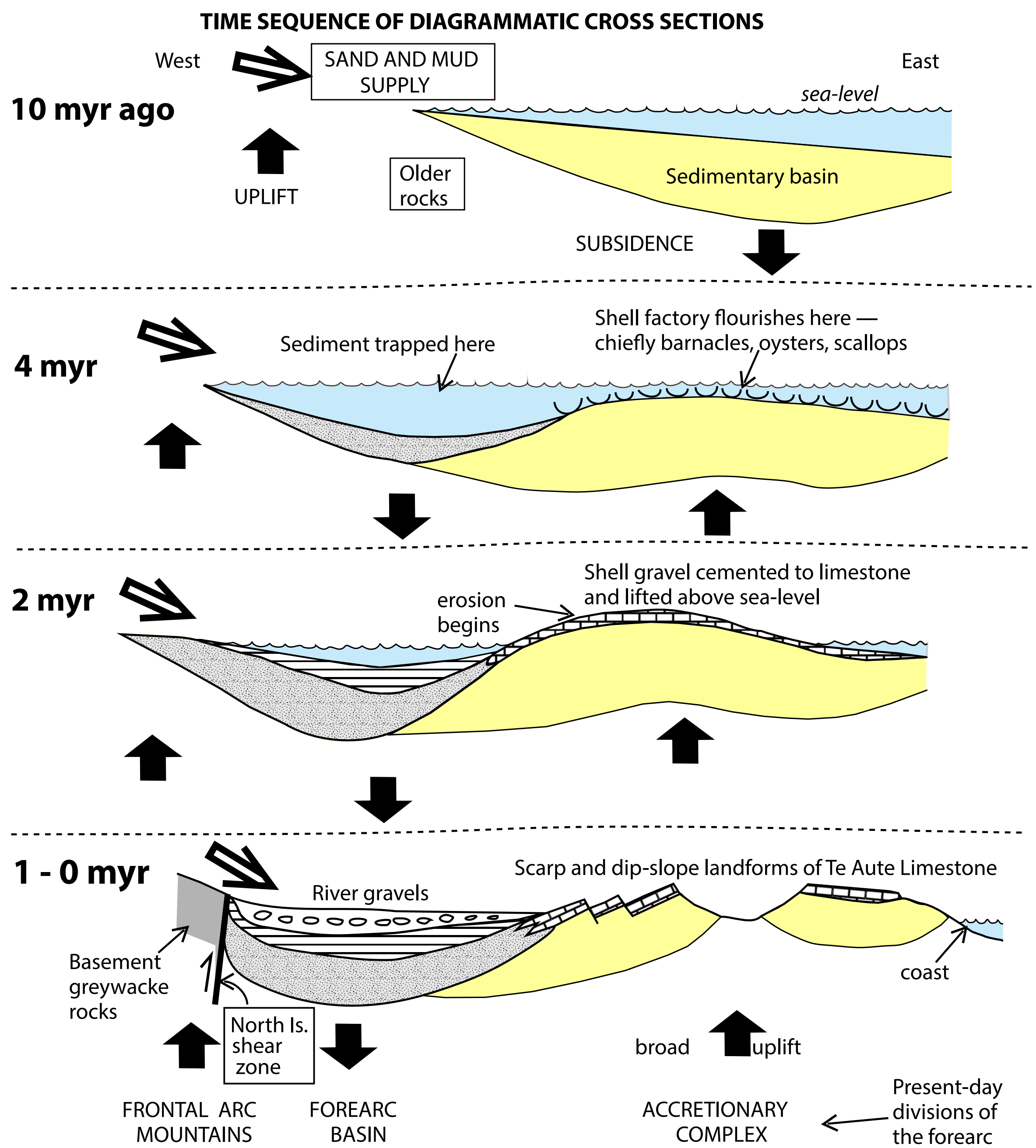
(6) Residual bluff landforms



e.g. Mt. Manaia, Whangarei Heads.

First, let's look at the Te Aute Limestone. These localised bodies of limestone tell an interesting story of the progressive uplift of the accretionary complex from continental slope depths, through continental shelf depths, to land. They are the products of 'shell factories' that developed here and there on the continental shelf, around 4 myr ago, on up-folded high points that were isolated from most of the continental shelf sediment (Boxes 12.7 A, B). The resulting limestone bodies are found

Box 12.7 A. Localised shell factories – the special case of the Te Aute Limestone of eastern North Island. The width of the diagram is about 60 km. Vertical scale is greatly exaggerated. Detail varies up and down the North Island east coast.



Box 12.7 B. Te Aute Limestone.

The prominent Te Aute Limestone of eastern North Island is present everywhere south of Gisborne. It is not the remains of a single layer, but consists of many separate lenses of limestone. They vary widely in thickness and extent, and also in age, ranging between 5 and 3 myr. Each lens formed in an area of shallow sea that was starved of sediment and therefore was host to myriads of shellfish - a 'shell factory' (sometimes called carbonate factory, because shells are made of calcium carbonate).

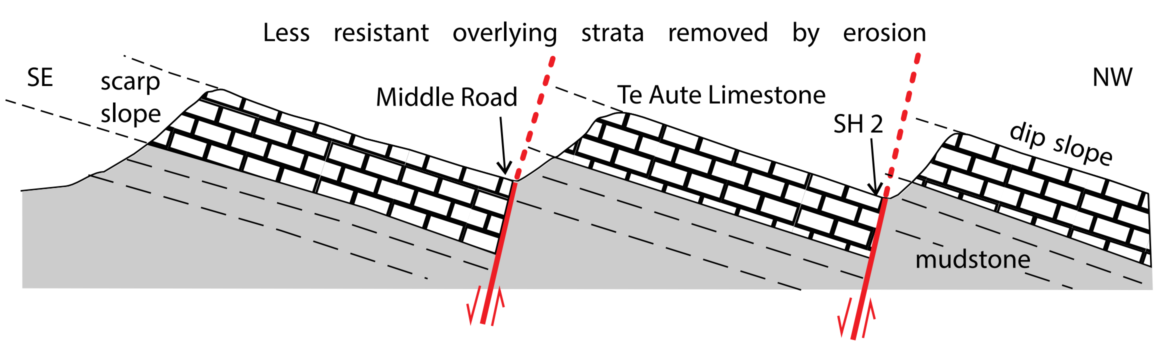
The commonest shells are barnacle plates (each barnacle shell being made of several plates), oysters, scallops, bryozoans (moss animals), brachiopods (lamp shells) and microscopic foraminifera. The shells were mainly pounded to fragments by wave action and predators. There is usually some quartz sand mixed with the shell fragments. Following deposition the shells were cemented by calcite to form limestone, as described in Boxes 7.2 A, B. Cementation of Te Aute Limestone is generally weak.

The commonest scenario for creation of the necessary shallow seafloor, isolated from sediment input, is shown in the time sequence of cross sections. As discussed at the beginning of this chapter, the forearc region of eastern North Island has been accumulating sediment in various ways for 25 myr, and, as a consequence, is being uplifted and added to the land area of the North Island. Each part of eastern North Island has been through the same sequence, beginning as a deep-water part of the trench slope, before being uplifted to form first the continental shelf, and then land. While at continental shelf depth there was commonly a deeper trough nearer to the land, the forearc basin, which trapped sediment coming off the land, thus allowing the shell factory to get established.

The localised uplifts which gave rise to the many separate bodies of limestone were commonly converted into anticlinal (arch) folds as uplift continued and lifted the area above sea-level. Erosion of the anticlines has given rise to the many limestone scarp and dip-slope profiles (Box 12.6) that we see throughout eastern North Island.

from just south of Gisborne to the central Wairarapa, and form spectacular scarp and dip profiles (Fig. 12.11).

The best-known Te Aute Limestone landform is Te Mata Peak (399 m) at Havelock North, along with its extensions well to the south. North of Hawke Bay, Whakapunake reaches 962 m and is prominent from many viewpoints, including Te Mata Peak. Both examples dip to the northwest. The Puketoi Range east of Pahiatua (to the southwest of Hawke Bay) is 30 km long, reaches a height of 803 m and also dips to the northwest. Mt Kahuranaki, in the Tukituki Valley south of Hawke Bay, is a large mass of limestone forming a stand-alone mountain 646 m high that is visible from a wide area of southern Hawke's Bay (Figs 12.12 and 12.14).



How one bed of limestone can form two or more parallel ridges having the same scarp slope-dip slope asymmetry — repetition by faults. Compare Fig. 12.14, repetition across a fold.

Fig. 12.11. Scarp and dip in Te Aute Limestone.

Fig. 12.12. The Te Mata Peak scarp and dip-slope ridge is formed of hard, erosion-resistant Te Aute Limestone that accumulated as shell banks in shallow seas about 4 myrs ago. It has subsequently been uplifted and tilted to the northwest and partly eroded away. Photographer Egon Eberle.

Next, there are various sandstone strata to explore. Because of the flysch character of most of the continental slope basin deposits that are now uplifted as part of the accretionary complex, there are millions of individual sandstone layers. These can become prominent in forming the landscape when they are particularly thick or resistant. Quoting examples from north to south, the Tokomaru Sandstone, north of Gisborne, is typically very thick and gently dipping, and forms bluffs and high plateaux, e.g. to either side of Tokomaru Bay.

Around Lake Waikaremoana, the Otaunoa, Kotore, Panekiri, Ngamoko and Matakuhia ranges are all magnificent sandstone escarpments. The Lake Waikaremoana Great Walk climbs the Panekiri Range dip-slope from the road to the hut at 1180 m, and then clambers down the escarpment to the next hut on the lake shore. Lake Waikaremoana is New Zealand's largest landslide-dammed lake. The Waikaretaheke River cut a deep gorge through the sandstone escarpment, in response to rapid uplift. Landslides of sandstone blocks from both sides of the gorge then formed a dam through which lake water escaped by seepage, before the seeps were plugged by hydroelectricity engineers to create an overflow for power generation.

In the Tukituki Valley, south of Havelock North, the Silver Range is a series of razor-back ridge segments extending 30 km from north to south (Fig. 12.13). The landform here is determined by the fact that the resistant sandstone layers are only a few metres thick and dip westwards at around 45°, the ideal angle for a symmetrical scarp and dip profile. Extensive areas of sandstone bedding plane are exposed on the western dip slope, best seen from the St Lawrence Road, west of Elsthorpe (Fig. 12.14).

Fig. 12.13. The razor-back Silver Range south of Havelock North is composed of an erosion-resistant sandstone bed that dips at 45° to the west (right). Photographer Egon Eberle.

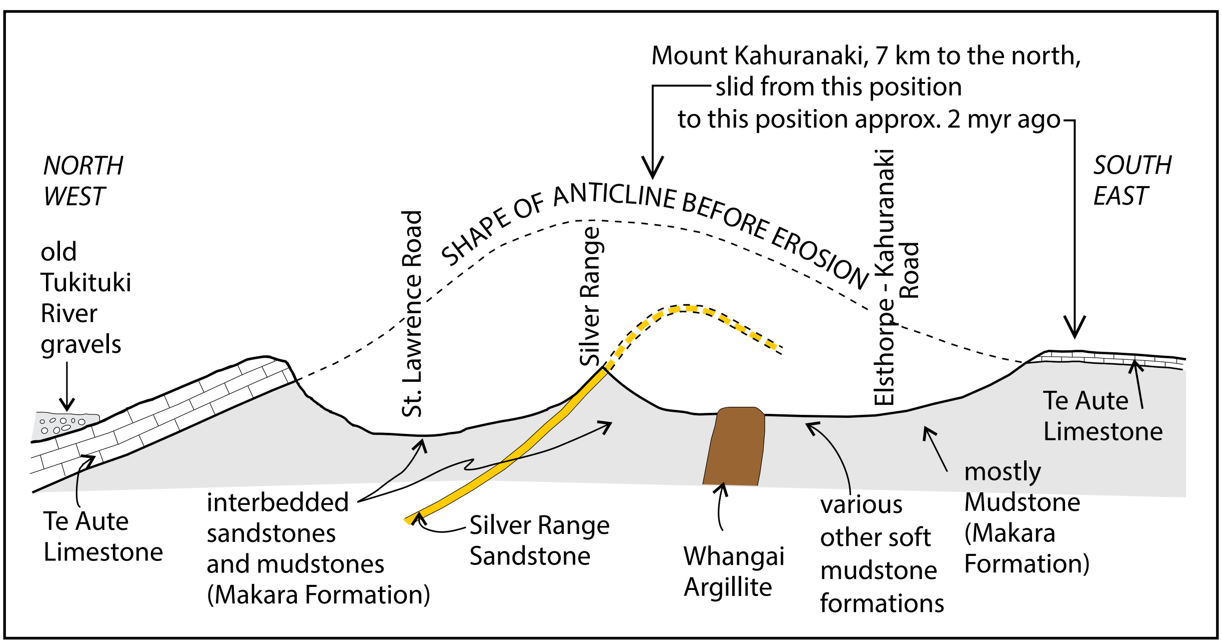


Fig. 12.14. Sketched geological cross section across the St Lawrence and Elsthorpe valleys. Vertical scale is exaggerated. Length of section is 7 km.

Chapter 13

The Wellington – Hutt Valley Region

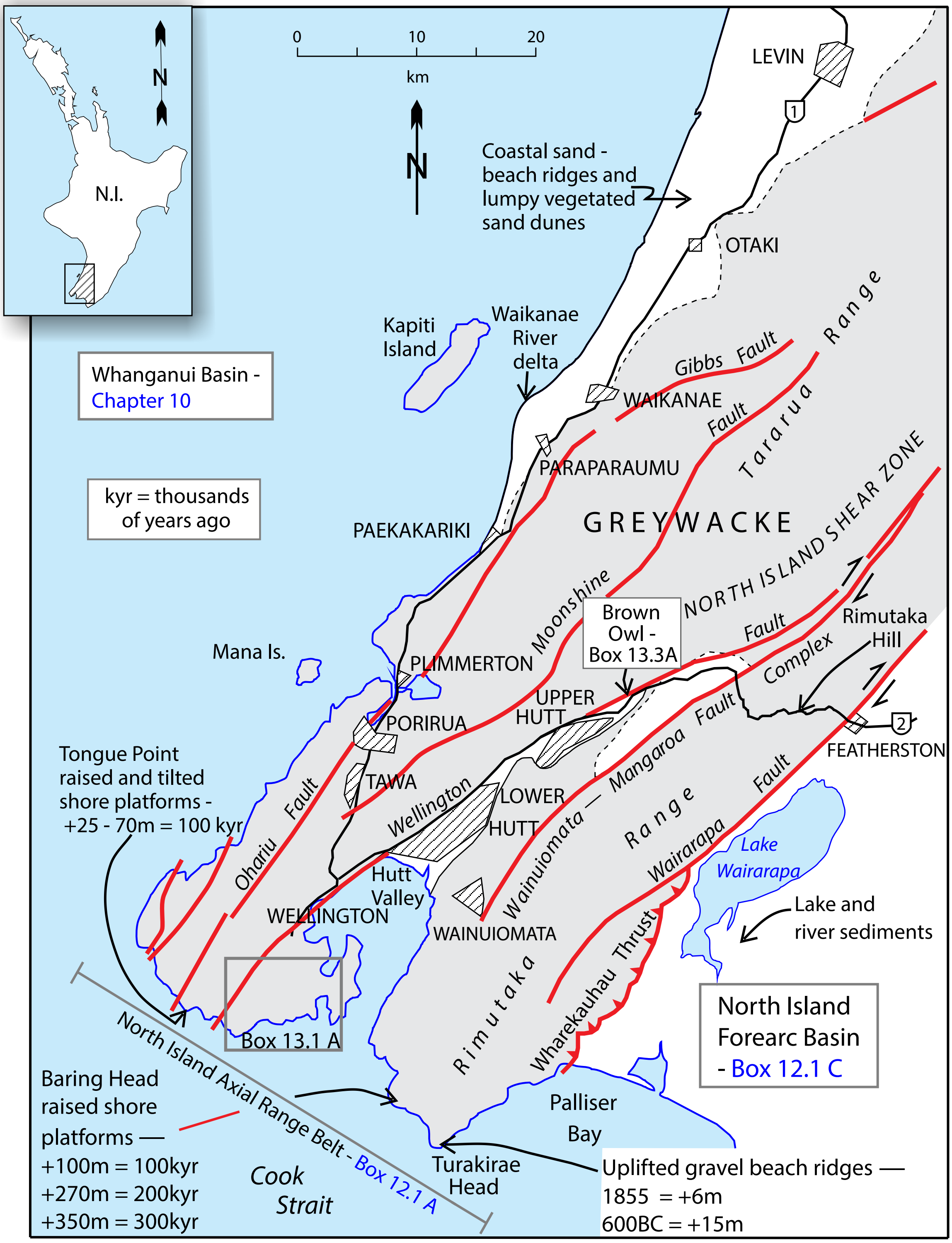


Fig. 13.1. Simplified geology of the Wellington and Hutt Valley region.

Wellington can legitimately claim several capital titles. As well as being the administrative and political capital of New Zealand, it is the greywacke capital of the world, the earthquake capital of the world and the tectonic landscape capital of the world. San Francisco might contest these titles, but few other cities could.

Greywacke Capital

Wellington lies at the southern end of the long, narrow North Island axial ranges, which in geological terms comprise the frontal arc of the North Island subduction zone and forearc (Chapter 12). Fig. 13.1 indicates the extent of the greywacke rock association that makes up the ranges, underlies Wellington and surrounds the city on all sides. In Wellington and the Hutt Valley, greywacke is the only rock you see, apart from the rich variety of imported rocks used in building façades and monuments. The best place to examine the greywacke rock association – so called because it contains a characteristic association of rock types that directly reflect its origins in a deep-sea trench and accretionary wedge above a subduction zone – is on Wellington's south coast from Lyall Bay along to Red Rocks (Boxes 13.1 A–C).

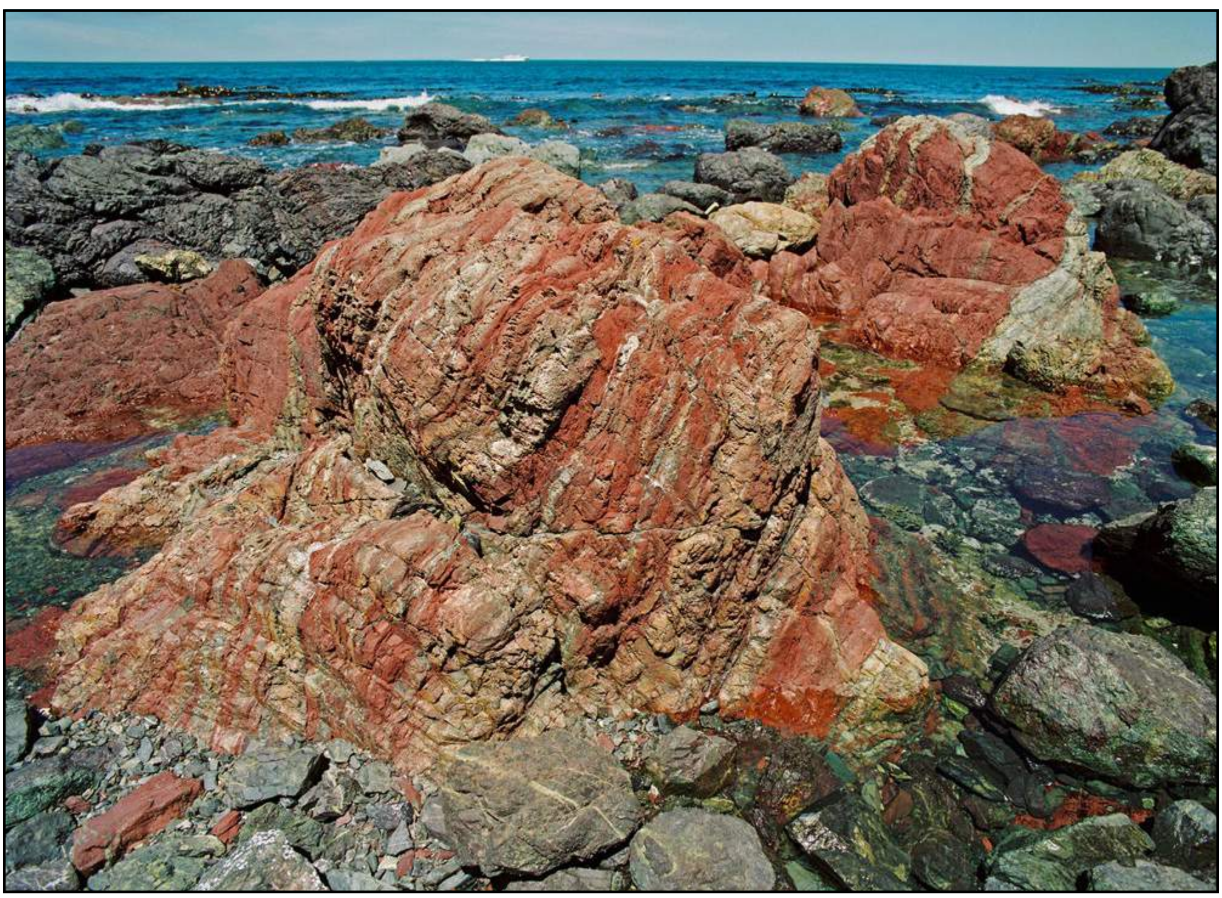
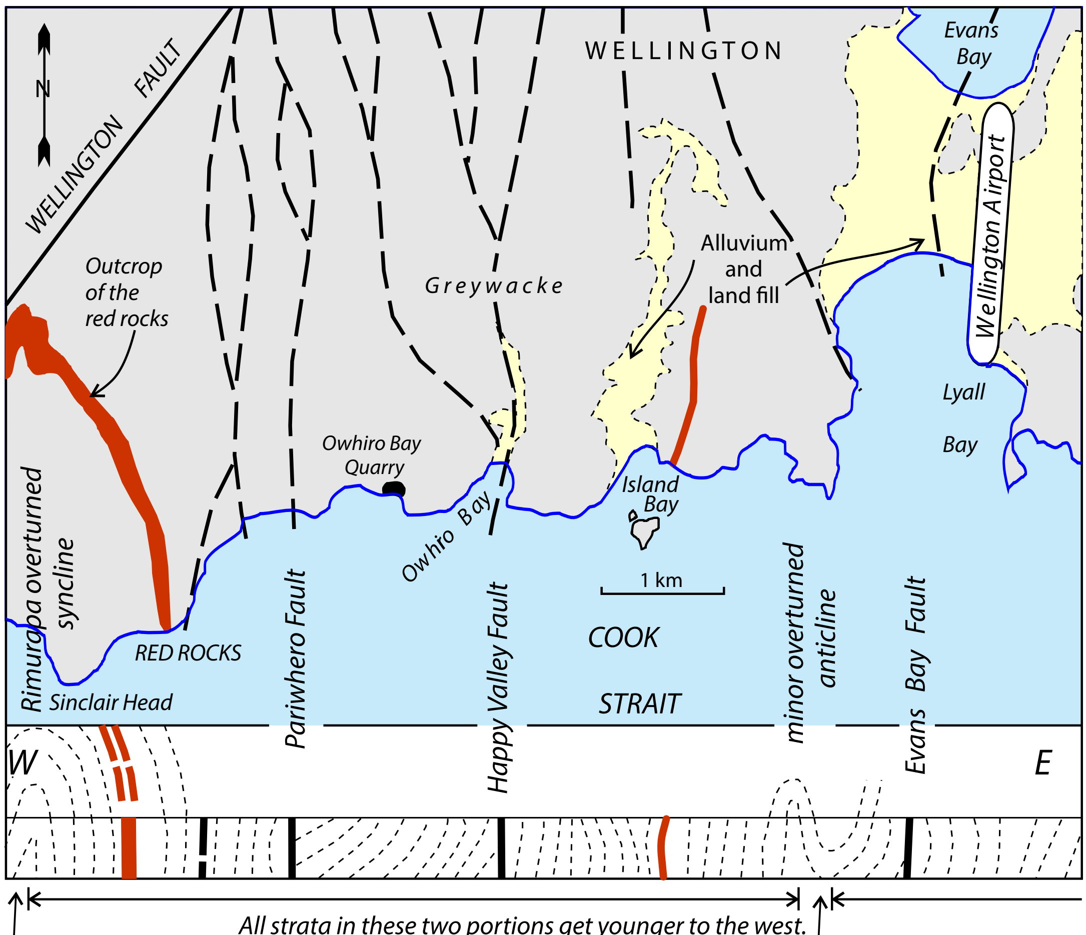


Fig. 13.2. Bedded red chert and argillite are the basis for the name of Red Rocks, Wellington. The chert was formed by the accumulation of the siliceous shells of microscopic radiolaria on the floor of the deep ocean in the late Permian. Photographer Lloyd Homer, GNS Science.

Box 13.1 A. The greywacke rock association of the Wellington south coast. Map and cross section.



Strata from here westwards to Oteranga Head, 8 km away, get younger to the east.

Strata in this short segment on the SW corner of Lyall Bay young to the east, defining a small 'parasitic' anticline on the limb of the much larger overturned syncline.

NOTE that these strata appear to define a major syncline, the Rimurapa Syncline, at least 16 km wide, which is entirely upside down. It has to be a syncline (trough-fold) because strata get younger towards the middle. In an anticline (arch-fold) strata get older towards the middle (Box 6.6 A). This structure is older than 100 myr. The unknown factor is the extent to which the younger faults shown have shuffled together segments of different structures, because they all have some sideways motion.

Box 13.1 B. Red Rocks Scientific Reserve and the greywacke rock association of the Wellington south coast. Discussion and diagrammatic column.

Access

There is a continuous waterfront road (courtesy of the 2 metres of uplift in the 1855 earthquake) from Lyall Bay to Owhiro Bay Quarry. West of the quarry there is a vehicle track to Red Rocks and Sinclair Head which passes through the quarry and is available to the public six days a week. The gate is closed on Sundays. Foot access to the coast is available at all times. To see the rocks well you need low tide. This is a south-facing coast in the roaring forties and on the shores of Cook Strait — watch the sea and weather. You should not damage rocks in the Scientific Reserve, or remove specimens.

Greywacke rock association

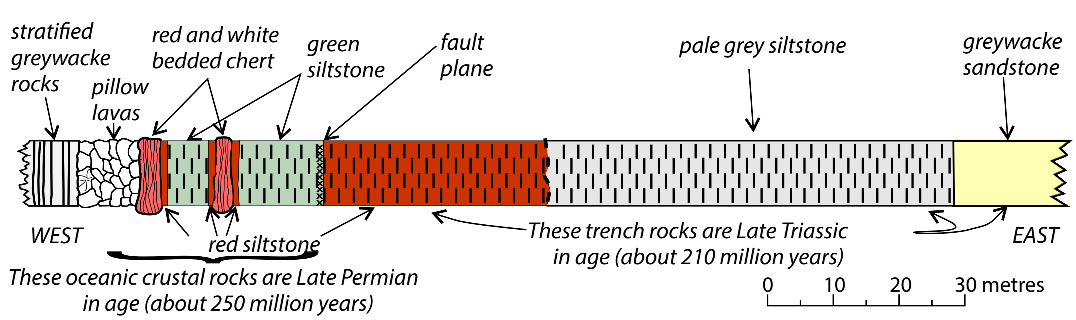
The coastal rocks from Lyall Bay to Red Rocks make an excellent case study of the greywacke rock association, as described in Boxes 6.2 A-C, Fig. 6.7 and Boxes 6.6 A-F. Most of it is the greywacke part of the association. The oceanic crustal rocks are seen only in a narrow band at Island Bay (map) and in the Red Rocks Scientific Reserve. All strata dip steeply.

From Owhiro Bay Quarry westward to Red Rocks, things to look for include:

1. Sedimentary features of the flysch sedimentary association (Boxes 6.5 A, B). Try to figure out the direction of younging (original upward direction) of the sequence from the size grading and abrupt bases of the sandstone layers — it should be generally to the west, but because small metre-scale folds are present you may find a bed younging to the east. A few conglomerate layers are present, including an imbricate arrangement (uniformly inclined like a shelf of tilted books) of the dark argillite rip-up clasts in a graded sandstone bed.
2. Metre-scale fold and shear structures. Some folds of this size show extreme thickening and thinning of the sandstones, and are interpreted as sea-floor slumping structures (Box 6.6 B).
3. Vertical to steep westerly dips (in the walls of the quarry, and in places in the cliff), as depicted in the cross section. The overall structure between Wellington Harbour entrance and the Wellington Fault (and possibly further west to Oterangi Head) appears to be a giant syncline (a trough-shaped fold) more than 16 km wide, centred just west of Sinclair Head. The syncline is entirely overturned (so it is now arch-shaped, like an anticline). See the cross section on previous page (Box 13.1 A).

Oceanic crustal rock association

The particular point of interest at Red Rocks, and the reason for the Scientific Reserve, is the excellent display of these rocks. What you see in the shore platform and cliff is laid out in the diagrammatic column —



Box 13.1 C. Red Rocks Scientific Reserve, continued.

Rock ages are determined by the species of fossil radiolaria (microscopic animals) present. Note that the rocks all get younger towards the west; that the older oceanic crustal rocks are bracketed by trench sediments 40 million years younger; and that the lower (righthand) contact of the oceanic rocks is a fault plane. These facts all fit the "offscraping model" of origin of these rocks (Boxes 6.2 A-C, Fig. 6.7). The puzzling aspects are red siltstones in both the trench sediments and the offscraped oceanic crustal rocks; and pillow lavas apparently lying on top of the deep ocean sediments (cherts, red and green siltstones) rather than underneath them.

Key points to locate yourself are the chert bands, thinly bedded and folded, and in part brick red in colour; and the pillow lavas, bulbous features up to a metre across, in a pile, and generally glistening in appearance. Try to use the shape relationships of the pillows (Box 6.4 A) to determine original top of the pile - later pillows mould themselves around existing pillows; you need more than one pillow shape to be confident of a younging determination.

Note that there are shear surfaces here and there, so that the original sequence has been disturbed.
For more detailed information see:

"Red Rocks: a Wellington Geological Excursion" by Rodney Grapes and Hamish Campbell, 1994, Geological Society of New Zealand Guidebook 11, 32 pp. Available from www.gsnz.org.nz/guides.php

"Geology of the Wellington Area, scale 1:50,000" by J.G. Begg and C. Mazengarb, 1996, Institute of Geological and Nuclear Sciences geological map 22, 1 map sheet plus a 128 page book. Available from GNS Science, www.gns.cri.nz/home/products/Publications/Maps

Earthquake Capital

Earthquakes (Boxes 13.2 A–C) and fault fractures (Boxes 6.6 C, D, G) are opposite sides of the same coin. Earthquakes occur from time to time in order to relieve stress built up in rocks at plate boundaries by plate movement. Much of the resulting strain is expressed in movement along a fault. Wellington is bisected by the Wellington Fault, and the city's claim to the title of earthquake capital of the world was established back in 1855.

The Wairarapa Earthquake

The 1855 Wairarapa Earthquake, which caused considerable damage and uplift around fledgling Wellington, was, in plain English, a whopper. There were no seismographs at the time and, fortunately, few people, because the magnitude has been estimated at 8.1. For comparison, the 1931 Napier earthquake (Chapter 12) measured 7.8; the Chile earthquake of 1960 measured 9.4–9.6 (the largest ever recorded); the Anchorage, Alaska, earthquake of 1964 measured 9.2; and the Boxing Day 2004 Indian Ocean earthquake west of Sumatra, which caused a huge tsunami and humanitarian disaster, measured 9.3, making it the second-largest earthquake ever recorded. The Richter scale is logarithmic, so that a jump from one level to another, e.g. from 7 to 8, represents a tenfold increase in the energy released.

The effects of the 1855 earthquake were dramatic. The fault that ruptured was the Wairarapa Fault (Fig. 13.1), which forms the eastern margin of the Rimutaka Range. This fault fades away south of Lake Wairarapa, where its role is taken over by a thrust fault, the Wharekauhau Thrust. A thrust fault takes up tectonic shortening and causes uplift. During the earthquake, the Rimutaka Range

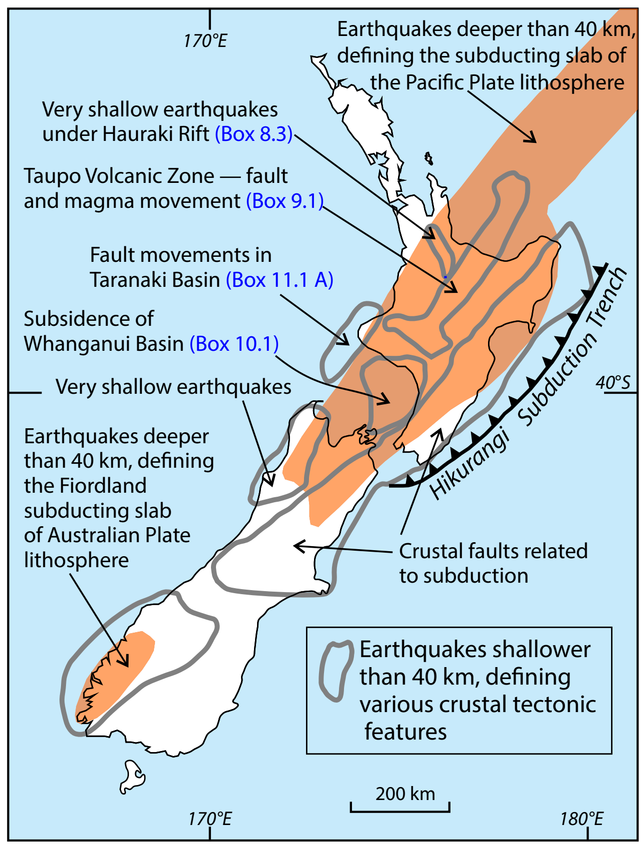
Box 13.2 A. Earthquakes and the North Island subduction zone.

An earthquake is a sudden release of energy into the earth's crust, causing vibration, shaking and — most destructive to property and hence human life — the passage of waves across the earth's surface. Some earthquakes are accompanied by visible and measurable deformation of the earth's crust —uplift, subsidence, breakage and offset along fractures (see Boxes 12.4 and 13.3 A, B for examples). Other earthquakes are not accompanied by ground-breaking deformation.

Source of earthquake energy
Most earthquakes occur in the regions where tectonic plates interact with each other (Boxes 2.1, 2.2). Individual plates move at speeds of between one and about twelve centimetres per year, so that where two plates converge at a subduction zone, and one plate either passes beneath the other (North Island situation) or rubs alongside the other (South Island situation), the relative movement between the two plates can be twice that, i.e. up to 25 cm per year. The fastest plate convergence so far recorded by global positioning satellite data is 24 cm per year, at the northern end of the New Zealand-Kermadec-Tonga subduction zone. 23 cm per year amounts to 230 kilometres in a million years; geologically that is an extremely rapid process.

Accumulation of stress, and creep versus breakage
Interactions between plates at those rates, even at the lower rate of less than 10 cm per year that is calculated (not measured yet) for the North Island, generate large amounts of frictional energy. We know from observations of fault creep (on the San Andreas Fault, California, and from long-term (50-100 years) surveys of deforming triangulation networks in New Zealand, that in some places and at some times the frictional energy is absorbed progressively, and there are few earthquakes. However, more commonly friction prevails and the plates lock together. The energy in the rocks accumulates, and is released periodically in an earthquake.

DISTRIBUTION OF EARTHQUAKES IN AND UNDERNEATH NZ.



Ground deformation
The sudden jolt of movement between the plates can be expressed at the surface by ground deformation. For example,
(a) the 1931 Napier earthquake was accompanied by folding of the Earth's surface in response to reverse slip on a west-dipping thrust fault (Box 12.4).

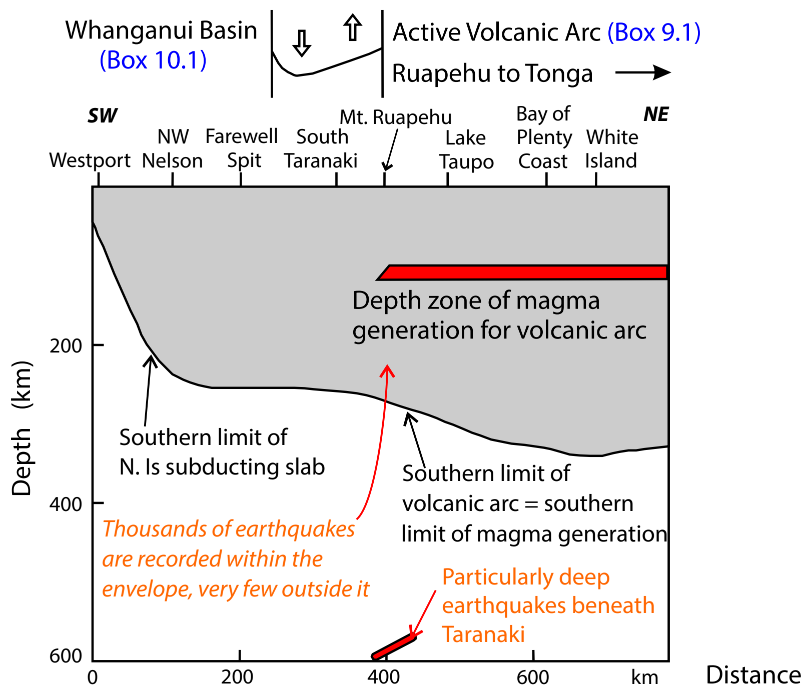
(b) the 1855 Wellington earthquake, which was enormous, was accompanied by up to 6.4 m of uplift and up to 18 m of right-lateral sideways displacement on the Wairarapa Fault (Box 13.3 B). The entire southwestern corner of the North Island was uplifted and tilted to the northwest.

(c) the 1987 Edgecumbe earthquake was a shallow event associated with the widening of the Taupo Volcanic Zone. There was subsidence of the Zone and ground breakage along the fault forming the southeastern margin of the Zone (Box 9.1). The movement in this case was 'normal', without any sideways motion, and resulted in one metre of extension. In other words, the Taupo Volcanic Zone (and hence the North Island) became one metre wider in the vicinity of Edgecumbe during the 1987 earthquake.

(continued on next page)

Box 13.2 B. Earthquakes and the North Island subduction zone.
(Continued from previous page. More cross sections on next page.)

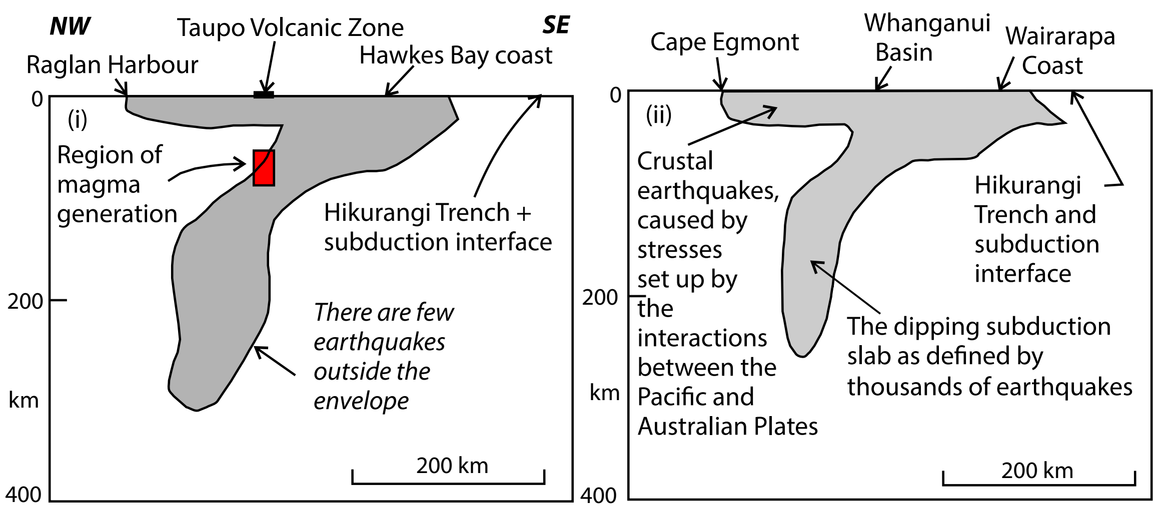
Length scales and time scales.
These effects of accumulation of stress, creep and breakage in earthquakes occur independently in different segments of the plate boundary. The effects of the 1855 earthquake were confined to central New Zealand, and, huge though it was, there was displacement only on the Wairarapa Fault.
The Wellington Fault did not move. Segments are typically no more than 100 km long.
Within individual segments — which are not clearly defined — the recurrence interval for major earthquakes is a few hundred years, according to several recent studies of the frequency of movement on major faults.



DEPTH DISTRIBUTION OF EARTHQUAKES BETWEEN WESTPORT AND BAY OF PLENTY

The earthquakes tell us the extent of the subducting slab of Pacific Plate lithosphere that dips to the north-west and underlies the North Island. At the southern limit of the slab, motion between the Pacific and Australian Plates changes from subduction to sideways slip along the Alpine Fault.

Box 13.2 C. Earthquakes and the North Island subduction zone.
(Continued from two previous pages.)

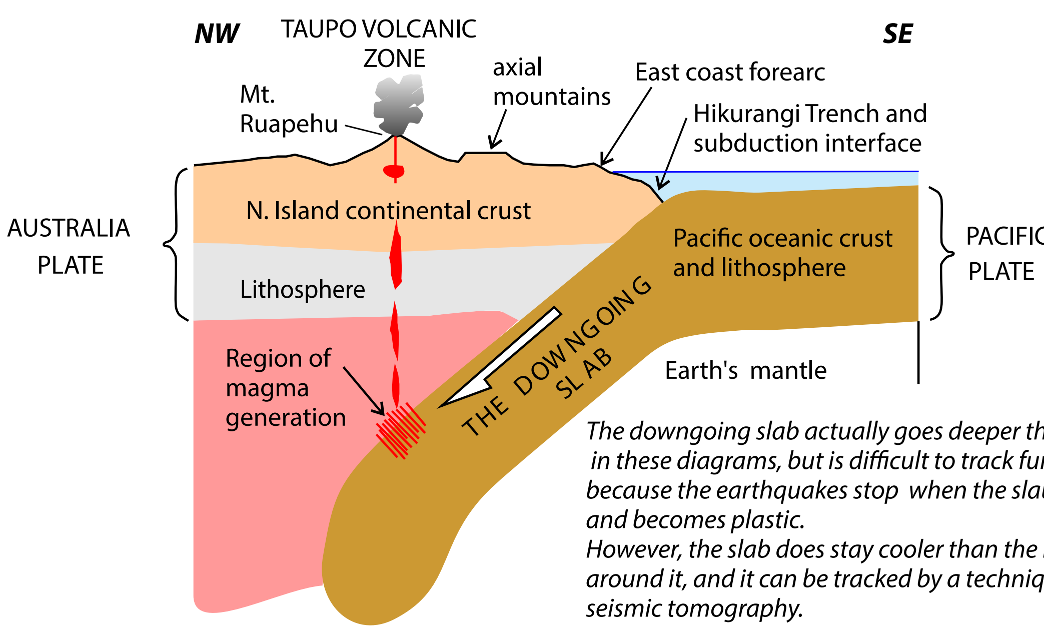


Two vertical sections across the North Island subduction zone showing the shape and extent of the downgoing slab of Pacific Plate lithosphere, as defined by the occurrence of thousands of earthquakes.

(i) Raglan Harbour to southern Hawke Bay, through the active Taupo Volcanic Zone

(ii) Cape Egmont to the Wairarapa Coast. Despite the fact that the slab has penetrated well past the critical depth for generation of magma (about 100 km), there is no indication of volcanic activity. Instead there is the Whanganui Basin. (Box 10.1)

SCHEMATIC CROSS SECTION THROUGH THE NORTH ISLAND SUBDUCTION ZONE



The downgoing slab actually goes deeper than shown in these diagrams, but is difficult to track further down because the earthquakes stop when the slab warms and becomes plastic.
However, the slab does stay cooler than the mantle around it, and it can be tracked by a technique called seismic tomography.

adjacent to the two faults rose by at least 6 m, as measured by uplift of the gravel beach ridge and its accompanying barnacle-encrusted rocks at Turakirae Head (Fig. 13.3). The range was also displaced northward relative to Lake Wairarapa (i.e. in a right-lateral or dextral sense) by up to 18.7 m. Long-term average rates of movement here are about 3.5 mm/year of uplift and 11 mm/year sideways. Both rates are higher than those along the Wellington Fault, and the ratio is different as well – there is more uplift per millimetre of sideways movement on the Wairarapa Fault, which explains why the Rimutaka Range is higher than the hills west of Wellington.

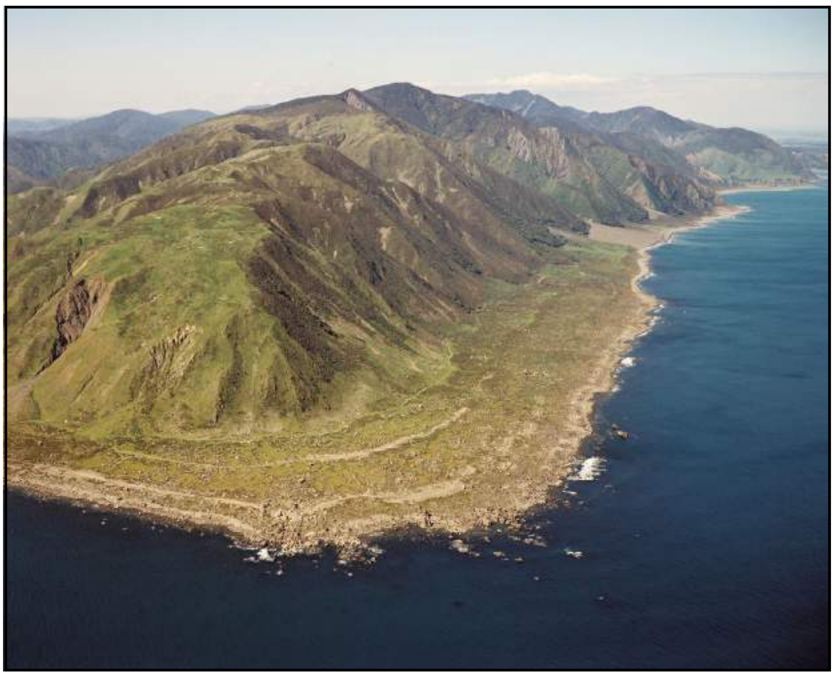


Fig. 13.3. The low coastal terrace around Turakirae Head has been uplifted during ruptures on the Wairarapa Fault in the last 7000 yrs. The lines visible on the terrace are a succession of five beach ridges, all but the lowest of which were uplifted by major earthquake events. The lowest ridge is not earthquake-related – it is the modern storm beach. Photographer Lloyd Homer, GNS Science.

The earthquake recurrence interval on the Wairarapa Fault is estimated to be between 1040 and 1420 years, compared with 715–1575 years for the Wellington Fault. A longer interval combined with higher displacement rates equates to more powerful earthquakes for the Wairarapa Fault – they are predicted to occur at magnitudes of 8.0–8.3 (as experienced in 1855), compared with 7.3–7.9 for the Wellington Fault.

Effects of the earthquake in Wellington

As a result of the 6 m of uplift at the eastern margin of the Rimutakas, the entire Wellington region (most of the area in the southern half of Fig. 13.1) was tilted to the northwest. The uplift diminished to zero at about the present west coast. The effect around Wellington city was an uplift of about 1.5 m, which was enough to raise the wave-cut intertidal rock platform above sea-level. This high and dry platform can be seen today at such places as the western headland of Lyall Bay and at Taputeranga Island in Island Bay. It has proved to be a highly convenient surface on which to build roads and railways. This is why Wellington has so many waterfront roads, which in most coastal cities are very difficult to build.

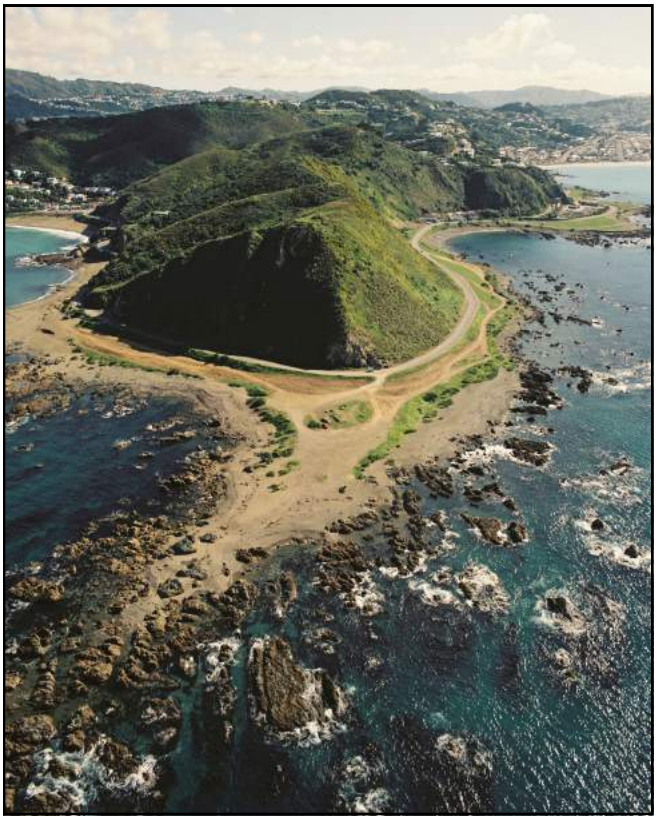


Fig. 13.4. The western headland of Lyall Bay, Te Raekaihau, Wellington, is surrounded by a low coastal terrace. This is a former intertidal platform that was uplifted by 1.5 m during the 1855 Wellington Earthquake.
Photographer Lloyd Homer, GNS Science.

Tectonic Landscape Capital

Wellington's faults have a major influence on the landscape, hence its claim as the tectonic landscape capital of the world. The Wellington and Wairarapa faults are just two of many that comprise the Wellington–Hutt Valley portion of the North Island Shear Zone, the belt where most of the sideways displacement across the North Island subduction zone is concentrated (Chapter 12). In total, there are five major active faults (Fig. 13.1) and a large number of minor ones. Together, they form a pattern that is characteristic of right-lateral transcurrent (i.e. predominantly sideways-moving) fault systems.

Fault pattern

The five dominant, through-going, right-lateral faults trend northeast. They are parallel to the plate boundary, which lies about 100 km to the southeast, at the bottom of the Hikurangi Trough (Box 2.3). A larger number of lesser faults trend north to north-northeast, as shown in Fig. 13.5. Generally, they do not cross the major faults, but instead join them at an acute angle. In theory, they should also have right-lateral displacement, though this is hard to prove because of a lack of distinctive markers that can be matched across them. Some of the lesser faults are shown in Box 13.1 A – these particular ones all join the Wellington Fault, either within the city or in the harbour. The geological map of Wellington also shows a few faults trending northwestward; in theory, these should have the opposite, left-lateral, sideways displacement, but, like the minor faults above, this is hard to prove.

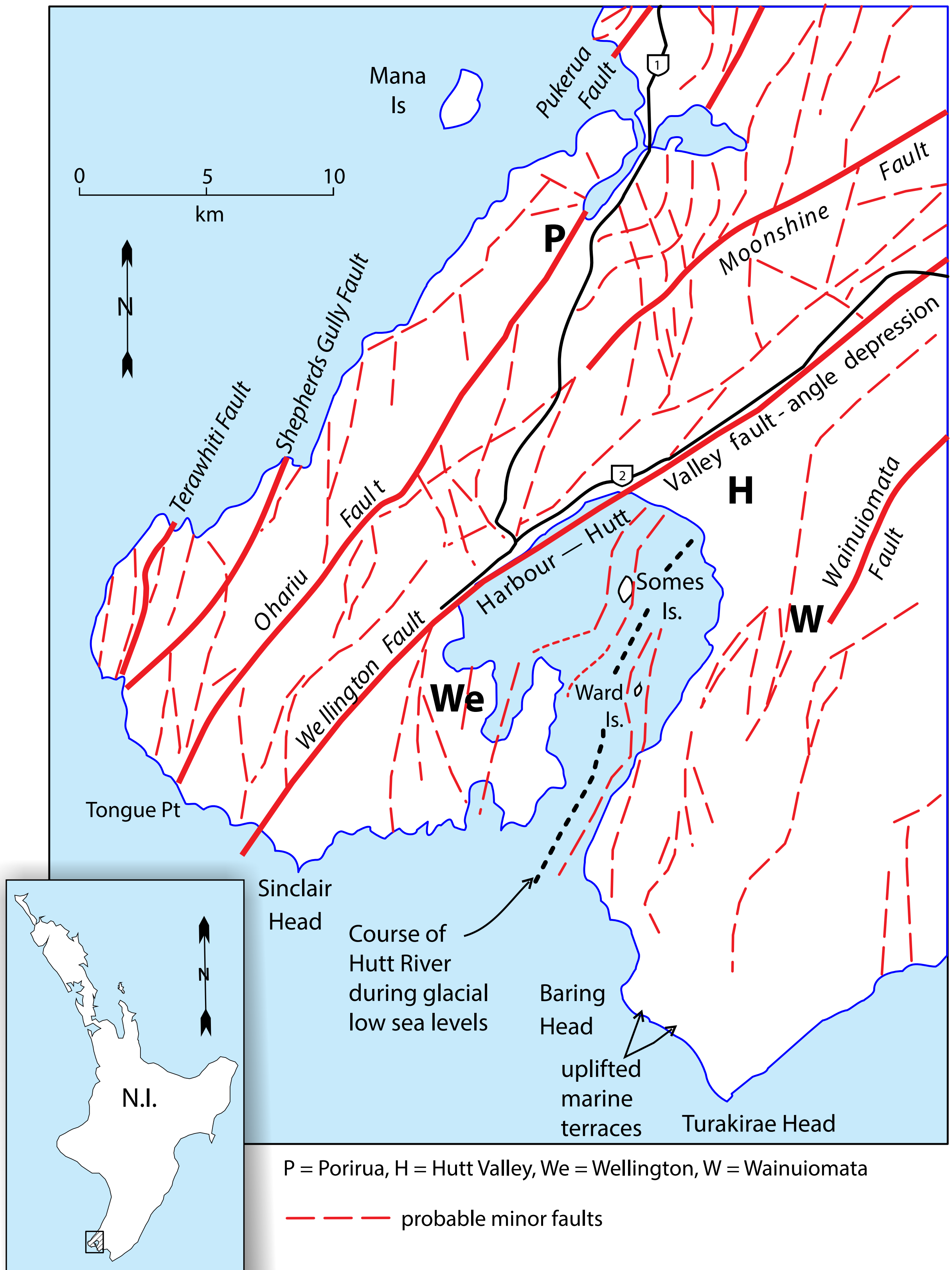


Fig. 13.5. Fault pattern in the Wellington Region.

The general braided fault pattern shown in Fig. 13.5, with major faults parallel to the direction of offset (i.e. trending northeast–southwest) and minor faults between them forming acute angles pointing north and south, is typical of right-lateral fault systems worldwide.

Vertical fault movements

The other important component of displacement on Wellington's faults aside from sideways movement is up and down movement. Few faults have 'pure' sideways motion – there is almost always another motion factor, tending either to pull the two sides of the fault apart or to push them together. It is this other factor that causes the up and down motion, which may be either 'normal' (extensional, pulling apart and causing a lengthening of the terrain) or 'reverse' (compressional, pushing together and causing a shortening of the terrain).

Reverse faults

The Wellington region is adjacent to an obliquely compressive plate boundary. That is, the Pacific Plate is subducting underneath the Australian Plate along the eastern seaboard, causing compression in the vicinity of the plate boundary. However, the Pacific Plate is moving towards the southwest and approaching the Australian Plate at a map angle of about 45° to the plate boundary (Box 2.3). This is the oblique component that gives rise to the right-lateral nature of the North Island Shear Zone. From what we have learned so far, in a right-lateral, obliquely compressive fault regime, we should expect the up and down component of the faults to be reverse (shortening) in nature.

And this is exactly what we find in Wellington. All the faults here are either vertical (i.e. neutral in terms of normal or reverse separation), or non-vertical and reversed. Furthermore, as you would expect, there is a gradient across the region in the degree of reverse displacement. Faults nearer to the subduction trench, where compression is strongest, tend to have the most marked reverse aspect (i.e. they tend to dip at a low angle), while those further away have a smaller reverse aspect and tend to dip more steeply.

Thus, faults in southern Wairarapa, the southeast corner of the North Island, closest to the plate boundary, tend to be reverse in nature. They dip to the northwest at angles of 60–70° from the horizontal. The Wharekauhau Thrust (a thrust is a low-angle reverse fault), shown on Fig. 13.1, carries the Rimutaka Range up and over the forearc basin of Lake Wairarapa and Palliser Bay. It dips underneath the Rimutakas at an angle of about 45°. Around Wellington, further away from the plate boundary, the faults are steeper – mostly near-vertical.

Upthrow and downthrow

With compression coming from the east, pushing against a wide backstop of continental crust west of Wellington, faults are overthrust back towards the east. This determines that the upthrown side of the fault, which is important in terms of effects on the landscape, is usually the western side. So, getting back now to Wellington's tectonic landscape, we can see its fabric is determined by a large, homogenous rock mass (greywacke) that is carved up by northeast-trending and north-trending faults that are generally upthrown on the west side.

Faults usually follow valleys, for two reasons. First, currently active faults like those around

Wellington physically push some ground upwards and other ground downwards, creating a fault-angle depression (Box 6.6 G). The fault scarp forms one side of the valley, and the fault is located in the bottom of the valley, which is, of course, where the river (e.g. the Hutt River) is going to flow.

Second, fault movement grinds up rocks, forming a zone of minced-up 'fault gouge' that can be many metres wide. Fault gouge is weak, and erosional forces excavate it preferentially, forming valleys. That is why major faults are seldom exposed to view. As it happens, the fault gouge of the Wellington Fault is exposed in the right bank of the Hutt River, about 1 km north of Silverstream Bridge and directly north of a prominent knob of greywacke called Mains Rock, which has a cliff against the river.

The Wellington region's pattern of minor north-trending and major northeast-trending ridges and valleys shows clearly on topographic maps and on high-altitude aerial photos (e.g. Fig. 13.6). Because of the general sameness (some would say monotony) of greywacke, and because fault planes are seldom exposed, it is often impossible to prove that a fault is present in a given valley. However, if the valley trends north or northeast there is a strong likelihood that a fault is indeed involved.

The Wellington Fault

The Wellington Fault dominates the Wellington–Hutt Valley landscape. It can be seen from satellites and it figures prominently on aerial photographs (e.g. Fig. 13.6). The reason the fault is so obvious is, first, that it is close to vertical and therefore forms a long, straight valley. Second, it has a strong component of western uplift and eastern downthrow, thereby creating the asymmetrical Hutt Valley and inner part of Wellington Harbour (Port Nicholson) as a fault-angle depression.

Fig. 13.6. Looking north along the line of the Wellington Fault, running straight through the Karori Reservoir and the west (left) side of the downthrown Wellington Harbour and Hutt Valley beyond. Photographer Lloyd Homer, GNS Science.

While the western side of the fault is uniformly high-standing owing to uplift, the eastern side has not been uniformly downthrown. In part this is because of intersecting north-trending lesser faults. Thus, Upper Hutt, Lower Hutt and the inner harbour mark three areas of strong downthrow, allowing the Hutt River to form extensive alluvial flats on which the towns are built. Separating these areas are bedrock highs, which mark areas of lesser downthrow: northeast of Upper Hutt; between Silverstream and Taita (Taita Gorge); and southwest from downtown Wellington to the coast of Cook Strait.

In other words, the bedrock surface on the downthrown side of the Wellington Fault undulates, with a wavelength of 10–20 km. This pattern can be seen on Figs 13.1 and 13.5. The maximum known depth to bedrock in the fault angle is more than 600 m, beneath the inner harbour. The outer harbour, meanwhile, contains several fault-controlled NNE-trending ridges and hollows.

Fig. 13.7. View south along the Wellington Fault and Hutt Valley showing the Upper Hutt depression filled with alluvium (foreground) separated by a forest-covered structural high of basement greywacke (Silverstream-Taita) from the Lower Hutt depression beyond. Note the active trace of the Wellington Fault in the foreground.
Photographer Lloyd Homer, GNS Science.

Tilting of the bedrock surface into the fault angle (i.e. towards the northwest) has had the effect of pushing the Hutt River up against the fault escarpment for most of its length (Box 6.6 G). In fact, the erosion associated with having the river hard up against the scarp has caused the scarp to retreat from the actual position of the fault in places, particularly along the inner harbour edge, where it is several hundred metres west of the fault.

East of Upper Hutt, the Wainuiomata-Mangaroa Fault complex (an association of lesser faults rather than a single major fault like the Wellington Fault) has created an alluvium-filled fault-angle valley. This is a mirror image of the Hutt Valley, and contains the Mangaroa River and settlement.

Origin of Wellington Harbour

As noted above, Wellington Harbour (Port Nicholson) occupies one of the more depressed areas on the downthrown side of the Wellington Fault. Within the harbour, Somes and Ward islands sit on separate fault-bounded, north-trending, structural highs (Fig. 13.5) . Somes may link up with the prominent ridge that bounds Evans Bay, Lyall Bay and the airport (according to Maori traditional history, that ridge itself was an island, before an earthquake struck in the 1400s).

Between these two greywacke ridges lies a fault-bounded depression that forms the harbour entrance. This depression would have been the valley of the Hutt River during the low sea-levels (up to 130 m below present) experienced during glacial periods (Chapter 10). In fact, the final stretch of the Hutt River still follows the depression to the harbour shore. The effect of this is to move the river away from the Wellington Fault in the northern corner of the harbour, something that roading engineers and transport planners are no doubt thankful for.

The last glaciation reached its maximum extent 20,000 years ago, and during it and all earlier glaciations Wellington Harbour was dry land. The harbour is surprisingly shallow – it does not reach 30 m. So, for that matter, is western Cook Strait, which is less than 100 m deep through the South Taranaki Bight, with the result that there was a dry land connection between North Island and South Island during each glaciation (Chapter 14).

Terraces, faults and rates of movement

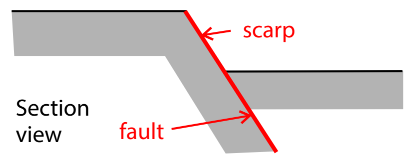
As explained in Box 10.3, terrace landforms are portions of river floodplains or coastal rock and sand platforms that have been left high and dry by a drop in the local base level of erosion. Various factors can cause the drop, but in the Wellington area the cause is almost always tectonic uplift.

The combination of river terraces (which can sometimes be dated, e.g. by radiocarbon or other radiometric means) and an active fault gives rise to the possibility of working out the movements on the fault over the past few tens of thousands of years. This is explained in Box 13.3 A, with the Wellington Fault at Emerald Hill, near Brown Owl, northeast of Upper Hutt, given as an example. The rate of uplift on the north side of the fault here is less than one-tenth of the right-lateral sideways movement, which has a long-term average rate of 6–7 mm/year.

Box 13.3 A. Active faults (earthquake faults).

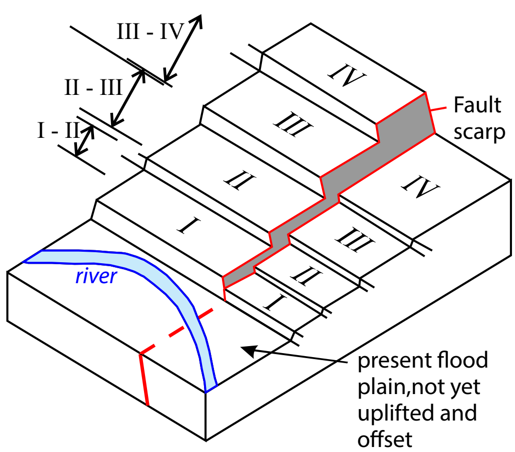
The name signifies a fault that has ruptured the ground surface in the recent past - recently enough that erosion has not destroyed the landform. Despite being characteristic of New Zealand's earthquake belt, they are not easy to see on the ground. It is generally not possible to see them from a moving vehicle.

Simplest Case —
flat ground offset by 1 m to several m.
Offset only vertical
Scarp straight.
Fault vertical or inclined.

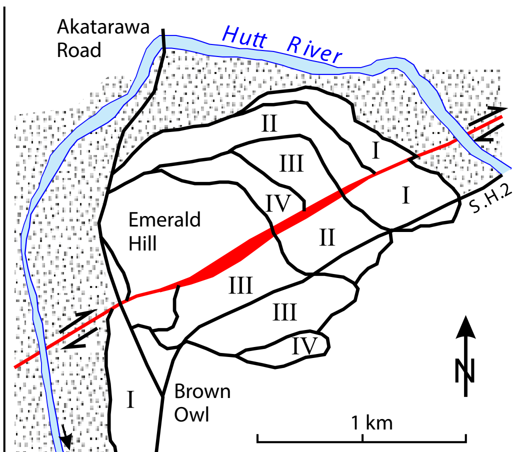


Complication—
the landform is very similar to a river terrace. It needs careful examination. However, the case where a recent fault displaces terraces is very informative, because higher (older) terraces may be offset more than lower (younger) terraces, and because terraces can be dated we can derive a history of fault movement over some movement over some tens of thousands of years.

Second complication—
in many cases offset on the fault is oblique, ie. a combination of vertical and sideways movement. To determine the sideways offset is difficult, unless linear features like streams, ridges, or terrace risers are displaced.



Idealised diagram, showing four river terraces (IV oldest, I youngest) offset by progressively greater amounts. Straight terrace risers give a measure of right-lateral sideways displacement, which is greater than the uplift. The double arrows show the cumulative oblique displacement.



The Wellington Fault at Emerald Hill near Upper Hutt.
Terraces cut by the Hutt River are numbered the same as the left diagram. Their ages and sideways displacements are as follows:

Dotted	Present floodplain 0-c.12,000 years	a few metres
I	14,000 years	100 m
II	70,000 years	440 m
III	140,000 years	940 m
IV	190,0000 years	not measurable

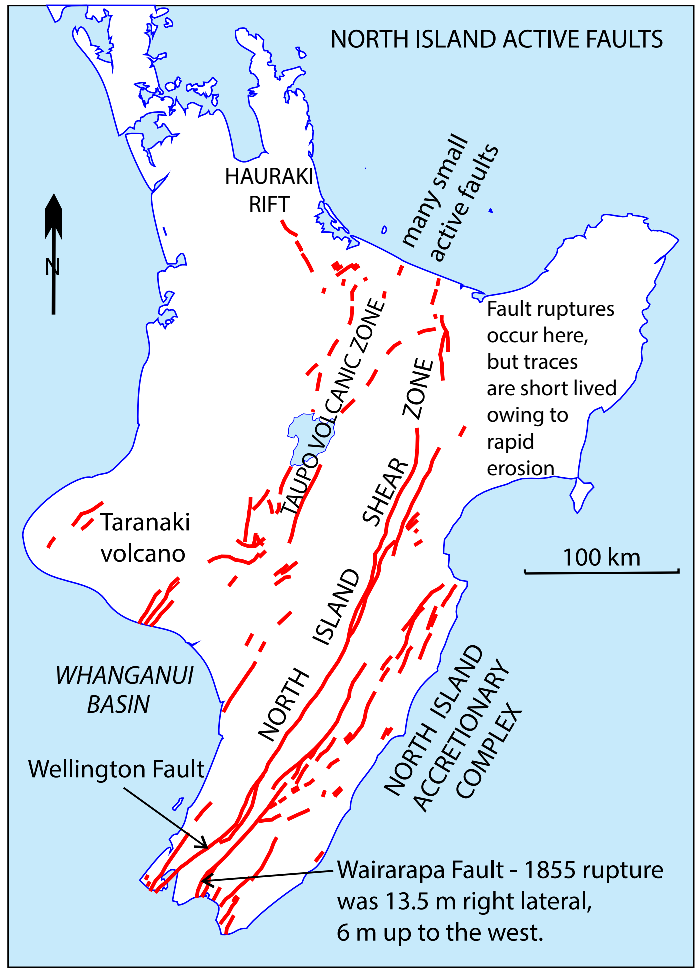
Uplift is on the north side, but is less than 10% of the sideways offset. The latter, like most New Zealand active faults, is right lateral (dextral) - the opposite side of the fault moves to the right.

Box 13.3 B. Active faults – earthquake scenario.

Earthquake scenario

From data like that presented, from precise survey data, and from examining trenches dug across fault planes, the following earthquake scenario has been derived for the Wellington Fault in the Hutt Valley:

1. Long-term sideways displacement averages 6-7 mm/year
2. An earthquake of magnitude 7.5 is likely to occur about every 1145 years, and to cause 4 m of right-lateral sideways displacement
3. Gradual sideways shear affects a zone about 7 km wide, centred on the Wellington Fault. Between 1929 and 1969 there was a total right-lateral offset of 170 mm
4. Earthquakes occur when the cumulative plate tectonic stresses exceed the capacity of the rocks to absorb them.



In the case of coastal terraces, there are good examples on Wellington's south coast at Baring Head and at Tongue Point (Fig. 13.1). They give information on uplifts over the past 300,000 years. Each terrace represents coastal erosion and cliff formation during an interglacial high-stand of sea-level, which occurred at approximately 100,000, 200,000 and 300,000 years ago (and before this, although these terraces are not preserved here any more). Being located on opposite sides of the Wellington Fault, the two localities have different uplift records, as might be expected. At Baring Head, the long-term uplift rate (350 m in 300,000 years) is approximately 1 mm/year, while at Tongue Point (40 m in 100,000 years) it is 0.4 mm/year. The latter rate is in the same ball park as Emerald Hill, where it is about 0.5 mm/year.

Subsidence rates

The mirror image of uplift is subsidence in the fault angles. This can be measured by obtaining the sediment accumulation record from boreholes – in the Hutt Valley there is good information from water bores. The deepest part of the Hutt Valley sector of the fault angle (i.e. depth to bedrock) is 480 m below sea-level. The oldest (i.e. deepest) sediments have been dated to about 550,000 years, giving a long-term average subsidence rate of about 0.8 mm/year. In global terms, this is a high rate.

History of the Wellington Fault

The age of the deepest part of the Hutt Valley fault angle is thought to mark the beginning of movement on the Wellington Fault. If that is the case, the fault has been in existence for about half a million years. A simple (and simplistic) calculation, putting together the two figures of age and recent rate of right-lateral movement at Upper Hutt (6–7 mm/year), would indicate a total right-lateral displacement on the Wellington Fault of about 3.25 km. The uncontrolled assumption in this calculation is that movement has always been at the same rate. There is no way of confirming that assumption, but by matching up a pair of valleys that seem to have been offset by the fault (Island Bay to Porirua, and Evans Bay to Korokoro Stream), a displacement of 4–5 km is suggested, which is close enough to the other figure to suggest that it may be right.

Fleshing out the Structural Bones

Wellington is renowned for steep slopes, and for the ingenuity of house and road construction they have spawned. Given the uplift that is taking place at geologically rapid rates, the presence of these steep slopes is understandable. Rivers and streams are given energy by uplift, and so deepen their valleys to form a graded profile to sea-level. As noted above, at the height of the last glaciation, 20,000 years ago, sea-level stood 130 m below its present level. Thus rivers at the time were working to a much lower base level and had correspondingly greater energy. In turn, deepening river courses provide gravitational energy for gravity-driven slope-forming processes, as outlined in Boxes 12.5 A–C.

A fine-textured landscape

Given the overall uniform nature of greywacke rocks and their high susceptibility to chemical weathering (Box 6.1 B), the actual slope-forming processes work mostly on a surface mantle of clay. Clay is uniform in character and impervious to water. The combination of deeply excavated valleys, moderate to high rainfall that cannot soak in very far, and uniform substrate, gives rise to a fine-textured landscape of small ridges and valleys that dissect the major fault-determined blocks. Within the parameters set by climate and rock type, the retreat of slopes takes place uniformly. This means that the slopes remain constant, and where they intersect others retreating from different directions, they form sharp-crested ridges. The overall effect is to create regularly spaced herringbone patterns of ridges.

Effects of glaciation

The last glaciation, 20,000 years ago, enters the story yet again. Apart from small ice caps on the highest Tararua Ranges, the Wellington region was not covered in ice. However, there was no forest cover either, and there was plenty of rain and snow, and much freezing and thawing of the surface rocks. These factors, in combination with frequent earthquakes, gave rise to the formation of extensive screes and glacier-like flows of mud and rock rubble (so-called solifluction debris). Deposits of this material can be seen in road cuttings near the foot of slopes. They consist of angular boulders and smaller rubble in a stony clay matrix. There were earlier glaciations at approximately 120,000 and 220,000 years ago, and more before that, so it is scarcely surprising that the landscape is complex.

Floodplains

All slope-forming processes deliver rocks and mud to rivers, which transport them to the sea. Along the way, river sediment is stored temporarily in floodplains. If tectonic subsidence is taking place, as along the Wellington Fault, this storage can be long term – up to 500,000 years. If localised tectonic uplift is taking place, river terraces result. The river sediments stored in the Hutt Valley fault-angle depression are very important water-bearing aquifers.

The K erosion surface

In Chapter 4 we described the flat to gently undulating surface of erosion that was carved across the basement rocks of most of the country, between 100 myr and 25 myr ago. In most places this erosion surface was eventually buried beneath sedimentary rocks, especially the Te Kuiti/Waitomo limestones (Chapter 7). Increased tectonic activity from 25 myr ago has affected the surface, folding it, burying it underneath younger rocks in places, and uplifting it, exposing it and causing it to be destroyed by erosion in others. The Wellington region, as we have noted, is a terrain of pervasive uplift and erosion, and so there are very few remnants of the sedimentary rocks that originally covered the surface, aside from a small area near Paraparaumu.

However, a different erosion surface, known as the K surface, has survived quite widely in the Wellington region. This surface was named by the late Professor Sir Charles Cotton, the noted Wellington geomorphologist and author of many books on Wellington and New Zealand landforms. The 'K' is short for both Kaukau, a peak in the Wellington suburb of Johnsonville, and key, because it enables us to track the region's recent uplift and subsidence, from –600 m in Wellington's inner harbour to +450 m in places west of the Wellington Fault, and to +800 m and more in the Rimutaka Range. The K surface is therefore not horizontal, because of all the recent tectonic activity, but it can be recognised from accordant ridge-top levels and from tilted areas of gently undulating ground lying between the steep valleys that are eating into it. As you would expect, the surface is preserved most widely where the uplift rates are lowest. The suburbs of Paparangi and Newlands, east of Johnsonville, are built on an extensive remnant of the surface, while Johnsonville itself is built on young alluvial sediments that mantle the surface in places.

The actual age of the K surface is not known – because of the general lack of young sedimentary rocks in the region, there are few clues. The analogy made above with the rest of the country would suggest it is the pre-25 myr-old surface. However, the paucity of remnants of the rock sequence that would have overlain it opens up the possibility that it records instead a prolonged period of subaerial erosion that pre-dated the beginnings of tectonic uplift about 1 myr ago.

It is important to note that the tectonic deformation we have concentrated on in this chapter, dating from the last 1 myr, has not only created fault ruptures but also folded the region into broad anticlines and synclines (Box 6.6 A). However, the highly complex small- and medium-scale fold structures that you see in exposures of greywacke are much older, and date from accretionary wedge days, around 120 myr ago (Chapter 12).

Fig. 13.8. The flat surfaces and flat-topped ridge crests in this southeastward view, from south of Porirua across to the head of Wellington Harbour, are the dissected remnants of the K erosion surface. Photographer Lloyd Homer, GNS Science.

PART 3: REGIONAL GEOLOGY OF THE SOUTH ISLAND

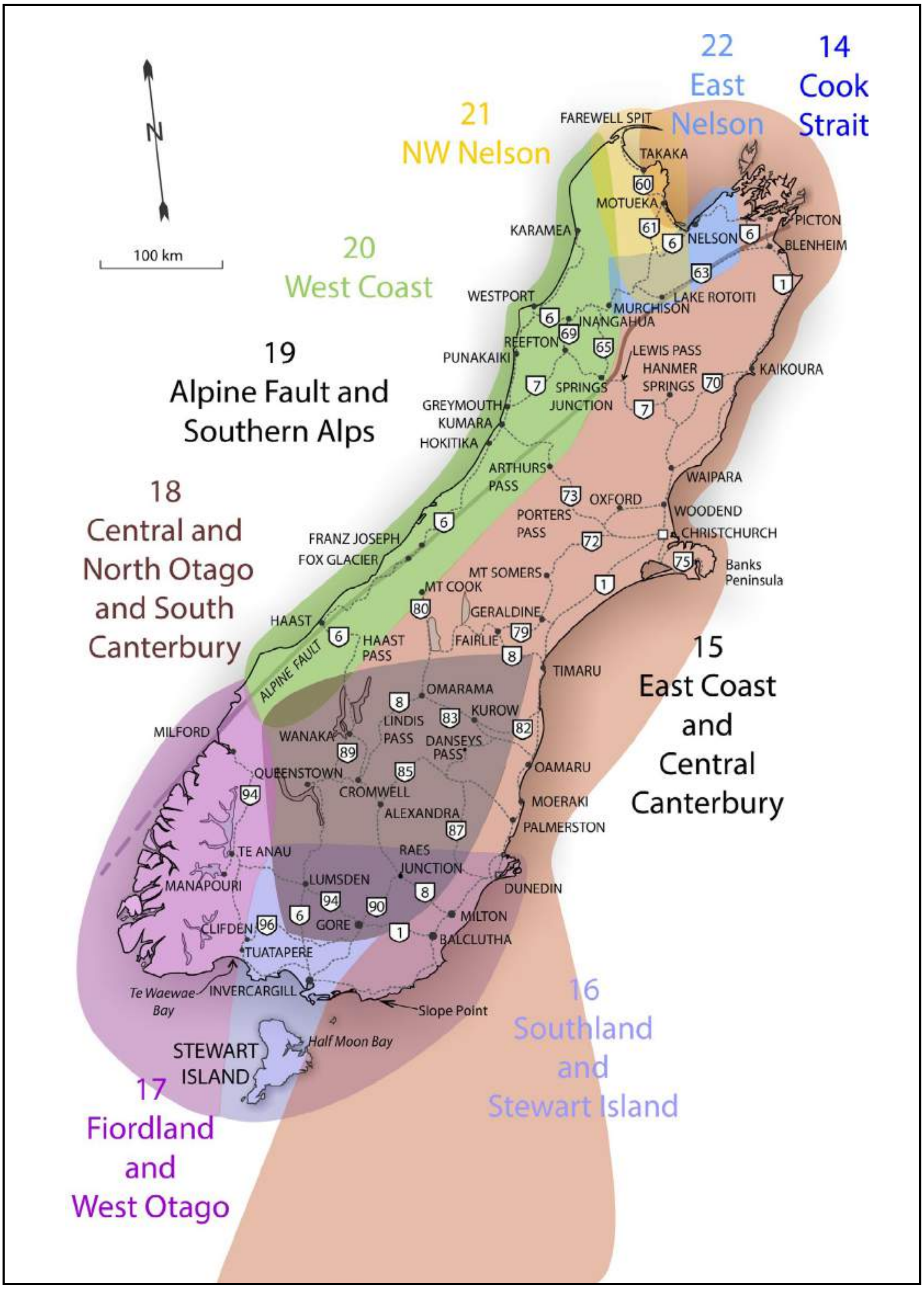


Fig. 14.0. South Island of New Zealand, showing the area covered by each chapter.

Chapter 14

Origin of Cook Strait

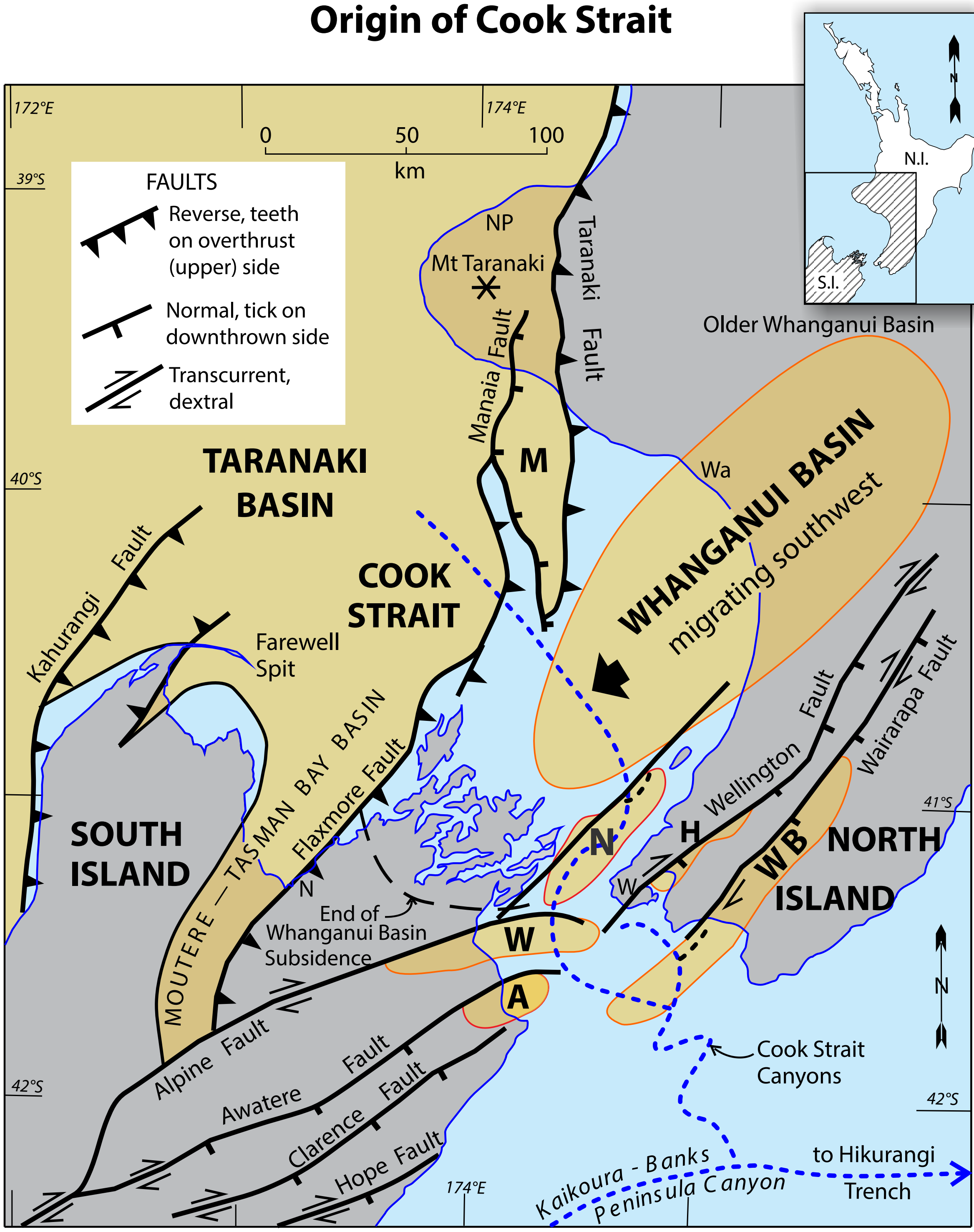


Fig. 14.1. Origin of Cook Strait – components map.
NP = New Plymouth, Wa = Whanganui, N = Nelson, W = Wellington.
Bold letters denote some of the basins - **M** = Manaia, **N** = Narrows, **H** = Wellington Harbour-Hutt Valley, **W** = Wairau, **A** = Awatere, **WB** = Wairarapa.

Cook Strait is a major Z-shaped strait separating North Island from South Island, and is a logistical barrier of the first order. However, in geological terms it is a linking feature, connecting the subducting plate boundary of the North Island with the transform boundary, through continental crust, of the South Island. The strait is actually there because of a transitory set of geological circumstances.

Because it has always been such a dominating factor in human affairs, we tend to assume Cook Strait has always been there. In fact, it is quite a recent phenomenon, a chance alignment of a number of tectonically controlled sedimentary basins. Looking back to how much has happened in the Wellington–Hutt Valley region over the past 1 myr (Chapter 13), it shouldn't be too surprising that something as major as Cook Strait could be even younger than that. Furthermore, it trends broadly northwest–southeast, i.e. at right angles to all the tectonic trends. This in itself indicates there is something special about Cook Strait – it cuts across all the major geological structures.

Basin Line-up

In a nutshell, right-lateral sideways shuffling along several faults of the North Island Shear Zone (Chapter 12), and its continuation in the South Island's Alpine Fault and Marlborough Fault Zone (Chapter 15), has brought into chance alignment no fewer than eight sedimentary basins, as shown in Box 14.1. They are, from west to east: Taranaki, Moutere–Tasman Bay, Manaia (indicated as 'M' on the map), Whanganui, Narrows ('N'), Wairau ('W'), Awatere ('A') and Wairarapa ('WB').

Taranaki Basin and its younger extension to the south, Moutere Depression–Tasman Bay Basin, are foreland basins, i.e. they were, or are, being caused by depression of the crust ahead of rocks being pushed westwards up the Flaxmore, Manaia and Taranaki thrust faults. They are a distant response to compression generated within the Australian–Pacific plate boundary zone.

Manaia Basin exists because of a divergence between the Flaxmore and Taranaki thrust faults, which has allowed a narrow arm of the Taranaki Basin to extend southwards. As the Whanganui Basin migrates southwestwards, pulling down the crust as it goes (Chapter 10), it has moved nearer and nearer to the Manaia Basin, which is older. This has created a breach in the ridge of high-standing basement rock that forms the upper plate of the thrust system. Without that breach there would have been a continuous physical barrier between the Whanganui and Taranaki basins, and possibly no Cook Strait.

Whanganui Basin is related in some way to processes taking place in the subduction zone (Box 10.1). It continues the line of the Taupo Volcanic Zone (TVZ) volcanic arc, and is migrating southwestwards in concert with the arc. The Marlborough Sounds are symptoms of the initial subsidence of the approaching basin.

Narrows Basin is the critical basin that links the other seven. It forms the portion of Cook Strait that is narrowest, and is aligned northeast–southwest, along the tectonic trend rather than across it – the central arm of the lazy-Z shape of Cook Strait. It and the Wairau and Awatere basins (as well as the Wellington Harbour–Hutt Valley Basin ('H' in Box 14.1), although this does not contribute to Cook Strait) are fault-angle valleys (Box 6.6 G) in the zone of right-lateral transcurrent faults. These mark the start of the transform portion of the plate boundary – the South Island Alpine Fault and its associated Marlborough faults (Awatere, Clarence and Hope) – and its overlap with the North Island Shear Zone.

Finally, at the southeastern entrance to Cook Strait is the Wairarapa Basin. This is the southernmost part of the North Island forearc basin (Chapter 12).

So, we have four different kinds of sedimentary basin that are all related to the Australian–Pacific plate boundary. They can all exist in the same region because this happens to be where the North Island subduction zone and the South Island transform boundary overlap (Box 2.3).

Basin Jostling and the Evolution of Cook Strait

Because they are contained between right-lateral moving faults, the three central basins (Narrows, Wairau and Awatere) are continually changing their positions relative to one another. It was during this jostling process that the Narrows Basin was brought into close proximity with the Wairau and Whanganui basins 1–0.5 myr ago, to form the vital central link joining the western and eastern parts of Greater Cook Strait.

The broad western section of Cook Strait, comprising bigger and, in part, older basins, has existed for longer. Even here, the geography 10 myr ago was such that large volumes of sand derived from the granite belts of northwest Nelson were being transported northwards into the Taranaki Basin (Chapter 11), indicating that the western Cook Strait embayment was not in existence at that time – if it had been, it would have trapped all of that sediment. Box 14.1 shows the evolution of Cook Strait, from the time before growth of the North Island mountains began 1 myr ago.

Cook Strait during glaciations

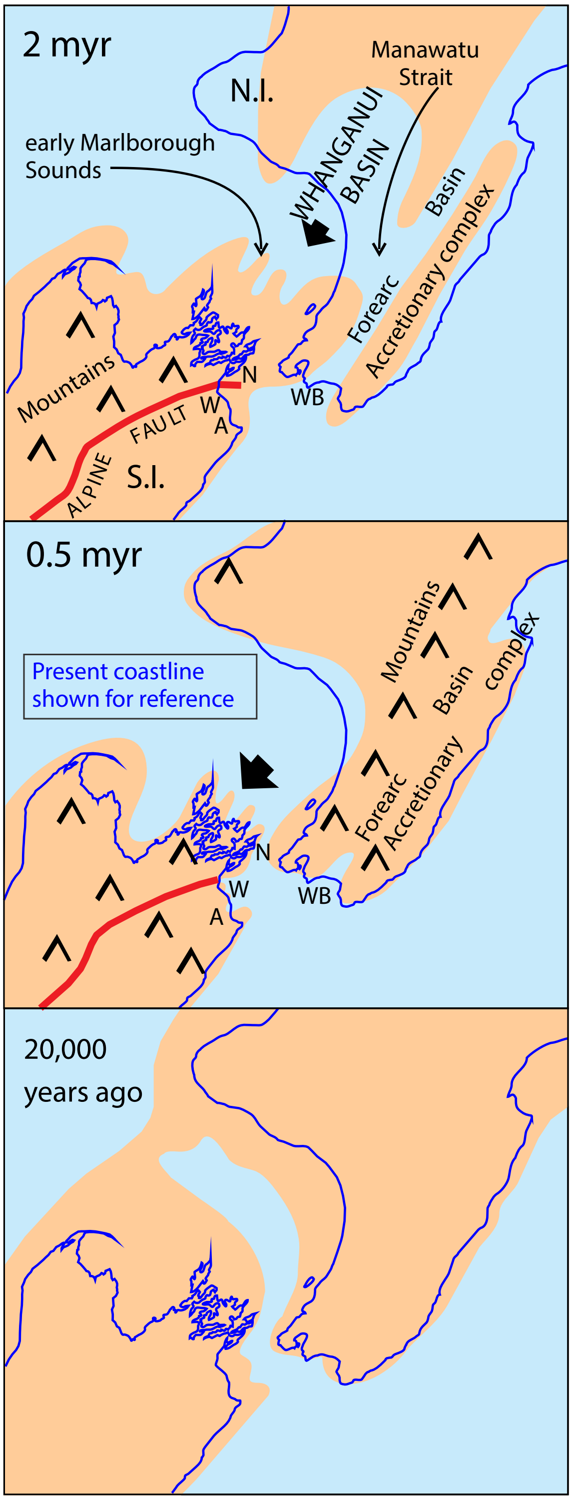
Western Cook Strait, between Farewell Spit and the Taranaki Peninsula, is shallow – less than 100 m. We know that during glaciations world sea-level falls by up to 130 m. We also know that glaciations happen approximately every 100,000 years, and that the last one peaked about 20,000 years ago (Box 10.2 A).

It is thought that the shallow western part of Cook Strait provides a land bridge during glaciations, connecting North Island with South Island. This has important implications, because it regularly renews genetic links of flora and fauna between the islands, but at the same time repeatedly cuts them off, allowing subspecies of snails, for example, to evolve in isolation. The lowest map in Box 14.1 shows Cook Strait during a glacial low level, about 20,000 years ago. We are presently in an interglacial high sea-level.

Tidal Flows and Currents through Cook Strait

Cook Strait presents a wide funnel-mouth at both ends and is flanked by mountains on both sides. It is located within the 'Roaring Forties', and therefore its western end funnels the prevailing westerly and northwesterly winds into the Narrows. Similarly, from the other side, eastern Cook Strait funnels southerly storm winds into the Narrows. Cook Strait and Wellington city and harbour are notorious for both northerly and southerly gales, as regular air passengers into Wellington and travellers on Cook Strait ferries know only too well. Big southerly swells generated by distant storms pound the Wellington coastline, and are likewise funnelled into the Narrows, where wave heights of 8–10 m are not uncommon. Gales from both north and south push water through the Narrows at speeds of up to 0.6 m/second. These flows can either reinforce, or counteract, tidal flows.

Box 14.1. Evolution of Cook Strait.



2 million years ago.
Whanganui Basin migrating south,
Narrows Basin forming and moving north.
No North Island mountains.

land

Half a million years ago.
Whanganui Basin and Narrows Basin have overlapped.
Manawatu Strait and Forearc Basin seaway closed by uplift.

Cook Strait during the last glaciation,
20,000 years ago.
Coastline at minus 130 metres.

There is a big tidal flow through the Narrows. Because the twice-daily tidal wave travels anticlockwise around New Zealand, the opposite ends of Cook Strait have – by chance – almost exactly opposite tides, so that when it is high tide at one end, it is low tide at the other. Thus there is a reversing tidal flow through the Narrows, which reaches speeds of up to 1.5 m/second at the seabed. This flow has scoured the bottom of the Narrows to a depth of 225 m – 100 m deeper than to either side. It also creates strong eddies around headlands, which scour local seabed holes to depths of 300 m. These tidal flows represent a large, albeit technically challenging, opportunity for electricity generation. They also make for challenging Cook Strait sea crossings!

The sediment that forms the seabed in the strait reflects the vigorous wave, current and tidal regime found here. It is mostly sand, although there is sandy mud and shell gravel in sheltered areas, including the Whanganui Bight and Tasman and Golden bays. There are also areas of gravel around the Taranaki Peninsula (from the volcanic gravels that form the peninsula) and through the Narrows (left behind after the winnowing effect of the tidal currents has removed sand and mud).

The Cook Strait Canyons

These are a set of submarine valleys that penetrate eastern Cook Strait and follow the Narrows Basin into western Cook Strait (Box 14.1). The shallow D'Urville sea valley extends the system well into western Cook Strait. The canyons have formed quite independently of the sedimentary basins. In association with the Kaikoura–Banks Peninsula canyon system, they are a headwards extension of the Hikurangi Trough, the subduction trench that lies alongside eastern North Island.

Submarine valleys form in similar ways to valleys on land. Gravity-driven slope-forming processes deliver sediment to the central valley, where it is removed downstream. In the case of land valleys, sediment is removed by a river, whereas in the case of submarine valleys, it is removed by periodic sediment gravity flows (e.g. turbidity currents – Boxes 6.5 A, B). The valleys extend themselves headwards.

The Cook Strait canyons cut across the sedimentary basins, and in the case of the Wairarapa Basin have cut deeply into it. Both the Cook Strait and Kaikoura–Banks Peninsula canyons play an important role in the sediment budget of New Zealand, as described below. The sediment budget is the broad picture of sediment production, sediment transportation and delivery of sediment to a long-term resting place.

Cook Strait and the New Zealand Sediment Budget

A landmass that is rising rapidly and has highly productive volcanoes and high rainfall generates large volumes of sediment. That's us. The amount of rock added to the South Island mountains each year by average uplift rates of 8–11 mm is matched by the amount of fractured and weathered rock removed by erosion and delivered to the sea, perhaps after temporary storage along the way. Extra-large slugs of gravel produced by landslides take only a few years or decades to move down to the sea.

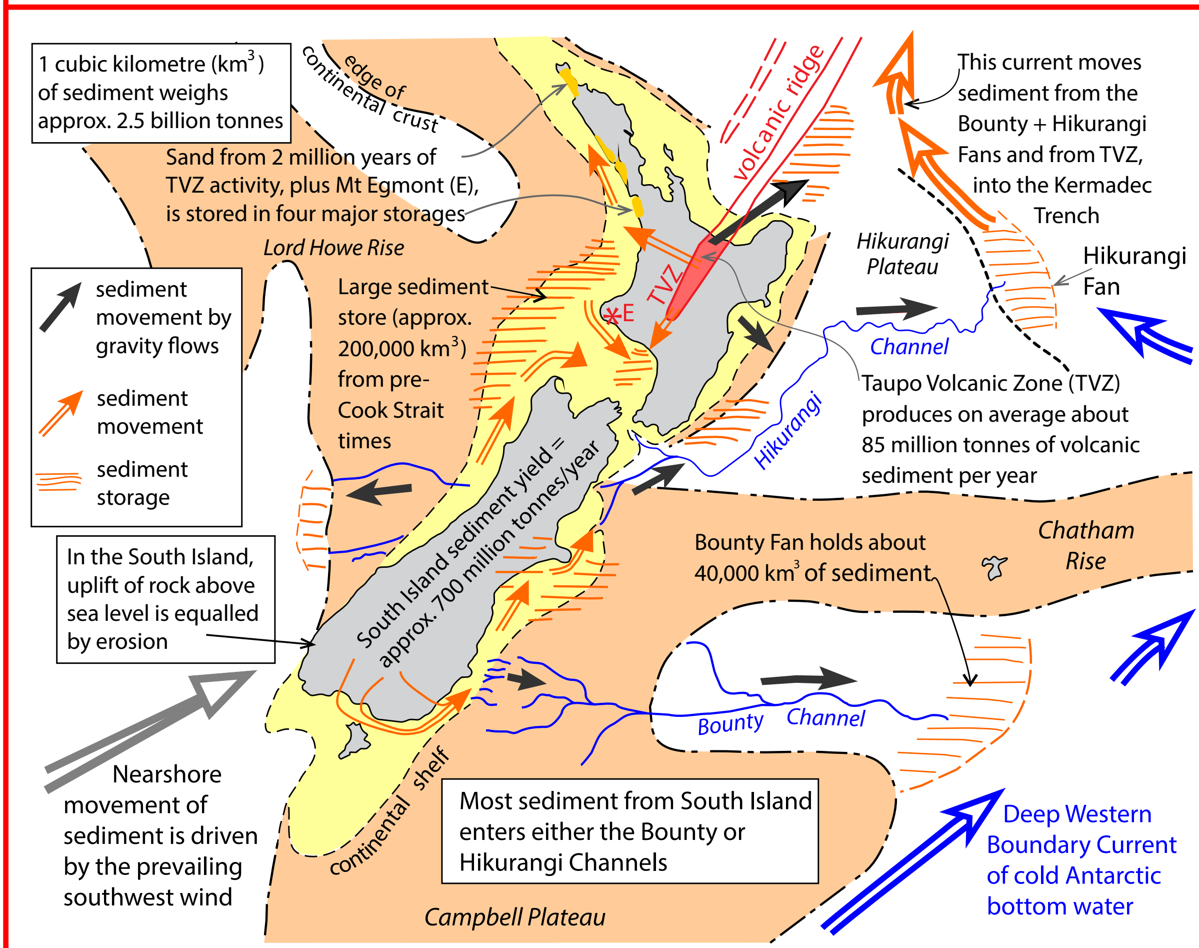
At the coast, the dominant influence in both islands and on both coasts is wave action from the south or southwest. Only the eastern coast of Northland and the Bay of Plenty are isolated from this influence – Auckland and Northland have small rivers and generate little sediment anyway, while the dominant storm direction is from the northeast.

South Island sediment

Waves move sand along the coast in zigzag fashion by longshore drift, as shown in Box 6.10. In most of New Zealand, sand thus moves predominantly to the north and east. As Box 14.2 shows, some of the sediment rounding Otago Peninsula is caught by the Otago submarine canyons and funnelled off by turbidity currents into the Bounty Channel and thence to the deep sea. The remainder travels up the coast to Banks Peninsula.

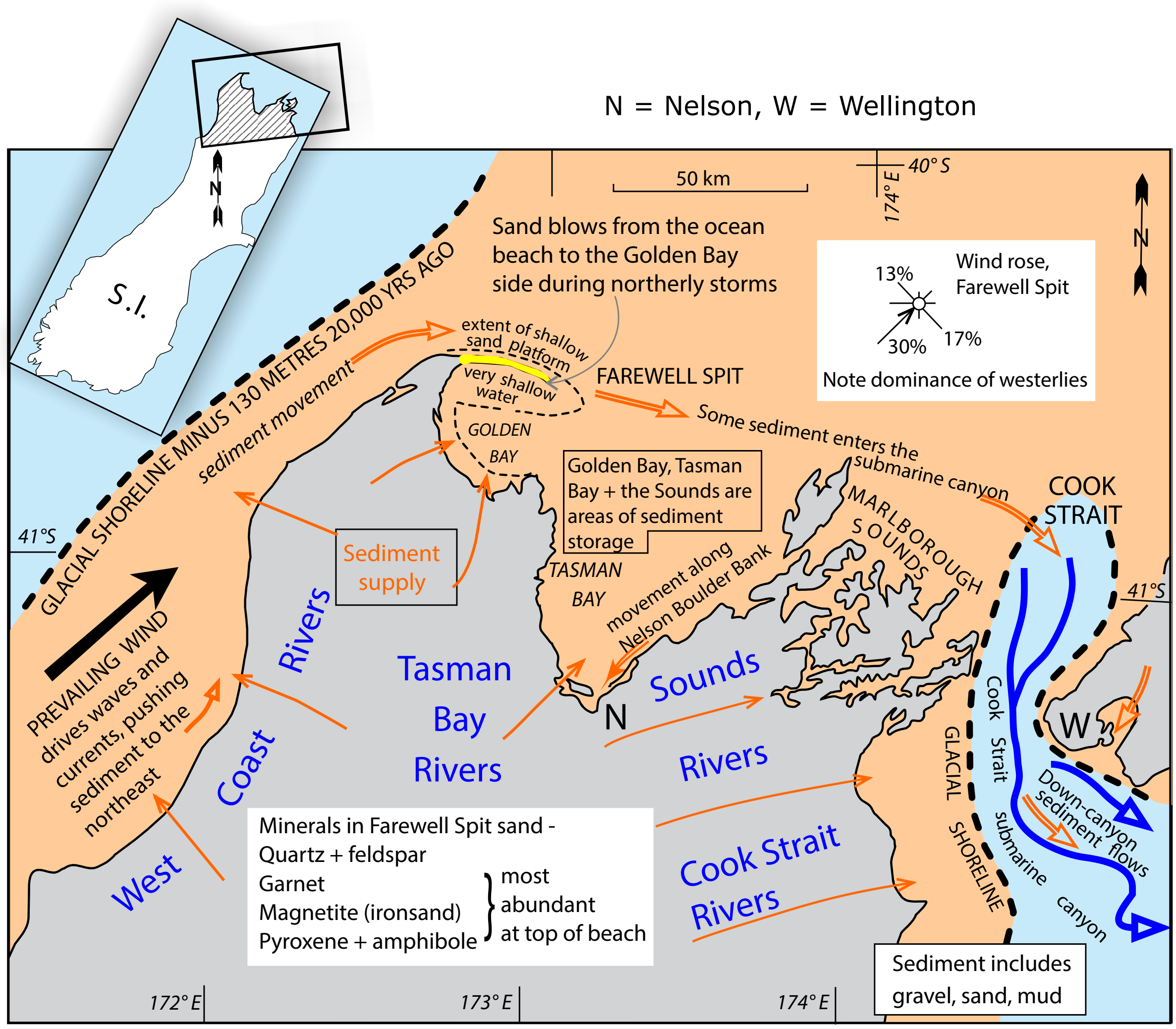
The longshore drift process is enormously abrasive, and all weaker rock and mineral grains being carried along are worn down and destroyed by the time the sand gets to Banks Peninsula. Only resistant minerals and rocks – chiefly quartz and chert – survive. That sand then passes around the peninsula and enters the Banks Peninsula submarine canyon. Along with sediment entering the Kaikoura Canyon, it travels in periodic turbidity currents into the Hikurangi Trough alongside the North Island.

Box 14.2. New Zealand sediment budget.



This situation is mirrored on the west coast of the South Island. At the present time, some sediment enters the small west coast submarine canyons and is taken westwards down into the Tasman Sea. Most, however, travels to Farewell Spit, there building the South Island's longest sand spit (Boxes 14.3 A, B and Fig. 14.2). It is notoriously difficult to measure how much sand is moved by such a system, although it has been estimated that 5–6 million cu m/year moves up the west coast, and the figure could be even higher. Only 10% or less of that amount has been stored in Farewell Spit and its sand flats over the 8000 years the spit has been forming at the post-glacial high sea-level. The remainder has gone somewhere else – some into storage on the continental shelf, and some around the corner into Cook Strait.

Box 14.3 A. Farewell Spit. *(Preamble to accompany this map is on following page).*



Box 14.3 B. Farewell Spit – Preamble *(to accompany map on previous page).*

Farewell Spit is New Zealand's largest sand spit, by far, and it exists because of a large supply of sand from the southwest, in conjunction with the opening into Cook Strait. Most sediment reaching the west coast from the Southern Alps is moved northeastwards by waves driven by the prevailing southwesterly wind (an example of longshore drift). At Cape Farewell, the northern tip of the South Island, waves and associated storm-driven currents sweep the sand round the corner into the mouth of Cook Strait, and into the lee of Cape Farewell. Here, protected from the westerlies, it is building an extensive shallow platform of sand in Golden Bay. The Spit is merely the line of dunes built by onshore winds, a small part of the total volume of sand involved, but important because it protects Golden Bay from direct wave action.

Farewell Spit in its present form has been build during the last 8000 years, since world sea level returned to its present height, following the low sea level of minus 130 metres during the last glacial maximum, 20,000 years ago. During glacial low sea levels Cook Strait does not exist, and West Coast sand moves on northwards to Taranaki and the North Island west coast, building the large sediment store (200,000 cubic kilometres) shown on the Sediment Budget map (Box 14.2) west of Cook Strait. At such times, Farewell Spit and similar spits are dry land, a long way from the sea and from supplies of sand.

The spit and its associated sand platform have presumably been build during several glacial-interglacial cycles, but would have been modified by weathering and erosion during each glacial low sea level.

Much sediment is stored in Tasman and Golden Bays during high sea levels like the present, but some makes its way through Cook Strait and into the Cook Strait submarine canyons. During low sea levels, rivers deliver sediment more-or-less directly into the canyons. Once in them, sediment moves down-canyon as periodic sediment gravity flows (Boxes 6.5 A, B).

Farewell Spit is a minor component in the bigger picture, the New Zealand-wide sediment budget. The large amount of sediment produced each year reflects New Zealand's high rate of tectonic activity (mountain building, chiefly South Island) and volcanic activity (North Island). The map summarizes the major sediment sources (South Island mountains, North Island Taupo Volcanic Zone plus Mount Egmont/Taranaki); the three major sediment transport routes – around the continental shelves (wave- and storm-driven), off the continental shelves along submarine canyons (by sediment gravity flows), and around the eastern margin of the New Zealand continent (by the Deep Western Boundary Current of cold Antarctic bottom water); and finally the major areas of sediment storage.

As noted, glacial-to-interglacial sea level movements of more than 100 metres, on a time scale of 100,000 years, play a big role in the workings of the system. The sediment in some of the storages has been accumulating for some millions of years.

Fig. 14.2. The 30-km-long Farewell Spit has built out from the northwest corner of the South Island in just the last 8000 yrs since sea-level returned to its present level after the end of the Last Glaciation. Photograph courtesy of Google Earth.

Waves and tidal flows drive sediment into western Cook Strait from both northern and southern sides. Quite a bit is stored there, but some enters the submarine valleys and travels in periodic turbidity currents through the strait into the Hikurangi Trough via the Hikurangi Channel. Once in the Hikurangi Trough, sediment from both sides of the South Island, plus sediment from eastern North Island, is driven by subduction of the Pacific Plate under the North Island, to form an accretionary wedge (Chapter 12). Growth of the wedge is helping to uplift eastern North Island – thus the North Island is growing at the expense of the South Island.

Effects of the glacial cycle

The regular 100,000-year glacial–interglacial climatic cycle adds a layer of complexity to the sediment story. The glacial-maximum low sea-level (about –130 m) moves shorelines out to the edge of the continental shelf, closing Cook Strait. New Zealand's land area is much bigger and weather systems are more severe, so the sediment budget is probably bigger and faster moving. During these periods sediment can enter submarine canyons more or less directly from rivers, so transfer to the deep sea is greater, building the Bounty, Hikurangi and west coast submarine fans. Sediment is also built up on the upper continental slope, as shown west of Cook Strait.

North Island sand barriers and tombolo

As described in Chapter 9, the rhyolitic volcanoes of the TVZ have been enormously productive over the past couple of million years. The eruptive products have been spread far and wide, but a large volume of sand – hundreds of cubic kilometres – is retained in the four major sand barriers and tombolo on the North Island's west coast north of the mouth of the Waikato River.

Chapter 15

Eastern South Island

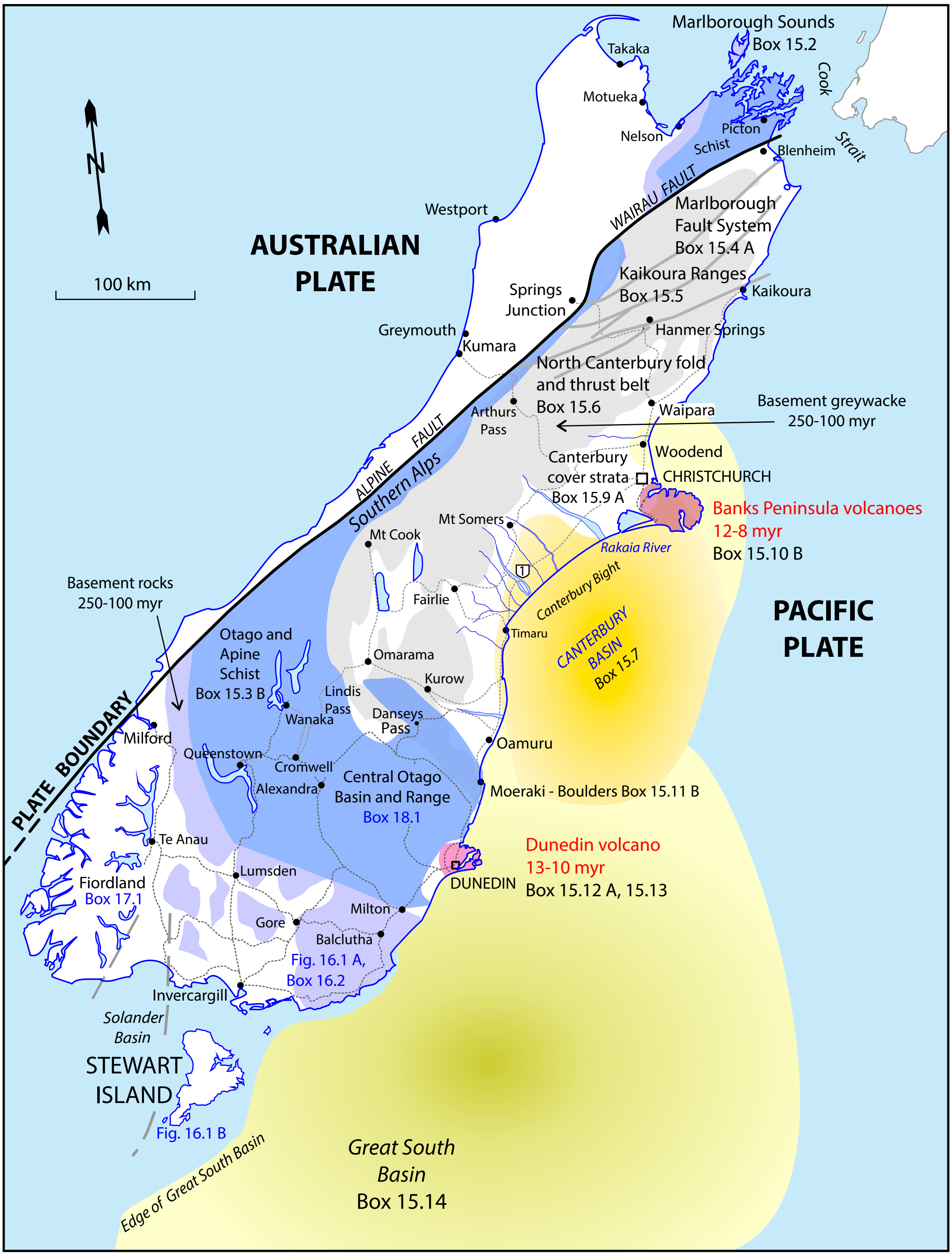


Fig. 15.1. Regional map of eastern South Island.

This chapter begins with some general points about the geology of the South Island, which in many respects is quite different from that of the North Island. Dominating everything is the Alpine Fault, one of the world's great tectonic faults. It forms part of the boundary between the Pacific and Australian plates (Box 2.4) and bisects the island, running in an almost straight line from Blenheim in the northeast to Milford Sound in the southwest. It is important to recognise that the Alpine Fault separates two chunks of continental crust, because this has a big influence on how the fault is expressed and what it does (see also chapters 3 and 19).

Everything to the north of the Alpine Fault (including the whole of North Island) is part of the Australian Plate and is moving northwards at a few centimetres a year, while everything south of it is part of the Pacific Plate and is moving southwestwards (i.e. nearly parallel to the fault) at around 4 cm/year. As a consequence of this, there has been nearly 500 km of right-handed sideways displacement on the fault (the opposite side of the fault is moving to the right). The fact that the two plates do not pull away from each other and create a large hole in the ground, but instead create mountains where they join, is because the direction of movement of the Pacific Plate is not exactly parallel to the fault, but towards it, at an acute angle of around 10° on the map (Boxes 19.1 A and 19.3). Thus, Pacific Plate rocks press against Australian Plate rocks and are pushed up to create mountains. The actual trajectory followed by the rocks is oblique, not straight up, but the uplift is always measured and expressed as vertical uplift, as this is much easier to comprehend.

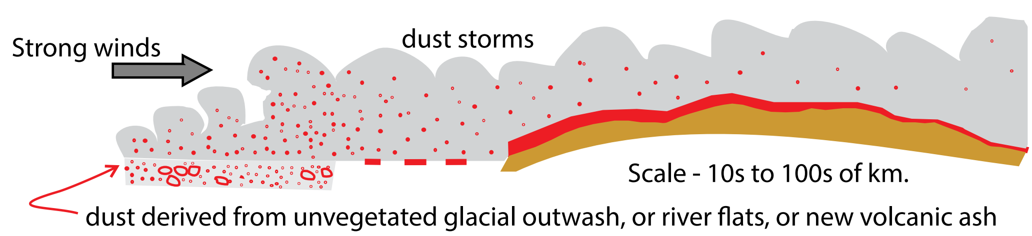
As a result of the activity at the plate boundary, the dominant geological feature of most of the South Island is rapid tectonic uplift. For rapid, read rates averaging between 1 mm/year and 11 mm/year in different places – on a global scale, that is fast. Uplift creates relief, and relief generates erosion. By chance, South Island mountains lie across the path of wet westerly winds of the Roaring Forties, causing the winds to rise steeply. This generates a high orographic rainfall, and in turn many streams and rivers. Throw into the equation the fact that the mountains were extensively glaciated during the most recent ice age, 20,000 years ago (and still are in places), and the result is gravel and dust, both in enormous quantities.

Everywhere you go in the South Island, you see gravel in cuttings – this is the material that has built the Canterbury Plains and many other river floodplains. And not all of it dates from the last glacial period. Gravel has been shed from South Island's rising mountains for at least 5 myr, and in a few places like Marlborough for even longer. When those sorts of ages are involved, old gravels themselves have been subjected to the tectonic forces that are raising mountains, and therefore you will see gravel beds that dip at various angles from the horizontal.

Dust is produced largely by the grinding action of moving glaciers. Owing to the prevailing westerly winds, most of it is found along the eastern side of the island. During and immediately following glaciations in particular, it is transported in large quantities. It is less easy to identify than gravel, but you can learn to recognise it. It falls to the ground and is trapped by vegetation, where it gradually builds up a layer of brownish silt beneath the topsoil up to 10 m thick. It is called loess (pronounced 'lerss'), and can be recognised by its uniform brownish colour, fine grain size (the grains are too small to see with the naked eye or with a 10× hand lens), smooth feel, lack of internal layering and tendency to break into vertical columns. If the cutting is deep enough, the local rocks will appear beneath the loess.

Fig. 15.2. The Canterbury Plains as well as many parts of North Canterbury and Marlborough have been built by river gravels eroded off the South Island's rising mountains. The impressive cliffs on either side of the Rakaia Gorge, Canterbury, expose vast thicknesses of gravel that built up the inner parts of the Canterbury Plains, especially during glacial periods of increased erosion. Photographer Bruce Hayward.

Box 15.1. Loess deposits.



Wind-blown dust layer:

- technical name loess (pronounced lerss)
- deposited in 100s to 1000s of years
- distribution governed by prevailing winds
- deposit thins away from source, but grain size (silt) does not vary much
- thickness ranges from less than 1mm (very hard to identify) to more than 10 m.

Can be dated:

- to a limited degree by physical methods (e.g. radiocarbon, thermoluminescence), by placement in a succession of dust and ash layers, by containing distinctive minerals from known ash layers.

Identification:

- cannot be identified as readily as volcanic ash, because the minerals come from many sources.

After deposition:

- as for volcanic ash.

Wind-blown dust layers in New Zealand form at all times, but are best developed during glacial periods when vegetation is reduced and there is a strong wind regime. They are most common in the southern half of the North Island and eastern South Island.

Fig. 15.3. At Dashing Rocks, Timaru, basalt lava flows are mantled by 4–5 m thickness of orange-brown, fine-grained loess. This is glacial flour that was blown here from the outwash fans of Southern Alps glaciers during cold glacial periods. Photographer Bruce Hayward.

A third outstanding feature of the South Island, after gravel and loess, that arises directly from the copious sediment supply and the pervasive uplift is the universal presence of magnificent river terrace landforms (Box 10.3). These are portions of former river floodplains, uplifted and abandoned when the river cut down its bed in response to the uplift. Whole staircases of terraces can be seen in many places, and the river alluvium that builds them is generally gravel, as noted above.

The Marlborough Sounds

The rocks around the Marlborough Sounds provide the most direct link between the geology of North Island and South Island, because they lie north of the Alpine Fault and are therefore on the same plate as the North Island – the Australian Plate. As noted in Chapter 14 and delineated in Box 15.2, the sounds are river valley systems that are being drowned by tectonic subsidence that is preceding the relentless southwards march of the North Island's Whanganui Basin (Chapter 10). This is a much more profound valley drowning than that caused by post-glacial rise of sea-level. Ultimately, the Marlborough Sounds area will disappear beneath the waves, although landowners can take some comfort from the fact that the subsidence is slow – around 1 mm/year, which is much slower than the sea-level rise expected from man-made global warming over the coming century.

One consequence of the tectonic subsidence is that the land drops about 100 m during every 100,000-year glacial–interglacial cycle of sea-level change. Thus, any river floodplain or coastal terrace constructed during an interglacial high sea-level is well below sea-level by the time of the next high sea-level, and as a consequence the Marlborough Sounds are the only part of the South Island that lacks river terraces.

Box 15.2. Marlborough Sounds.

The Whanganui Basin is described in [Chapter 10](#)

The rocks in the western Marlborough Sounds are a continuation of those found in east Nelson, and as such are described in Chapter 22. The central and eastern sounds, meanwhile, are made of the same rocks as Central Otago and the southern part of the Southern Alps, and have been separated from them by 480 km as a consequence of sideways displacement on the Alpine Fault. Matching rock sets on opposite sides of the fault are encountered a number of times around the South Island. The displacement is always right-lateral or dextral. This is a direct consequence of the anticlockwise rotation of the Pacific Plate.

Most of the rock seen in the sounds is broadly called Marlborough Schist. Boxes 15.3 A,B describes how schist is formed by the application of heat and pressure (metamorphism) to greywackes. There is a much larger area of schist in Central Otago (Box 15.3 B), and as happens there, the change from greywacke rocks to schist in Marlborough is gradual and progressive. There is a transition from northwest to southeast across the sounds, between 'regular' greywacke (itself a slightly metamorphosed rock) and high-grade mica schist, through a wide zone of what can be called semi-schist. Most of what is seen around the sounds falls into the categories of severely deformed greywacke (as seen at Pelorus Bridge on SH6) and semi-schist (as seen around Picton and the eastern sounds).

Fig. 15.4. View northeast over Kenepuru Sound with Pelorus Sound to the northwest (top left). The Marlborough Sounds are river valleys that are being drowned by the sea, partly in response to their tectonic subsidence. Photographer Lloyd Homer, GNS Science.

Box 15.3 A. Metamorphism – flysch to greywacke to schist, also gneiss, mélange and fault gouge.

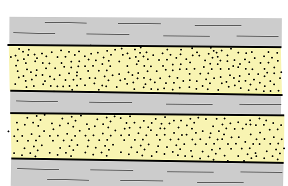
Metamorphism is the process by which rocks are changed by heat, hot water and/or pressure, to the point that existing minerals begin to change into new minerals. It is a matter of chemical equilibrium. Most minerals are at home, comfortable in, in equilibrium with, a particular set of chemical circumstances. If those circumstances change, then the minerals begin to change into a different form to suit the new conditions. Chemical weathering of rocks at the earth's surface is an example (Box 6.1 B) though by convention weathering is not called metamorphism. That term is reserved for changes going in the opposite direction, i.e. downwards into a hotter nether world. Either way, the changes of mineral type are glacially slow, but of course there is plenty of geological time available.

There are many variations in the metamorphic process. In New Zealand the commonest process is the one that takes flysch deposited in a subduction trench, shoves it into an accretionary wedge and changes it into greywacke (Boxes 6.2 B,C). That process can be taken one or two steps further, to change greywacke into schist. It requires some event to raise the temperature and pressure by another couple of notches. This can be extra-deep burial in the accretionary wedge, or it can be something unrelated, like a tectonic collision between two terranes (Boxes 5.1, 5.2, 5.3).

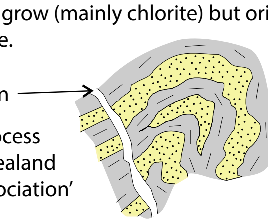
It is thought that most New Zealand schists are the result of collision between the Torlesse and Caples/ Waipapa greywacke terranes. There is very little schist to be seen in the North Island, just small areas within Torlesse greywackes in the Kaimanawa Ranges, and occasional blocks brought to the surface by volcanoes. There is far more schist in the South Island. You may see this rock in the North Island as decorative facings and paving stones. Look for a grey, strongly layered rock, with lots of sparkling grains of white mica, and a strong alignment of the grains. There may also be veins (crack-fillings) of white quartz, and scattered grains of red garnet.

SCHIST is a metamorphic rock. It began as some other rock, and has been changed (metamorphosed) by heat and pressure to the extent that the original minerals and structures have been totally reconstituted. Most New Zealand schists began as greywacke. A typical progression is as follows:

(1) Alternate layers of sand and mud (flysch) deposited in deep-sea trench

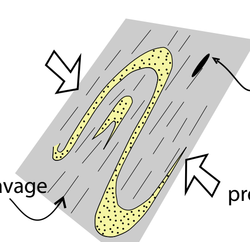


(2) Metagreywacke stage of metamorphism during deep burial (about 10km), folding, cooking (to 200° C). Some new minerals grow (mainly chlorite) but original layering is still visible.

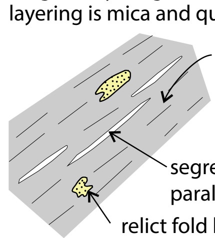


This stage of the process is the typical New Zealand 'greywacke rock association'

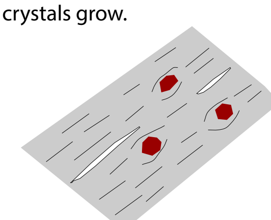
(3) Greater pressure causes flattening of folds, thinning of fold limbs, thickening of fold hinges, and re-alignment of flakey minerals (clays and mica) perpendicular to the pressure to form a new layering-cleavage (full name = axial plane cleavage, because it is parallel to the axial plane of the folds). Mica grows as a new mineral. This stage is not commonly preserved.



(4) Original layering has disappeared. The new layering is mica and quartz segregations.



(5) If taken to higher temperatures and pressures, red garnet crystals grow.



(6) The whole cycle can be repeated. An old schist can be re-constituted into a new schist.

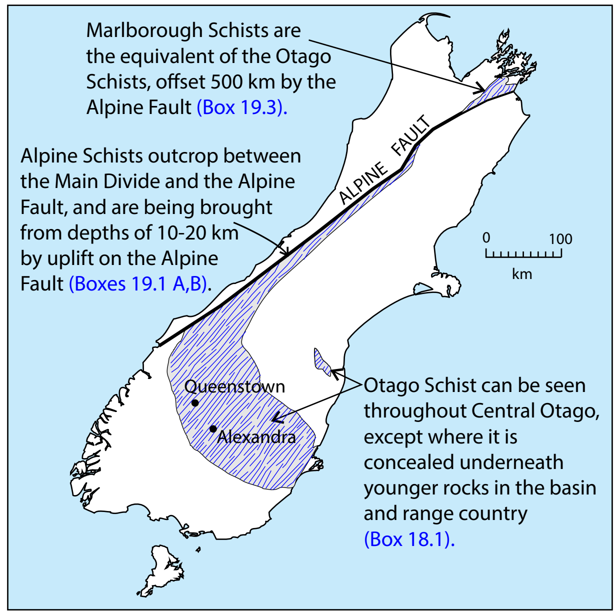
(continued in next box)

Box 15.3 B. Metamorphism – flysch to greywacke to schist, also gneiss, mélange and fault gouge *(continued from previous box).*

(7) <u>What you see in schist country (map)</u>
The rock breaks along the schistosity and along joint surfaces, so you see box-shaped landforms tilted at various angles. Schist splits easily along the schistosity, displaying white mica grains lying flat on the schistosity, and reflecting light. It is a popular ornamental stone.

(8) The main body of New Zealand schist, the Otago Schist, formed during the collision of two large (100's of km) bodies of metagreywacke (the Torlesse and Caples Terranes) as a result of continental drift about 150 million years ago (Box 5.3).

(9) Quartz veins and segregations in schist contain gold (Box 18.1).



GNEISS (pronounced nice) is the second major category of metamorphosed rock. It differs from schist in being coarser grained, and while it is distinctly layered (foliated is the technical term) the layering is coarser, and the rock does not split along the layering like schist does. It is coarser because it has been metamorphosed deep in continental crust, under high temperature and pressure which were close to melting conditions. Thus the minerals most typical of gneiss are quartz, feldspar and mica, which are the typical minerals of granite, and in nature there is a continuum between gneiss and granite. Local lowering of pressure may allow parts of the gneiss to melt, forming veins of granite in the gneiss (an association called migmatite). If the conditions exceed melting on a large scale, molten granitic magma results, which then moves to higher levels, to form granite batholiths, or perhaps reach the surface as volcanoes of rhyolitic (high silica) composition (as in the Taupo Volcanic Zone of the North Island).

MELANGE

Mélange is a French cooking term meaning mixture. It has been taken into geology to mean rocks that have have been mixed and stirred together by tectonic processes. Typically, mélanges form where two large bodies of rock grind together during a tectonic collision event, as when major rock terranes are brought together at a subduction zone. There is usually a ductile (i.e. easily deformed) matrix of clay-rich rock, or perhaps serpentinite, which is smeared around pieces of rock (from millimetres to kilometres in scale) which may be shattered but which have hung together. The pieces include rocks from a variety of sources, some of which may be metamorphic rocks from deep in the crust, or pieces from the oceanic crust which was carrying the terrane into the subduction zone. Mélange is commonly developed where 'obduction' occurs, the counterpart of subduction where slabs of oceanic crust are pushed up and onto continental crust. The grinding process tends to generate high pressures, but not necessarily high temperatures, and there is a characteristic set of new (i.e. metamorphic) minerals which form under those conditions. One of them, glaucophane, is blue, and the name "blueschist" is used for that general category of metamorphic rock. A geology graduate student who was both a good cook and a good cartoonist had a cartoon on his office door captioned "Spécialiste de Mélange".

FAULT GOUGE — CATACLASITE AND MYLONITE

On a more localised scale than mélange, rock gets ground up when it is trapped in a fault plane between opposing blocks of crust. It is commonly called fault gouge; the technical term is cataclasite. Fault gouge can be hundreds of metres wide, for example on the Alpine Fault in places. In extreme conditions, frictional heat actually melts the ground-up rock, producing a glassy rock called mylonite.

Fig. 15.5. The rocks that form the sides of the small gorges at Pelorus Bridge, on the road between Nelson and Havelock, is severely deformed greywacke that has not been as deeply buried and metamorphosed as the schist of eastern Marlborough around Queen Charlotte Sound. Photographer Bruce Hayward.

First encounter with the Alpine Fault

Our clockwise traverse of the South Island now crosses the Alpine Fault (here also called the Wairau Fault, after the Wairau River that follows it) at Blenheim, onto the Pacific Plate. The traverse stays on the Pacific Plate until we cross the Alpine Fault again, in Chapter 20, onto the West Coast, Buller and Nelson.

This being our first encounter with the Alpine Fault, we can begin to get acquainted. As it happens, the Wairau Fault sector of the Alpine Fault, between Nelson Lakes National Park and Cook Strait, is different from the rest. This is because it is a 'pure' sideways-moving fault – there are no components of convergence across it. Such faults are rare. The critical difference is the orientation of the fault – the Wairau sector is aligned 65°E of north, compared with 55° for most of the fault. That 10° difference puts the Wairau sector exactly parallel to the local movement direction of the Pacific Plate, which is rotating anticlockwise around its pole of rotation at latitude 60° S, longitude 180° (Box 2.1). The consequence of this is that there are no mountains adjacent to the fault on its southeastern side, as there are further south. The wide zone of ground-up rock that is characteristic of the actual fault zone has been excavated by the Wairau River.

The mountains to the northwest of the Wairau Fault, the Richmond Ranges, have a different origin that is related to the Marlborough Fault System (see below).

Boxes 19.1 A, B and 19.3 describe the Alpine Fault and how it causes the rise of the Southern Alps. Like most big faults, the Alpine Fault is seldom exposed to view. Having a 25 myr history, 480 km of sideways movement and perhaps 50 km of vertical displacement, rock along the fault is well and truly ground up. The fault-scarp traces of the most recent activity can be seen (with difficulty) at four places by a main road, as indicated in Box 19.3. One of these is in the Wairau Valley at Branch River, where the 2 m step up on the side road to the Branch River power plant is the fault trace. The Branch River is one of the rare localities where a series of river terraces has been displaced by the fault, allowing a reconstruction of the history of recent movements on the fault. This concept was explained in Box 13.3 A for similar features in the Wellington area.

Fig. 15.6. View southwest along the Wairau (Alpine) Fault trace (uplifted on left) at Branch River, Wairau Valley. Notice the sag pond on the trace and the sideways (dextral) displacement of terraces and stream channels in the fields. Photographer Lloyd Homer, GNS Science.

The big bend in the Alpine Fault

After it crosses the eastern car park at Lake Rotoiti in Nelson Lakes National Park (the recent trace forms the step up to the toilet block!), the Wairau/Alpine Fault enters its big bend – the only significant bend along its length. This is a lazy-S bend, giving rise to a left step in the line of the fault (left step because it steps to the left when viewed from either direction along the fault). As explained in Box 19.2 A, a left step on a right-lateral fault causes one side of the fault to press against the other side at the step. Compression equals uplift, because up is the only way the rocks can go to get out of the way. And that is where the Southern Alps begin – in the St Arnaud Range and Spenser Mountains. It is no coincidence that the strip of deep-seated schist that is dragged up along the fault for most of its length (Box 19.1 A) widens greatly in the bend, and then disappears abruptly as the fault bends into the Wairau trend and uplift ceases.

There is a reason for the big bend, and it lies in the way the Marlborough faults are behaving.

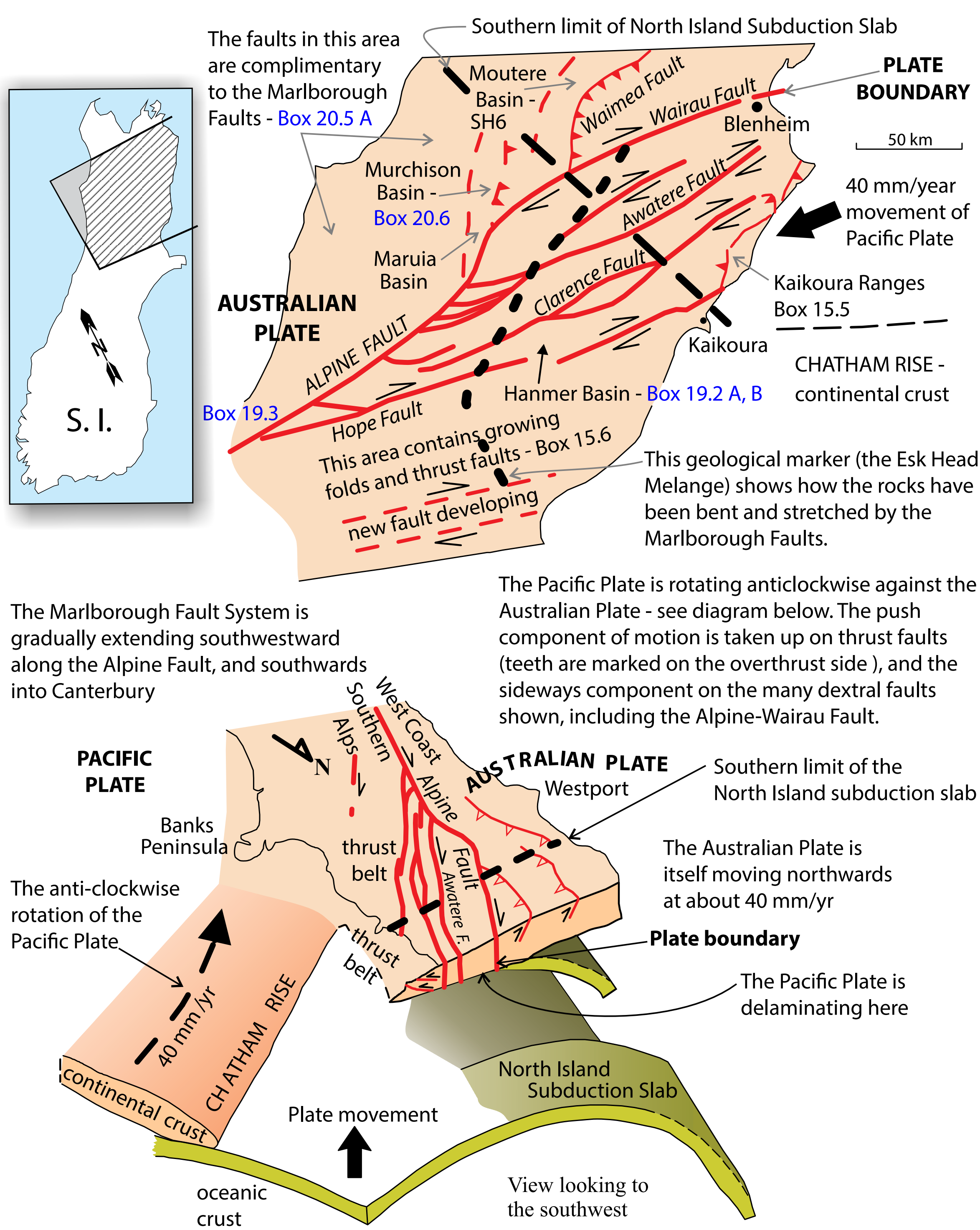
The Marlborough Fault System

This is a group of three main faults – the Awatere, Clarence and Hope – and some lesser faults. They each branch off the Alpine Fault, as shown in Box 15.4 A, and like the Alpine Fault they are right-lateral faults, with a component of compression – enough to raise the Inland and Seaward Kaikoura ranges, the highest mountains outside the Southern Alps (Box 15.5). As noted above, inland Marlborough is sliding along the Wairau/Alpine Fault with no cross-fault compression. However, as you go southwestward, Marlborough continental crust starts to push against West Coast/Australian Plate crust, first in the big bend of the Alpine Fault, as discussed, and then at 10° with the realigned Alpine Fault.

Marlborough is the meat in the sandwich. Not only is it pushing against a backstop of continental crust, it is also being pushed itself by a large salient of continental crust, the 1000 km-long Chatham Rise. The result is an unusual tectonic response. Continental crust is underlain by, and floats in, oceanic lithosphere/mantle, and the two normally move together, but in Marlborough's case they are being separated, or delaminated. While the heavy mantle rocks continue to dive down into the southernmost bit of the New Zealand subduction zone, the overlying continental crust is being peeled off and pushed backwards by the West Coast backstop. Box 15.4 A explains the situation.

The Marlborough crust that is being peeled off is piling up in a tectonic traffic jam, and is being pushed northeastward, in slices between the Marlborough faults. As well as forming the Inland and Seaward Kaikoura mountain ranges, the effect is the widening of Marlborough. In turn, the widening is elbowing everything to the northwest of it (i.e. the Nelson region, the northernmost sector of the South Island) further towards the northwest. The consequences of this push can be seen all across Nelson, as it determines the tectonics, earthquake hazard and topography of the region. We will enlarge on this subject in Chapter 22, but the two points to mention here are that it is the widening of Marlborough that is pushing the Alpine/Wairau Fault out of line with its southern continuation, thereby creating the 'big bend' in the fault; and that the resulting northwestward-directed compression is raising the Richmond Ranges immediately northwest of the fault.

Box 15.4 A. Marlborough Fault System. *(discussion in following box)*



The Pacific Plate is rotating anticlockwise against the Australian Plate - see diagram below. The push component of motion is taken up on thrust faults (teeth are marked on the overthrust side), and the sideways component on the many dextral faults shown, including the Alpine-Wairau Fault.

The Marlborough Fault System is gradually extending southwestward along the Alpine Fault, and southwards into Canterbury

Box 15.4 B. Marlborough Fault System – discussion.

The Marlborough Faults lie between the Alpine Fault (the boundary between the Pacific and Australian Plates in the South Island) and the coast between Blenheim and Kaikoura. They branch off from the Alpine Fault, and slice continental crust of the Pacific Plate into northeast-trending slivers which are all sliding to the right as you look at the map. They do not slice up the oceanic lithosphere of the Pacific Plate which lies underneath the continental crust — this point is important in understanding how they fit in to the dynamic picture of the plate boundary (below).

The effect of the fault movements is to "smear" rock units to the right (Esk Head Melange marker on the map); to widen this part of the South Island, thereby displacing the Alpine Fault to the north (causing the bend in it) and pushing the rocks of east Nelson towards the northwest over the Waimea thrust fault system (Fig. 22.1); and to raise the Kaikoura mountain chains by the same squeezing effect (Box 15.5).

What is driving all this? The same force that is pushing Canterbury to the southwest along the Alpine Fault, and raising the Southern Alps: movement of the Pacific Plate towards the southwest at around 40 mm per year. This movement is put in context in Boxes 2.1 and 2.4, and is further explored in Boxes 19.1 A and 19.3.

The plate boundary is obliquely convergent along its whole length through New Zealand, i.e. the convergence has two components, shortening which causes subduction, and sideways motion which causes faults with combined up-down and sideways displacements (Boxes 6.6 C, D, G). Alongside the North Island shortening is dominant, and oceanic crust is subducted underneath continental crust. In the South Island sideways motion dominates, and subduction fails underneath Marlborough. The other big difference from the North Island situation is that the plate boundary leaves the continent-ocean boundary, and enters continental crust, at Blenheim. There is still a small component of shortening, but now it is pushing continental crust against continental crust, and that is why the South Island has high mountains. Continental crust is too bouyant to be subducted. It can, however, be pushed up into the air and destroyed by erosion, and that is what is happening along the Southern Alps.

<u>How to read the block diagram</u>

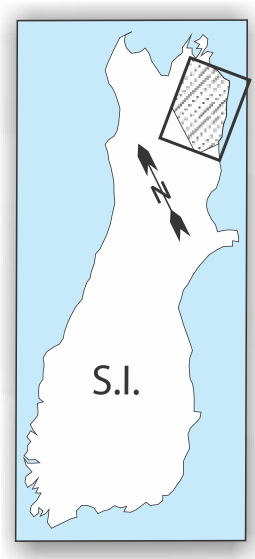
The block diagram (Box 15.4 A) is an oblique aerial view of the northern half of the South Island, looking southwest. The northernmost part of the island is cut away to show the subduction slab extending down from the North Island to its southern limit at Kaikoura. You are looking along the line taken by the Pacific Plate as it moves obliquely towards the Australian Plate.

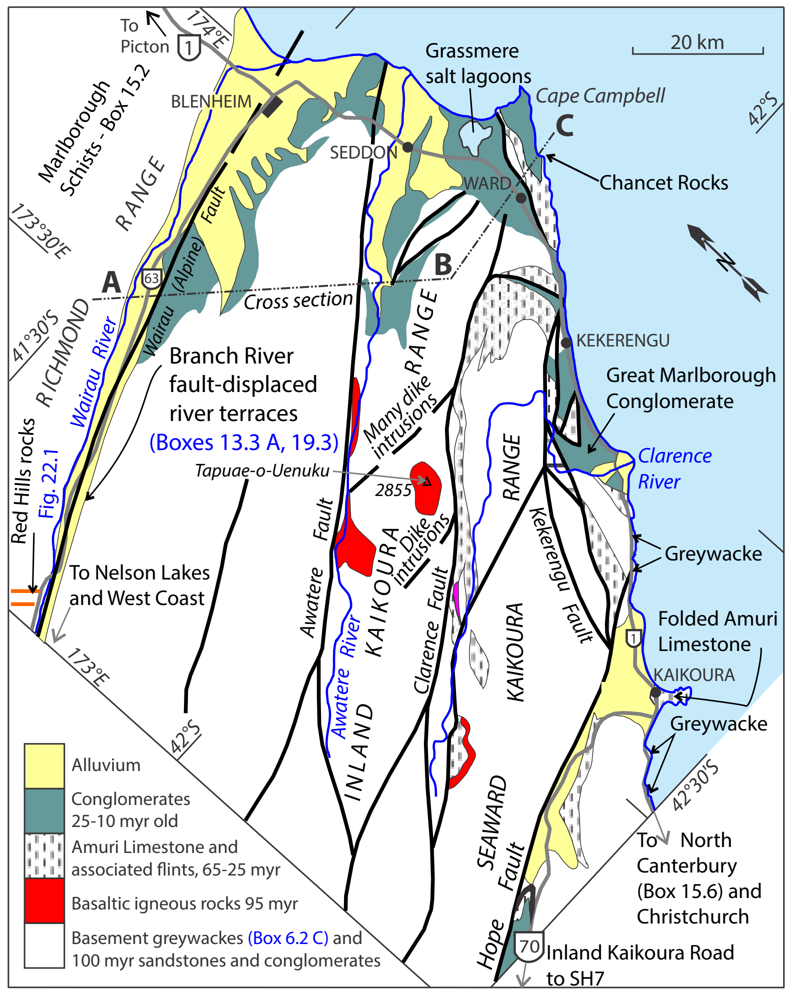
<u>Making a plate subduct under itself - a rare tectonic sublety</u>

Oceanic crust is available for subduction as far south as Kaikoura, where it is replaced by continental crust forming the Chatham Rise. The Pacific Plate is able to carry its oceanic crust downwards towards the southwest (i.e. to subduct it), but its efforts to push its continental crust in the same direction are resisted by a 'backstop'- the continental crust of the Australian Plate west of the Alpine Fault. The effect is that the Pacific Plate is delaminating, so that its continental crust is being pushed northward over its own oceanic crust. This is happening in the triangle formed by the Alpine (Wairau) Fault, the southern limit of the subduction slab, and the edge of continental crust (to a close approximation the coastline), i.e. the area of the northern Marlborough Faults. On the block diagram (Box 15.4 A) the black arrow at the lowest point shows how oceanic crust of the Pacific Plate is being taken underneath continental crust of the same plate. Globally, this is a rare phenomenon.

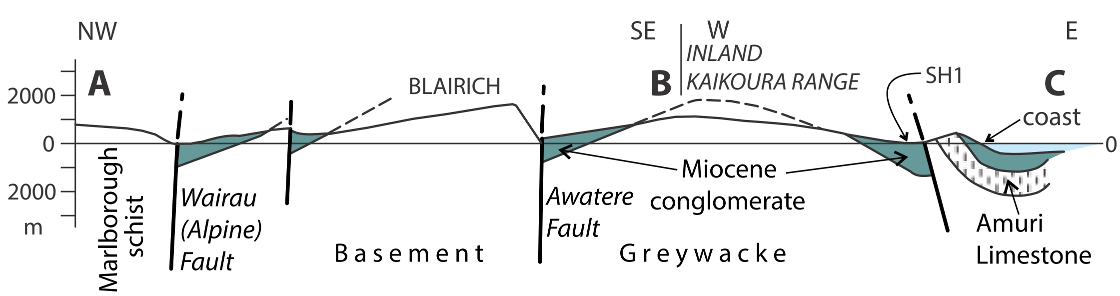
Editors' note: Since Peter Ballance wrote this section, it has been shown that the Hope Fault is more active now than the Wairau Fault. There is a shift in terminology to place the northeastern extension of the Alpine Fault along the Hope Fault rather than the Wairau Fault.

Box 15.5. Kaikoura Ranges.





This map and cross-section have been greatly simplified. The geology of this region is complex.



NOTE- 1) There is a 2x vertical exaggeration on this cross-section

2) The faults all have a dextral sideways displacement as well as a vertical. As you look at the section, the rocks to the right of each fault have moved towards you.

3) Amuri Limestone and associated rocks are not preserved inland.

Fig. 15.7. Oblique aerial view over the Kaikoura Ranges of Marlborough showing the three main faults of the Marlborough Fault System (Hope Fault – far left, Clarence Fault, and Awatere Fault – centre) and the Wairau/Alpine Fault diagonally across the top right. Photograph courtesy of Google Earth.

A Brief Geology of Marlborough

Amuri Limestone

Like the rest of the country, Marlborough can be described in terms of old, thick basement rocks (the terranes that accumulated against Gondwana; Chapter 5) overlain by younger cover strata. Because the region is being strongly deformed tectonically, cover strata have been removed by erosion in some places and preserved in others. The map and simplified cross section in Box 15.5 show the effect.

Cover strata in Marlborough include a prominent fine-grained white limestone (really a hard chalk) called the Amuri Limestone. It originated as a deep-sea calcareous ooze on the continental slope of the time, 65–25 myr ago, during phase three of New Zealand's formation. Like its English equivalent in the White Cliffs of Dover, it contains nodules of dark grey chert or flint, made of fine-grained silica. On the coast at Ward Beach, the limestone forms classic polished pebbles, coin-shaped from sliding up and down the beach. They display a wide range of trace fossils – structures left behind by burrowing organisms like worms (Box 6.5 B). The Amuri Limestone also forms Kaikoura Peninsula, where medium-scale folding is well displayed.

The Great Marlborough Conglomerate

There is one aspect of the geology of Marlborough that differs from the rest of the South Island. Mountains high enough to shed gravel were formed here shortly after the plate boundary became energised 25 myr ago, and a steep transition between these mountains and deep ocean was also in place. The thick conglomerate deposits (hardened gravels), aged 25–10 myr, can now be seen inland, while along the coastal strip is a remarkable formation called the Great Marlborough Conglomerate. This is the only rock formation in New Zealand to have the word 'Great' in its name, and is a series of submarine debris-flow deposits (Box 19.5) that were carried down a steep continental slope here 18–12 myr ago. The debris in the deposits comprises pieces of the local rocks – both basement greywackes and cover strata – in chunks up to 1 km or more long. Now uplifted above sea-level, the formation occurs chiefly between Kekerengu and the mouth of the Clarence River, but unfortunately is not well displayed near public roads.

The North Canterbury Fold and Thrust Belt

The scenario outlined above for the Marlborough Fault System is still developing, because the zone of compression and widening between the faults is actively extending to the south. A new fault is developing at Porters Pass on SH73, and between it (the Porters Pass–Amberley Fault Zone) and the Hope Fault, the southernmost of the Marlborough faults, the area around Culverden and Waipara is being squeezed. The squeezing has not reached the extreme seen in Marlborough, where it forms the Kaikoura Ranges, but it is enough to be forming a fold and thrust belt, to raise whaleback low ranges of basement rocks, to depress intervening basins, and to cause interesting growing folds, which are deforming the surfaces of river terraces (Box 15.6 and Fig. 15.8).

This kind of terrain is called basin and range, and we will come across it again, on a bigger scale, in Central Otago (Chapter 18). In North Canterbury, the Waiau and Hurunui rivers have cut 'antecedent gorges' through the rising ranges, and meander across the intervening basins, on their journey to the sea. The gorges also meander, indicating that they preserve the meander pattern that existed before the ranges began to rise – superimposed drainage with 'fossilised' meanders.

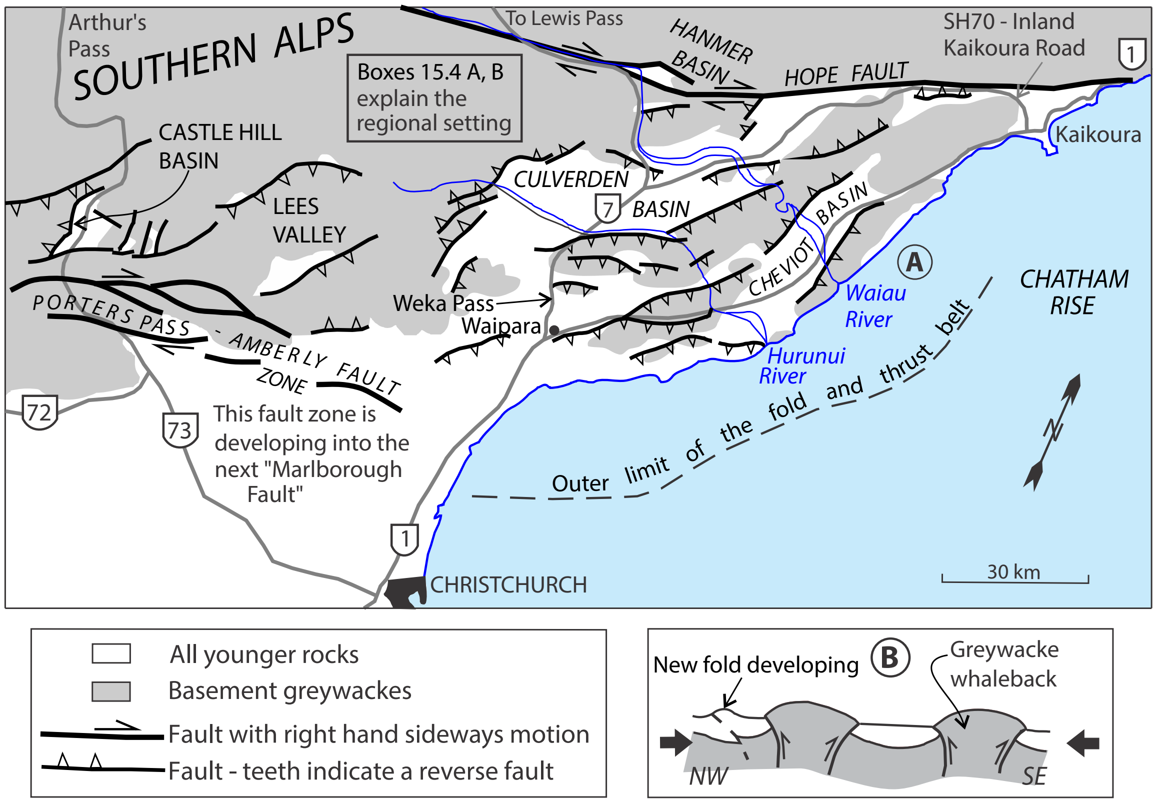
Cover strata in this area include two limestones: the lower 30 myr limestone found throughout New Zealand; and an upper one aged 15 myr, which is a local development. The two form a classic pair of dipping limestone escarpments at Weka Pass on SH7 (Fig. 15.9). There are fossiliferous sandstones in road cuttings on the seaward side of Weka Pass.

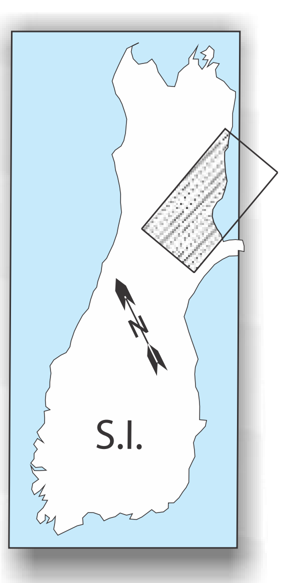
Hanmer Springs and the Hanmer Pull-apart Basin

The spa and ski town of Hanmer Springs lies in a basin at the foot of the Lewis Pass Road (SH7), which crosses the Southern Alps. The name refers to its hot springs, which in turn relate to the active Hope Fault (see above). There are many hot springs (27–60°C, not boiling) associated with the active sideways-moving faults of the Alpine Fault system (Box 9.5). They reflect the uplift of the Southern Alps, which brings hot rock near to the surface, while the faults provide the plumbing for surface water to circulate at depth.

The basin in which Hanmer Springs is located is New Zealand's best example of a kind of subsiding sedimentary basin called the strike-slip pull-apart basin. In our earlier discussion of the big bend in the Alpine Fault at Nelson Lakes National Park, we met the notion that a left step or bend on a right-moving lateral fault creates a region where opposite sides of the fault push against one another, causing uplift and mountains. Here we have the opposite scenario: a right step on a right-moving fault, creating a region where opposite sides of the fault pull away from one another. The resulting tension and extension allow the crust to subside, forming a basin. Boxes 19.2 A, B explain how it works, and shows the 1 km right step on the Hope Fault and the resulting 7 km-long basin.

Box 15.6. North Canterbury fold and thrust belt.





A

The rivers — note how the Waiau and Hurunui Rivers cross the structures. They cut gorges through the greywacke blocks, and spread out in wide shingle beds as they traverse the basins. That is, the rivers were established before the structures grew. They have been "superimposed" across the growing structures.

B

Think of the NE-trending ridges of greywacke as whale-back up-folds or anticlines, and the intervening basins as downfolds or synclines. The teeth on faults indicate the upthrust side of a reverse or thrust fault — one in which there has been overall shortening. Note how most greywacke whale-backs are bounded on both sides by outwards-moving thrusts, showing that the greywacke is spreading outwards over the basins to each side as it rises, as shown in the diagram. As explained in Boxes 15.4 A, B, this is because the region is caught in a closing vice, the jaws of the vice being the Chatham Rise on the east and the Alpine Fault/West Coast on the west. The growing structures extend offshore, and are extending southwards into the Canterbury Plains.

Editors' note: This chapter was written before the Christchurch Earthquake sequence (which was initiated in September 2010), where movement occurred on the Greendale Fault, parallel to and even further south, than the Porters Pass–Amberley Fault Zone.

Fig. 15.8. A view inside the Culverden Basin towards an approximately 3 m high fault scarp within the Hurunui Peak Fault Zone. Here, around 10 km northwest of Hawarden, the ground surface is the c.18,000 year old Masons Flat plain (foreground to middle distance). In this sector, which has been formed by the Waitohi River, the fault scarp is the product of at least two ground-surface rupturing earthquakes within that time period. Photo from David Barrell.

Fig. 15.9. View west over Weka Pass limestone escarpments in north Canterbury. Photographer Lloyd Homer, GNS Science.

There are other, smaller pull-apart basins along the Hope Fault, ranging in size right down to house-sized ponds. SH7 follows the Wairau River (which in turn is following the Hope Fault) for 25 km, before leaving it and turning right to climb to Lewis Pass. The broad basin seen as the road leaves the river here is another pull-apart.

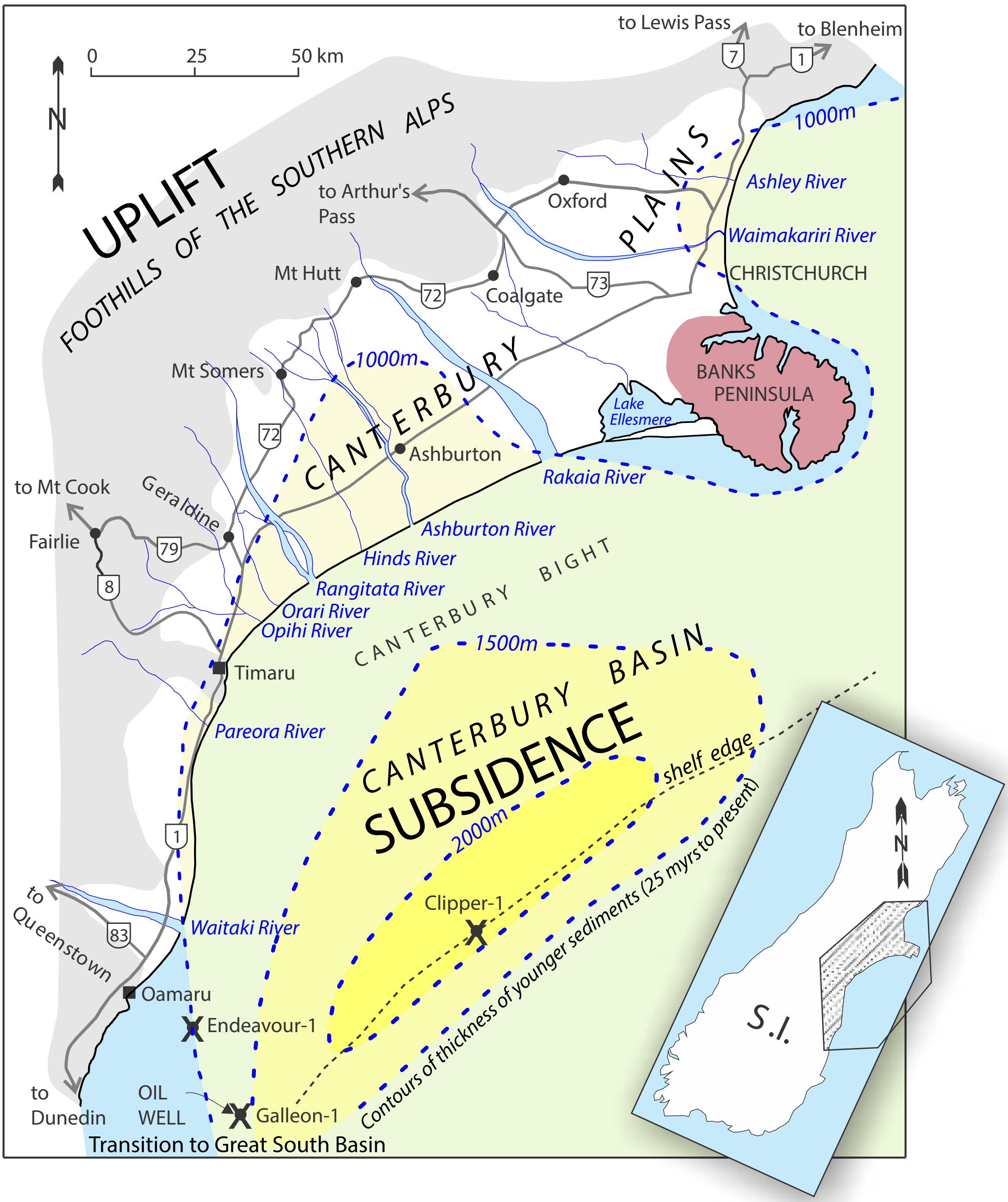
The Canterbury Plains

The Canterbury Plains (actually only one plain, despite the name) are one of New Zealand's best-known geological-cum-scenic icons (Box 15.7). Extending from Amberley to Timaru – a distance of 180 km – they are an apron, or piedmont, of river gravels, carried out from the Southern Alps by the big braided rivers (from north to south: Waimakariri, Rakaia, Ashburton and Rangitata) and built out across the underlying rocks as a gently seawards-sloping plain. A glance at the map shows that Banks Peninsula volcanoes (see below) have had a strong influence on the shape of the plains, by protecting the middle portion from direct wave attack and by allowing the plains to lap around the volcanoes.

Fig. 15.10. The Canterbury Plains seen here inland from Christchurch, have been built up by gravels eroded off the rising Southern Alps to the west.
Photographer Lloyd Homer, GNS Science.

Box 15.7. Canterbury Plains, Canterbury Basin.

The Canterbury Plains are the top layer of the Canterbury Basin, above sea level, comprising glacial and river gravels shed from the rising Southern Alps during the past 5 million years. (See the cross-section, Box 15.8)

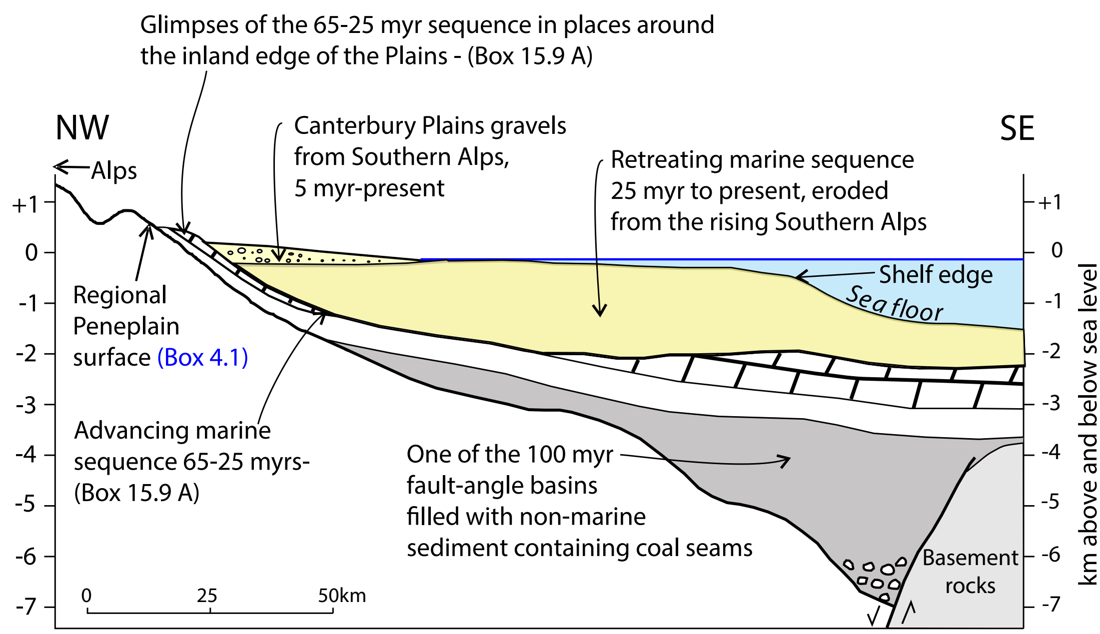


Box 15.8. Canterbury Basin.

The Canterbury Basin underlies the Canterbury Bight, with its deepest part (more than 6 km below sea level) located beneath the shelf edge. Like other important sedimentary basins around the country (Great South Basin, Box 15.14) it began life around 100 myrs ago as an extensional fault-angle basin (Box 6.6 G), a response to the crustal stretching that accompanied the beginnings of opening of the Tasman Sea (Box 5.5 A). The fault angle occupies the deepest part of the basin, and is filled with non-marine alluvial fan breccias (same as the Hawks Crag Breccia on the Buller River) and river sediments including coal seams. As elsewhere, the area slowly subsided after New Zealand separated from Australia and Antarctica, and a slow marine advance occurred towards the present-day Southern Alps (which were not mountains at that time), from around 65-25 myrs ago. At 25 million years ago the regime changed to one of compression across the country, the Alps began to rise, slowly at first, and sediment shed from them built out first the continental shelf, and later the gravel plains as the rate of uplift of the Alps increased about 5 myrs ago.

The cross section shows the three major rock units, and the map shows contours of sediment thickness for the topmost unit. A similar history is found for the past 100 myrs throughout most of the country, with local variations. See Boxes 15.9 A, B for more detail of the cycle of marine advance and retreat.

Oil exploration in the basin lead to the drilling of three exploration wells marked on the map. No commercial hydrocarbon accumulations were found.



Schematic cross section across Canterbury Plains and Canterbury Basin, showing the sequence of cover strata 100 million years to present. Vertical exaggeration 10x.

The best views of the Canterbury Plains are from aeroplanes circling to land at Christchurch airport, as many features are hard to see on the ground. Look for the common gravel pits, and the scroll patterns in the paddocks that mark the positions of abandoned braided river channels. The gravels are important aquifers, on which much of the agricultural irrigation on the plains relies.

The plains are being gently uplifted, as the neighbouring Southern Alps go up much faster, with the result that the inner margin of the plains is around 350 m above sea-level. This uplift has resulted in splendid river terraces alongside the big rivers (Box 10.3).

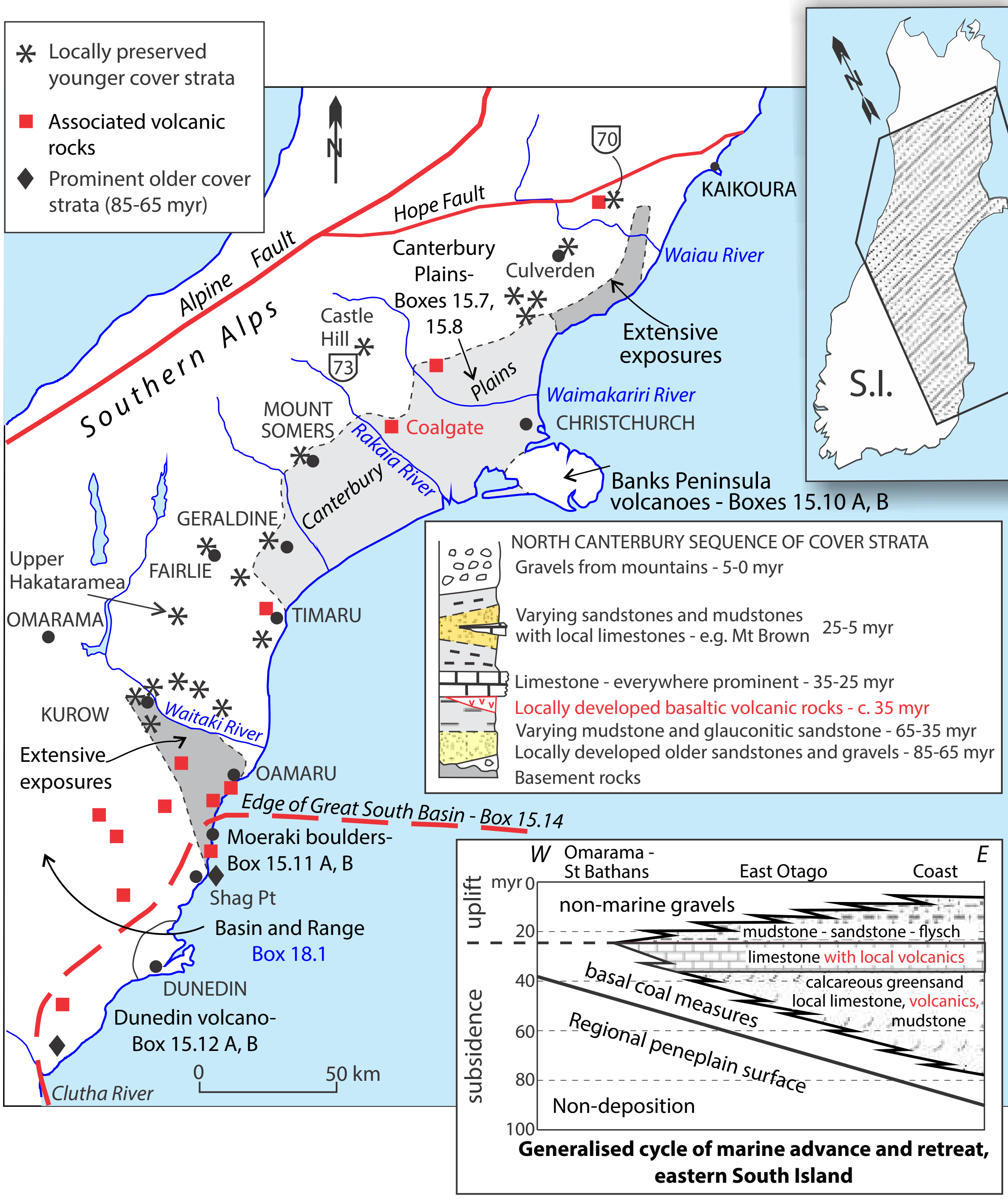
The Canterbury Basin

The Canterbury Plains are there only because there was a foundation across which they could build. In fact, the surface river gravels are simply the latest layer in a club sandwich of strata that date back 100 myr and comprise the fill of the Canterbury Basin, one of New Zealand's largest and longest-lived sedimentary basins (Box 15.8). Like the North Island's Taranaki Basin (Chapter 11) and the Great South Basin off the coast of Southland (Box 15.14), the Canterbury Basin began life as a rift valley on Gondwana in phase two of New Zealand's development, during the 30 myr interval between the start of the new spreading plate boundary and the actual first appearance of oceanic lithosphere that became the Tasman Sea (Chapter 4). Also like the other two basins, it was a 'failed arm' that did not proceed to ocean formation, and thus the coal-bearing strata deposited in the rift became a potential hydrocarbon source.

The Canterbury region went through the same prolonged period of gentle subsidence as the rest of the country during phase three of its formation (80–25 myr ago), and accumulated a similar sequence of marine strata. Although these are largely buried under the gravels of the Canterbury Plains or lie beneath the sea floor, they can be seen in small outcrops around the inner edge of the plains – windows into what is underneath the gravels (Boxes 15.9 A, B). They are seen more widely to the north and south of the Canterbury Plains, around Culverden in the north, at Castle Hill by the Arthur's Pass highway (SH7), and between Fairlie and Dunedin in the south. Further inland towards the rising Southern Alps, cover strata have, of course, long been eroded away – apart from the Castle Hill remnant.

There is one feature of that time interval in which Canterbury and Otago are unique, and that is in the presence of basaltic volcanic rocks interleaved in the sequence of strata from place to place. These vary widely in age (60–2.5 myr), and are recognised by their colour – black when fresh, brown when weathered. A good example is the famous cliff exposure of pillow lavas (Box 6.4 A) just south of Oamaru port (Fig. 15.11). Oamaru also gives its name to one of the many limestone strata dating from this time. Oamaru Stone is a soft freestone that can be sawn at any angle and is easily sculpted; indeed, the town has a famous historic precinct built largely from this cream-coloured stone (Fig. 15.12). Extensive outcrops of limestone are found inland from Oamaru, as far as Kurow, giving rise to a karst topography of bluffs, caves and sinkholes (Box 7.2 A, B).

Box 15.9 A. Cover strata – Oamaru Stone, other limestones and associated rocks of Canterbury and Otago. *(discussion in following box)*



Box 15.9 B. Cover strata – discussion.

Cover strata are relatively thin, younger rocks that lie on top of thicker, older, usually harder rocks (the basement). The cover strata of eastern South Island — Oamaru Stone and associated rocks — are the product of slow subsidence and marine advance (100-25 myr) followed by accelerating uplift and marine retreat (25 myr-present). Boxes 5.5 A-D put the cycle in context. Box 7.3A discusses the coastal plain swamp deposits (coal measures) that commonly begin the cycle, though they are generally not well developed in Canterbury and Otago.

Oamaru Stone is a fine-grained white limestone (rock made of calcium carbonate, $CaCO^3$). Unlike most limestones, which are hard and have bedding planes, seams and joint planes (Box 7.2 A, B) that govern how they can be cut and broken, Oamaru Stone is soft, uniform and homogeneous (a "freestone") and can be sawn into blocks of any size and shape. In Oamaru and elsewhere many buildings display its versatility: the "neo-classical" style of building was popular. It is a good sculpting rock, and is still in use for building, being exported as far afield as Auckland. Oamaru Stone is made of bryozoans (tiny marine moss animals) and is 60-70 metres thick. It is one of many limestone units, country-wide, which record the maximum marine advance between 40 and 25 million years ago (lower diagram in Box 15.9 A). The shelled organisms which form limestone dominated the continental shelf of the time because little mineral sediment was coming off the land area — it was small and well worn down (Box 4.1). Oamaru limestone is older than most, about 40 million years, and exists because local volcanic activity produced an elevated, sediment-free sea floor that was promptly colonised by moss animals. Limestone was deposited in a few areas later, during the marine retreat.

How to find where you are in the rock sequence : Look for the limestone — it forms bluffs and/or scarp-and-dip-slope profiles (Box 12.6) of pale-coloured rock, with overhangs and caves, and sinkholes on the top surface. Drainage may be underground. Exposed faces may show vertical fluting. Flagginess is rare in eastern South Island. As a secondary marker there are basaltic volcanic rocks (Box 6.4 A) conspicuous in places by brown weathering colours and forming some bluffs and ridges. Other cover strata, glauconitic sandstones (greensands), mudstones, sandstones and flysch (inter-layered sandstone and mudstone, Box 6.5 A) can be seen in cliffs and road cuttings. These rocks are followed everywhere by gravels recording the rise of high mountains from about 5 myrs ago, and buried by them underneath the Canterbury Plains. Cover strata are normally preserved from erosion only locally, typically in fault-angle depressions. (Boxes 15.5, 15.6, 18.1). The largest outcrop areas are near to the north Canterbury coast, and inland from Oamaru around Duntroon. The precise sequence of strata varies from place to place.

The volcanic rocks — while matching cover strata are found throughout New Zealand, the volcanic rocks are unique to Canterbury and Otago. Many small volcanoes were active in different places at around 60 myr, 40 myr, 15 myr and 2.5 myr. Like the much bigger and chemically different Banks Peninsula and Dunedin Volcanoes (Boxes 15.10 A, B; 15.12 A, B), the lava source was within Pacific Plate lithosphere (between 25-100 km deep). Just why this portion of the Pacific Plate periodically produced squirts of lava, over a long period of time, is a mystery.

Regional peneplain surface — slow subsidence and marine advance between 100-25 myr ago resulted in a country-wide erosion surface (a peneplain or near-plain) worn across all the different basement rocks. It is well displayed in central Otago and South Canterbury where it was exhumed when cover strata succumbed to erosion, and preserved because basement rocks are more resistant to erosion. Folding and faulting accompanying the recent rise of mountains cause the surface to slope at all angles, in all directions. Where uplift is fastest, and mountains highest (Boxes 4.1, 19.1 A, 19.3), the surface has been destroyed. In contrast to the generally sediment-starved nature of older cover strata, the sequence since 25 myr has received an ever-increasing supply of sediment, causing the coastline to retreat eastwards and culminating in voluminous outpourings of gravel. Boxes 5.5 A-D explain the change. In places, e.g. central Otago, extensive lignite immature coal) deposits formed during this phase, in sedimentary basins forming simultaneously with uplift of the Alps.

Fig. 15.11. Eroded cross-sections through late Eocene (about 35 myrs old) basalt pillow lavas can be seen in the coastal cliffs just south of Oamaru port. Width of photo 2.5 m. Photographer Bruce Hayward.

Fig. 15.12. Thirty million year-old Oamaru Limestone has been quarried for building stone for over 150 yrs and used in prominent stone buildings all around New Zealand and even in a few buildings overseas. Not surprisingly the best assemblage of Oamaru Stone buildings occur in Oamaru itself and include the National Bank (1871) and former Bank of New South Wales (1884) buildings in South Thames St. Photographer Bruce Hayward.

Timaru is built on the youngest and most extensive body of basaltic lava, 2.5 myr old and forming a low plateau. The dark basalt lava can be seen around the port, overlain by up to 5 m of wind-blown loess. Blocks of basalt have been used for cliff protection and in building breakwaters.

Again in common with the rest of New Zealand, Canterbury began receiving more sandstone and mudstone following the energising of the plate boundary 25 myr ago. The sediment was derived from the early manifestations of the Southern Alps, even though mountains as such did not appear until 5 myr ago. The new sediment built a thick, seawards-tapering wedge.

The situation has thus been set for the possibility of oil and gas in the Canterbury Basin. Old source rocks have been buried to more than 5 km depth, as shown in Box 15.8, and so should have been 'cooked' properly, and there are permeable potential reservoir strata in the younger rocks. Timing is critical: suitable traps – anticlinal folds or other types – must be in place before oil and gas are expelled from the source rocks. This may not have happened in the Canterbury Basin, for reasons connected with distance from the plate boundary and lack of compression to form folds, and so suitable anticlinal trap structures may not have been present at the critical times. Several exploratory wells have been drilled offshore (early ones are marked in Box 15.7 and more are likely), but so far without success.

There is more on hydrocarbon geology in Chapter 11, regarding the productive Taranaki Basin. The latter is situated at a similar distance from the plate boundary as the Canterbury Basin, and therefore we might expect that the two would show similar trap structures. The difference is that the Canterbury Basin is on the Pacific Plate, which is doing the pushing, while the Taranaki Basin is on the Australian Plate, which is being pushed. As a result, there has been more deformation and hence trap formation in the Taranaki Basin. The old rift-bounding fault shown at the bottom of the Canterbury Basin in Box 15.8 has not been compressed and reactivated in a reverse fashion, as such faults have in the Taranaki Basin.

Banks Peninsula – Two Basaltic Volcanoes

Before leaving Canterbury geology, we must visit the unique Banks Peninsula (Boxes 15.10 A, B). In a nutshell, Banks Peninsula is built of two large, adjoining volcanoes of basaltic composition, the older Lyttelton Volcano (active 12–9.5 myr ago) and the younger Akaroa Volcano (active 9–6 myr ago). The peninsula's two main harbours, Lyttelton and Akaroa, occupy the crater areas of the two volcanoes. They are not hot-spot volcanoes because they were active over a 6 myr span, during which time the Pacific Plate moved a long way – hence the lava source must have migrated with it. At present, the Banks Peninsula volcanoes are a geological mystery.

Box 15.10 A. Banks Peninsula – discussion.

Banks Peninsula
Banks Peninsula comprises two largely basaltic volcanoes, Lyttelton and Akaroa.
For more detailed information, see "Geology of Banks Peninsula". Institute of Geological & Nuclear Sciences Geological Map 3, 1992, scale 1:100,000, by R.J. Sewell, S.D. Weaver & M.B. Reay. Available from GNS Science, Box 30368, Lower Hutt.

Lyttelton and Akaroa were caldera volcanoes (having large collapsed crater areas). Lava flows from gently sloping surfaces on the hillsides, and in cliffs and cuttings you will see breccias and lavas with columnar jointing. You will also see a pervasive cover of loess (windblown dust) resulting from the position of the Peninsula downwind from the Canterbury Plains. Loess is commonly 4-5 metres thick, and as a result of rainwash in cuttings it can show flutings that somewhat resemble columnar jointing in the lavas.

Being old, and long inactive, the volcanoes are deeply dissected by river erosion, but retain their general shield volcano shape and radial drainage. The original caldera areas are now the site of Lyttelton and Akaroa Harbours, and the deeply indented coastline is the result of drowning of river valleys by sea level rise after the last glaciation 20,000 years ago (Box 15.13). The age of the volcanoes also means that surface weathering is well advanced, and brown iron-oxide colours are pervasive. You will see volcanic rocks in various stages of bleaching and alteration (Boxes 6.1 A, B).

Dikes of intruded lava up to several metres wide are generally arranged radially from the two volcano centres. They are best seen on Summit Road and in cliffs, where they commonly project as walls.

Lyttelton volcano rests on greywacke basement rocks (not well exposed) (Boxes 6.2 A, B), and on its own early white-coloured rhyolitic volcanic rocks, aged about 12 million years. Around Gebbies Pass you will see these massive, white-cream rhyolite lavas, flow-banded in places, and breccias. To confuse matters somewhat, there are also much older rhyolites and andesites here, related to the Mount Somers Volcanics in inland Canterbury, and aged about 90 million years. These rocks outcrop to the southwest of Gebbies Pass.

A road circuit which covers the essential rock and volcanic features is Christchurch-Lyttelton-Gebbies Pass-Summit Road-Christchurch. There are many other roads, and a walking circuit around the Peninsula. Similarly, a boat trip on Akaroa or Lyttelton Harbours is recommended.

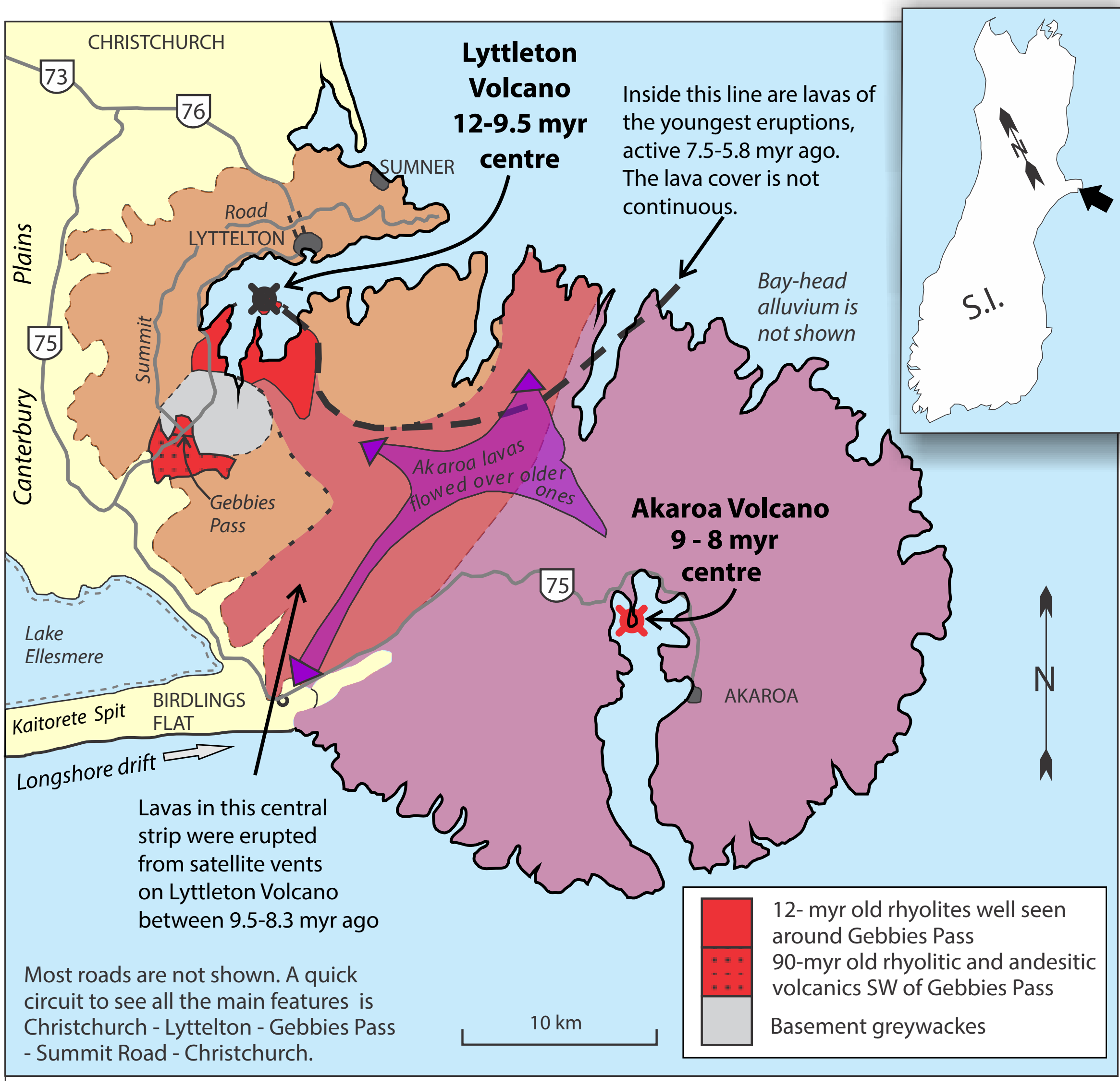
Banks Peninsula was an island, but has been linked to the mainland by the growth of the Canterbury Plains (Box 15.7). Lake Ellesmere lies adjacent to the Peninsula, caught between the Plains advancing from the northwest, and Kaitorete Spit advancing from the southwest by longshore drift.

Birdlings Flat lies at the northern end of Kaitorete Spit, and is well known among rockhounds for its concentration of semi-precious stones. The concentration arises because the spit is at the end of a long line that begins near Oamaru, a line of longshore movement of sand and pebbles towards the north, driven by the prevailing waves. During this northerly drift, most of the greywacke pebbles are ground down to sand, leading to greater concentrations of the rarer, harder, silica-rich stones like quartz, chert and jasper.

(continued on next page)

Box 15.10 B. Banks Peninsula – discussion *(continued from previous page).*

Why is Banks Peninsula here? It is a geological puzzle without an answer at present. If andesitic volcanoes were related to a deep-seated hotspot, like Hawaii or Yellowstone, movement of the Pacific Plate over the 6 million years of volcanism should have created a chain of volcanic centres more than 150 km long, but it didn't. All volcanism requires a localised source of extra heat, or water to facilitate melting, in the earth's mantle. In this case the source moved with the plate, and so must have been within the 100 km thickness of the plate, but we do not know what that source was. Geologically speaking, then, Banks Peninsula has no business being here, but it certainly is, and it greatly enhances the Christchurch district.



Moeraki Boulders

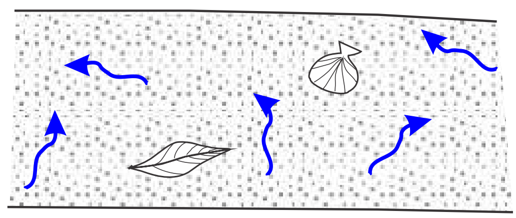
About 20 km south of Oamaru is Moeraki, whose large, spherical boulders, up to 3 m across, have become a famous tourist attraction. They are a fine example of the phenomenon of concretions formed in permeable sandstone layers following burial (Boxes 15.11 A, B and Fig. 15.13), and date from phase three of New Zealand's development. Similar concretions are found all over New Zealand.

Fig. 15.13. The Moeraki Boulders erode out of the soft sea cliffs about 30 km south of Oamaru. They are one of the best known examples of spherical concretions in New Zealand. Photographer Bruce Hayward.

Box 15.11 A. Moeraki – Concretions.

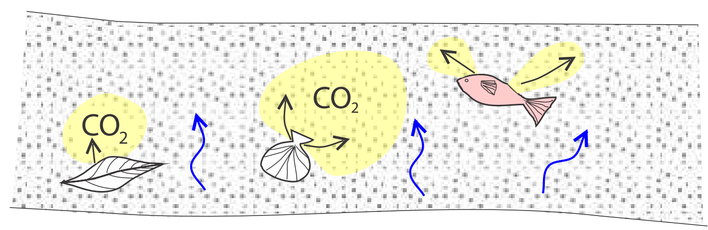
The celebrated Moeraki Boulders are excellent examples of a phenomenon called "concretions", but they are far from unique. Concretions are common in sedimentary rocks, and occur all over the country. Not all are spherical like the Moeraki examples, but they always have a rounded shape. This rounded form is original. It is not the result of abrasion on the beach or in a river. The concretions range in diameter up to 3 metres. Concretions form in sedimentary rocks after they have been deposited and buried, as follows:

(1) Layers of water-saturated porous and permeable sand or mud are deposited

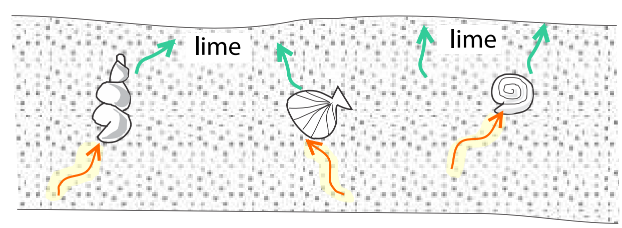


(2) After burial the pore water (up to 50% of the volume of the sediment) moves slowly through.

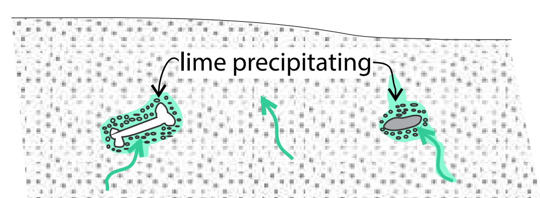
(3) Decaying plant and animal matter in the sediment releases carbon dioxide (CO_2) which dissolves in the pore water, making it slightly acidic (carbonic acid).



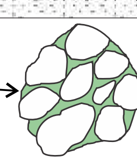
(4) Slightly acidic pore water corrodes objects like shells which are made of lime (calcium carbonate - $CaCO_3$) and carries away some lime in solution.



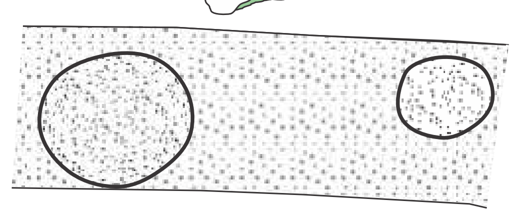
(5) Lime-charged pore water encounters suitable slightly alkaline conditions, commonly around bone, shell or mudstone fragments. The lime precipitates as small crystals filling the pore spaces.



The sand or mud is now firmly cemented into hard sandstone or mudstone <u>concretions</u>, surrounded by soft, uncemented sand or mud.



(6) In a uniform sediment concretions grow uniformly and become spherical or near-spherical, as at Moeraki; and more than 1 metre across.



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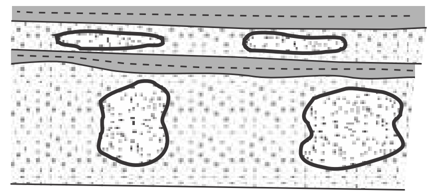
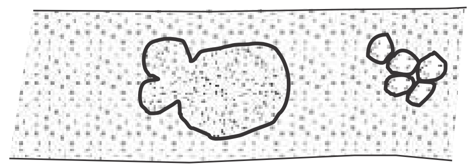
Box 15.11 B. Moeraki – Concretions *(continued from previous page).*

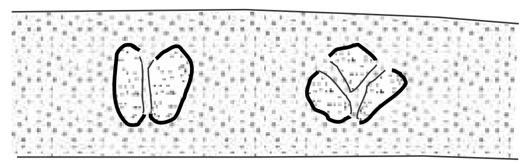
(7) In non-uniform settings concretions may be pancake or bun-shaped, or elongate like rods.



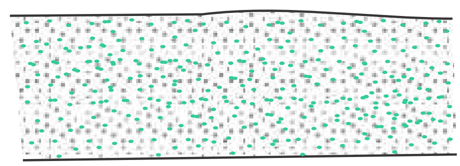
(8) Adjacent concretions can join or make popcorn- like aggregations



(9) Paramoudras - concretions that grow around mud-filled worm burrows - may end up with hollow centres.

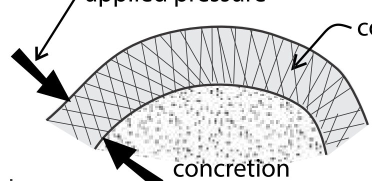


(10) If the concretion-forming process continues, the entire bed of sandstone or mudstone may be cemented by lime.

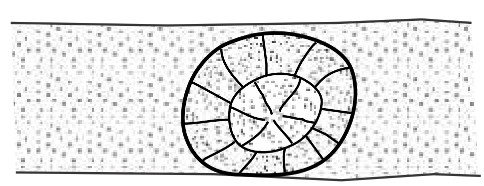


(11) The most common mineral found as the concretion cement is lime, calcium carbonate, or calcite. Other minerals that form cements are the iron-magnesium-calcium carbonate ankerite, and the iron carbonate siderite; these concretions tend to be brownish in colour (e.g. ankerite concretions at Shag Point, south of Moeraki).

(12) Growing concretions exert some pressure on the enclosing sediment, and this can give rise to a surrounding layer of fibrous calcium carbonate or siderite with a distinctive fibre-crossing "cone-in-cone" arrangement. This is an expression of the conjugate shear pair pattern that tends to develop in all matter when it is squeezed.



(13) At a late stage in their formation, perhaps when the confining pressure of the surrounding rocks is reduced, concretions tend to open up a little, forming a system of radial and concentric cracks. There is still circulating pore-water in the surrounding sediment, and this penetrates the cracks and lays down crystalline lime (calcite). These "septa" are commonly 1 to 2 cm across, and contain calcite crystals of similar size. Such concretions are called "septarian", and tend to break along the septa.



(14) Finally, when the rocks containing the concretions are exposed by erosion, the much-harder concretions are released by removal of the soft, surrounding sediment, as is happening at Moeraki.

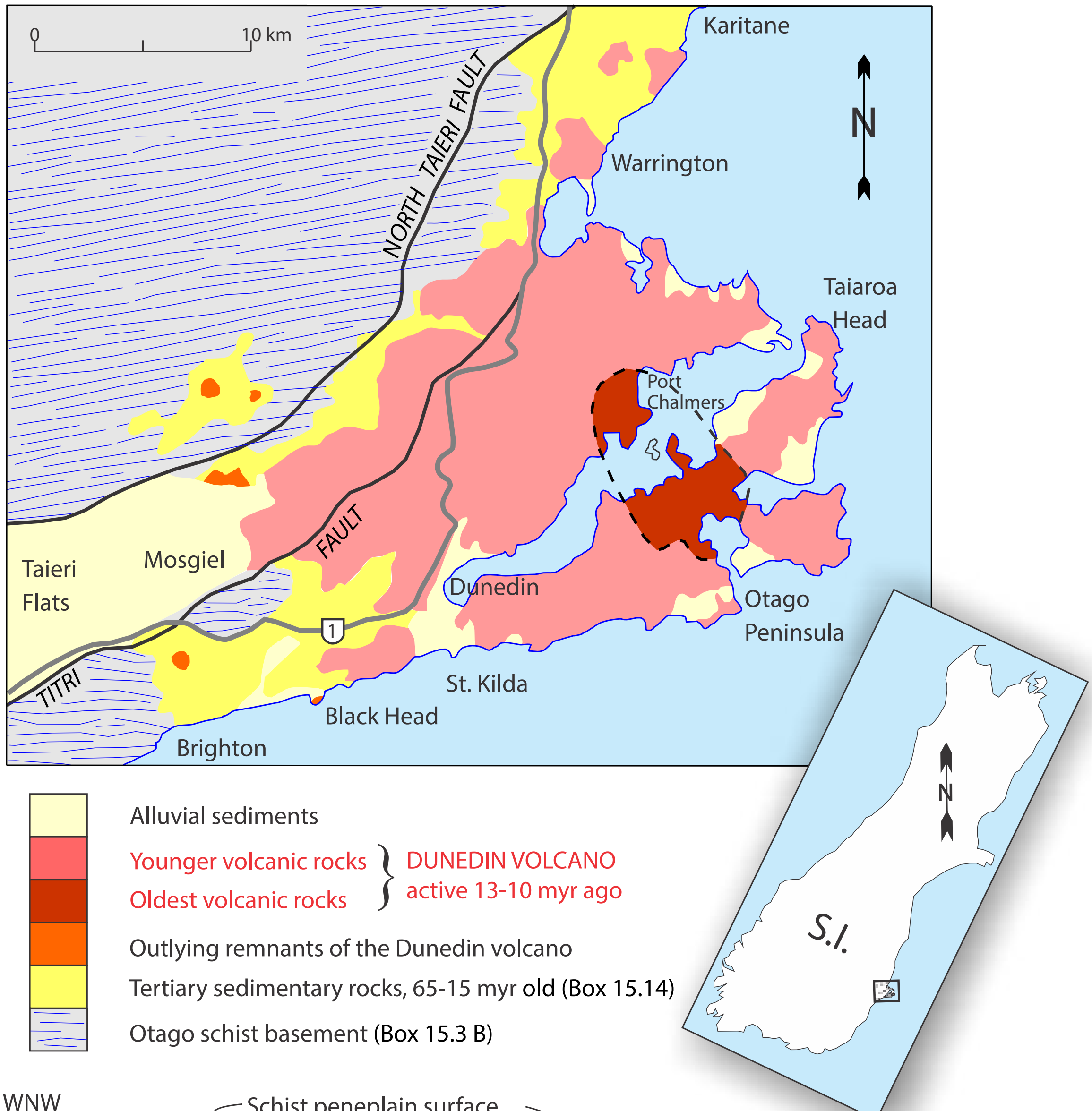
Geology of Dunedin

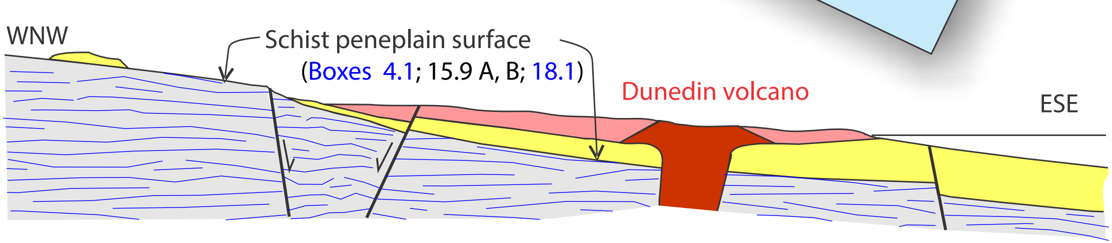
Dunedin is built partly on cover strata (including some of the country's older ones, dating from 65–70 myr ago), and partly on the Dunedin Volcano (Boxes 15.12 A, B), which has a lot in common with the Banks Peninsula volcanoes further north (see above). In both cases, the volcanoes have formed a prominent eastward bulge in the coastline, and they each have long, narrow harbours that cut through the old volcanic craters. They are similar in composition, and they overlap in age: 12–6 myr for Banks Peninsula, and 13–10 myr for Dunedin. Like the Banks Peninsula volcanoes, the Dunedin Volcano moved with the Pacific Plate, and there is no connection with subduction. It, too, is a geological mystery.

Basement rock hereabouts is Otago Schist, which forms the hills to the northwest and southwest of Dunedin city (Box 15.12 A). Those two ranges, and the intervening fault-bounded Taieri flats or Mosgiel depression, are the southeasternmost portion of the Central Otago basin and range province, a highly distinctive geological region that is described in Chapter 18.

Fig. 15.14. Aerial view of Dunedin Volcano. The view is northeast along Otago Harbour that cuts right through the middle of Dunedin Volcano and its crater, which was situated near Port Chalmers (near islands in middle of harbour).
Photographer Lloyd Homer, GNS Science.

Box 15.12 A. Dunedin Volcano.





Schematic cross-section through Dunedin Volcano, vertical scale exaggerated. The original volcano landform has been destroyed by erosion.

Box 15.12 B. Dunedin Volcano – discussion.

There are two major volcanic constructions in the South Island, Banks Peninsula (Box 15.10 A, B), and Dunedin Volcano. Dunedin Volcano is an old shield volcano, active between 13-10 myrs ago. The lava had basic to intermediate composition (Boxes 6.4 A; 15.10 A, B), and the activity produced a wide range of products: lava flows with columnar jointing, ashes and breccias, explosion breccias in craters, intrusive dikes and sills (perhaps intrusive into the sedimentary rocks that underlie the volcano), and lava domes.

Being old, and long inactive, the volcano has been extensively dissected by erosion. A wide range of exposures can be found around the shores and road cuts of Dunedin city and the area north and south of the harbour. The oldest part of the volcano lies around Port Chalmers and Portobello. Note that in many exposures the rocks are deeply weathered (Boxes 6.1 A, B); look for "onion skin" spheroidal weathering and red-to-brown colours.

Noted Dunedin vantage points include Soldiers Monument on Otago Peninsula (lava), The Pyramids on Otago Peninsula and the Organ Pipes (columnar-jointed lava). Look for the dark lavas and grey breccias in many Dunedin buildings, often in association with white Oamaru stone. The lavas are quarried for aggregate (e.g. Black Head).

Separating the Dunedin volcanic rocks from the basement schists (Box 15.3 B) are a sequence of sedimentary rocks. They record the subsidence and marine transgression that buried the Otago Peneplain between 100 and about 25 million years ago (Boxes 4.1; 15.9 A, B). The lowest levels are coal-bearing river and swamp sediments, followed by shallow marine sands, mudstones and limestones, with glauconite ("greensand") and in places phosphate. These particular Cretaceous and Tertiary sediments (as far south as Kaitangata) offer the only onland glimpse of the Great South Basin (Box 15.14) which is otherwise entirely offshore. Onland they comprise a thickness of between 2,000 and 3,000 metres, and the country they underlie tends to be landslip-prone (e.g. the famous Abbotsford landslide), especially where sandstone overlies mudstone. As elsewhere, limestone forms bluffs and ledges.

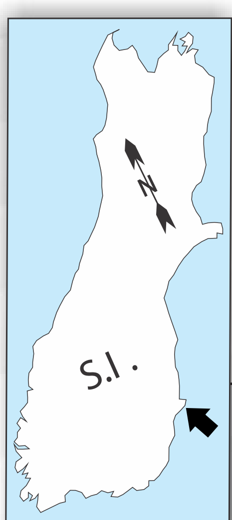
Coal is mined at Kaitangata (in 2000) and was formerly mined south of Dunedin. Limestone, quartz sand, clay and marl are or have been quarried.

Further reading: Geology of the Dunedin Area, compilers D.G. Bishop and I.M. Turnbull, Institute of Geological and Nuclear Sciences 1:250,000 geological map 21. One map sheet and 52p. booklet. 1996. Box 30368, Lower Hutt.

Box 15.13. Drowned valleys.

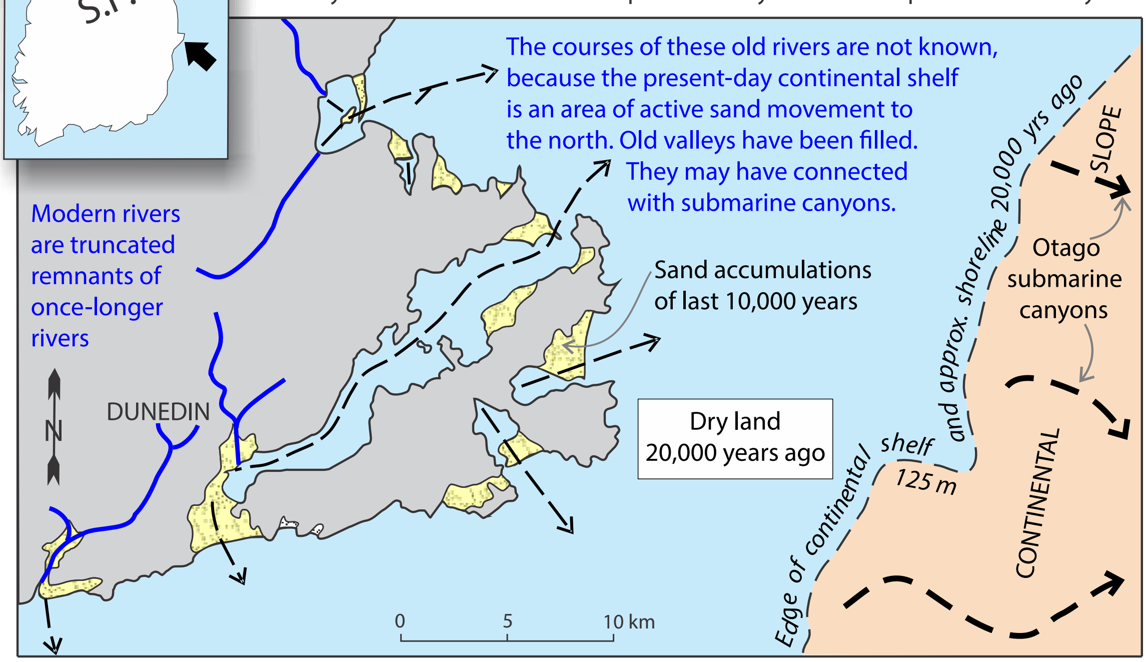
Most of the world at present has a "drowned" coastline. The cause is the glaciations. At the height of a glaciation (the last two were 20,000 and 150,000 years ago) the large quantity of water locked up in continental ice caps (mostly in the northern hemisphere) causes sea level to fall by 120 to 130 metres, exposing most of the continental shelf. Rivers excavate their valleys to that low base level. The valleys are then "drowned" by the post-glacial rise in sea-level, giving us our many harbours and coastal embayments.

The map illustrates the effects around Dunedin. In general, drowned valleys are not conspicuous around the South Island. This is because of the huge sediment supply arising from active uplift of mountains, and associated high rainfall. Most drowned valleys have been infilled in the past 10,000 years by sand and gravel delivered by rivers and longshore drift (Boxes 14.2, 14.3).



Exceptions are found in Fiordland, where the fiords are glacially deepened drowned valleys (Box 17.2 A) and modern rivers and sediment supply are small; around the Dunedin and Banks Peninsula old volcanoes (Boxes 15.10 A, B; 15.12 A, B), which are largely isolated from the large sediment supply; and in the Marlborough Sounds, where again the sediment supply is limited and there is the additional factor of tectonic subsidence to enhance the drowned valley effect (Box 15.2).

Sea level rose from minus 130 metres to its present level between 20,000 and 10,000 years ago, as ice sheets melted. That is an average rate of sea-level rise of more than one metre per century (it may have happened in fits and starts). That rate of rise is important in the context of the current debate about global warming. The up-and-down cycle of sea level has been repeated many times in the past few million years.



The Taieri River, west of Dunedin, is deeply carved into the schist, forming a narrow valley between broad areas of the peneplain surface (Chapter 18). This is clearly seen from the scenic Taieri Gorge Railway excursion from Dunedin to Middlemarch. The Taieri River discharges into the Mosgiel structural depression, and then in order to reach the sea it cuts a gorge through the coastal belt of schist. This belt is actively rising now, and the Akatore Fault that crosses the Taieri just offshore of its mouth is regarded as active. Hence the gorge is antecedent, i.e. formed by the river downcutting as fast as the rocks rise. However, the rate of downcutting is barely keeping pace with uplift, giving the river a very low gradient on the Taieri flats within the Mosgiel depression. This was the reason for a prolonged flood episode that closed Dunedin airport for more than a month in the 1990s.

The Great South Basin

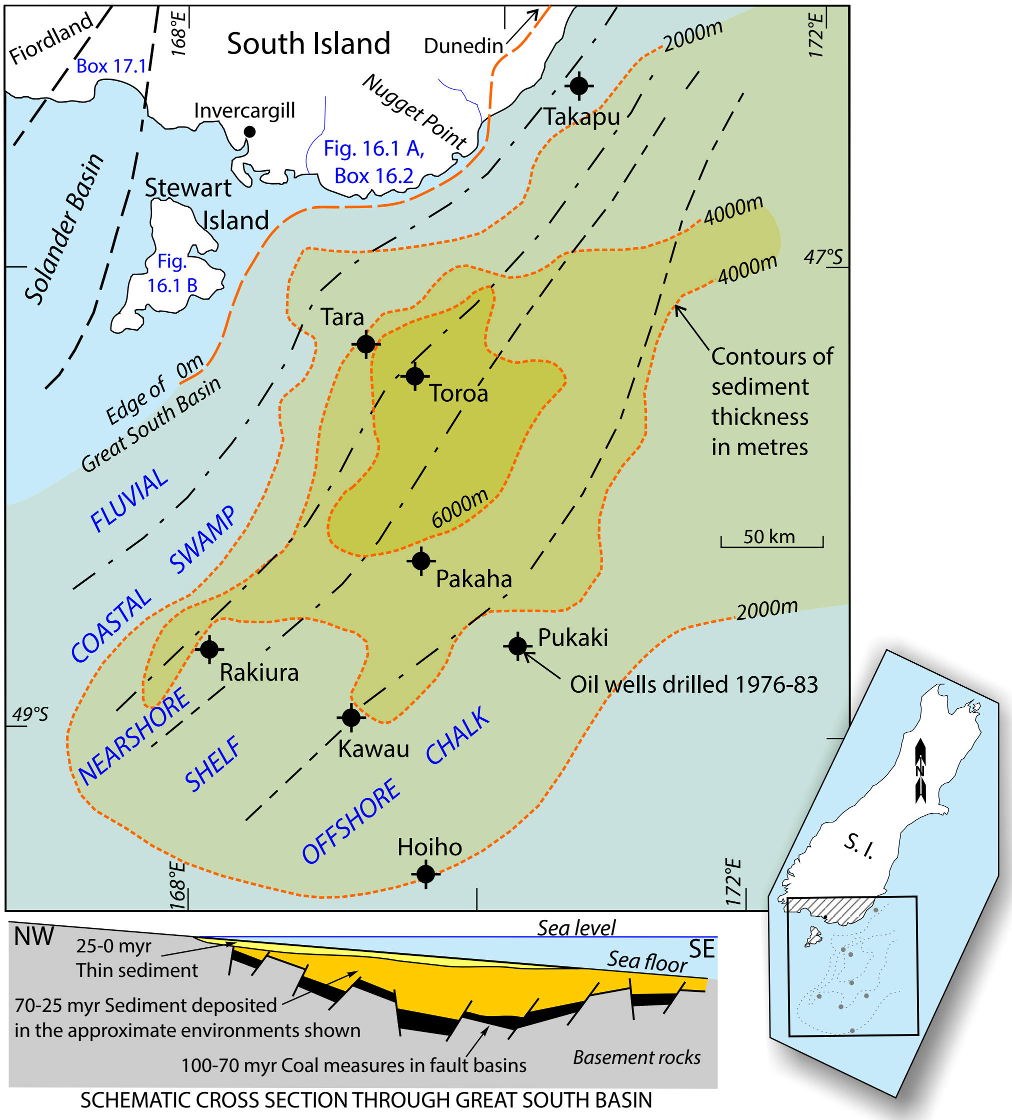
The cross section in Box 15.12 A shows how the thin veneer of cover strata that crops out around Dunedin is the feather-edge of a package of strata that thicken to the southeast – into the Great South Basin.

As Box 15.14 shows, the Great South Basin is big – an area measuring roughly 600 km by 350 km. It has exactly the same history as the Canterbury Basin (see above) and Taranaki Basin (Chapter 11), but was even further away from that vitally important supply of sediment derived from the vicinity of the plate boundary from 25 myr onwards. Nevertheless, strata are thicker than 4 km in the centre of the basin, which is regarded as prospective by oil geologists.

Slope instability and the Abbotsford Landslide

The cover strata around Dunedin comprise a wide variety of rocks, including coal measures, limestone and unconsolidated sandstone. They are mostly weak rocks and as such are prone to slope failure. The infamous Abbotsford landslide of 8 August 1979 involved a gently dipping bed of water-permeable sandstone that slid on a layer of impervious mudstone underneath it, when a quarry in the sandstone, at the base of the hill, removed the toe support that had been preventing the sliding. The block of sandstone was displaced downhill by many metres, opening up a rift that ran through a housing estate, in what was a classic example of a bedding-plane slide. Fortunately, no one was hurt, although the damage costs exceeded NZ$10 million.

Box 15.14. Great South Basin.



SCHEMATIC CROSS SECTION THROUGH GREAT SOUTH BASIN

This potentially important offshore sedimentary basin comes on shore only around Dunedin (Box 15.12 A, B) where you can see, between Shag Point and the Clutha River mouth, a complete selection of the sedimentary deposits of the basin (except where they have been covered by the Dunedin Volcano).

Great South Basin was the subject of intensive exploration by Hunt International Petroleum Company in the mid 1970's, and by Phillips Petroleum in the early 1980's. Several exploratory offshore wells were drilled, but no commercial accumulations of gas or oil were found. The map shows a highly generalised picture of the basin, summarizing the results of the petroleum exploration.

The sedimentary sequence found in the Basin is the result of the same cycle of marine advance (100-25 myr ago) and retreat (25 myrs to the present) as in Canterbury and Otago (Boxes 15.7, 15.8; 15.9 A, B). But there is one big difference: because of the greater distance of Great South Basin from the Alpine Fault, there has been only a small supply of sediment reaching the Basin from the uplift of the South Island mountains (see the cross section of the Basin). This has been the big problem in the petroleum exploration— in general there has not been sufficient burial to cook the coal measures and make hydrocarbons.

The raw material of oil and gas is organic carbon, dispersed through sedimentary strata or in coal seams. The cooking process requires temperatures of 100° to 120°C, which under normal geothermal gradients (25°C per kilometre depth) needs burial of at least 4 km. It is often more than 4 km because geothermal gradients are commonly reduced in sedimentary basins.

Chapter 16

Southland and Stewart Island/Rakiura

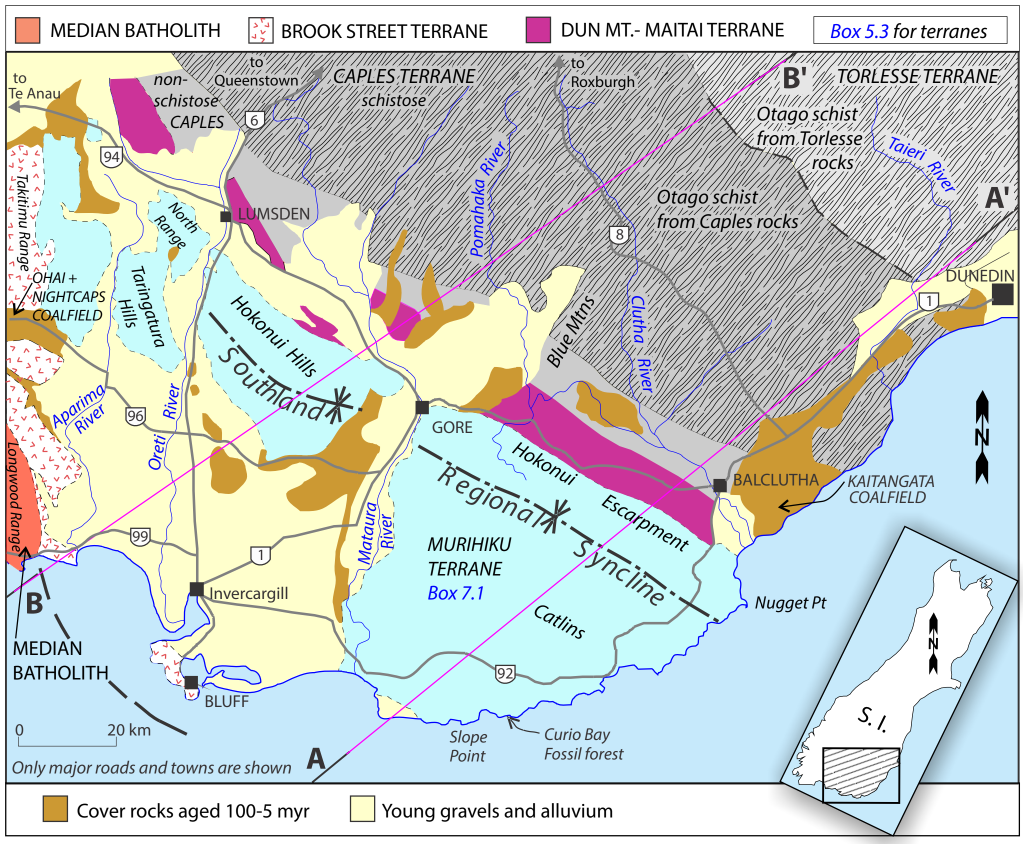


Fig. 16.1 A. Simplified geology map of Southland. Cross sections A–A′ and B–B′ are shown in Box 16.1.

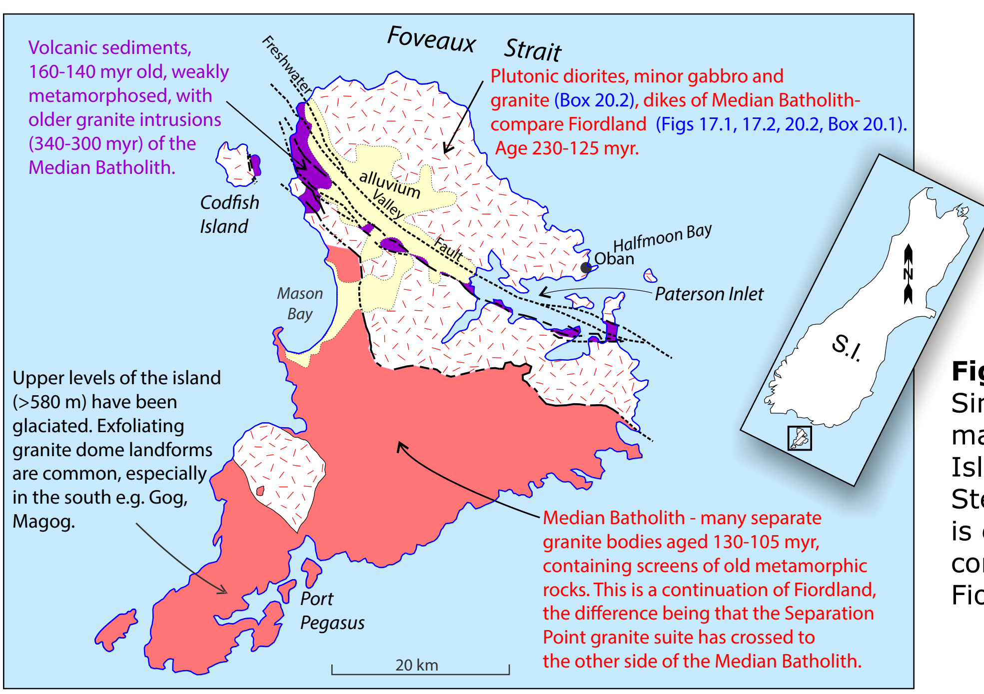


Fig. 16.1 B. Simplified geology map of Stewart Island. Stewart Island is essentially a continuation of Fiordland geology.

Southland and Stewart Island/Rakiura are best thought of as being made up of six major strips or terranes of old basement rock (Boxes 5.1, 5.2, 5.3), trending from northwest to southeast, and overlain in places by thin cover strata aged between 100 myr and about 5 myr. Extensive young gravels and alluvium overlie both the basement rocks and cover strata in places, reflecting uplift of mountains, and glaciations, over the past few million years.

Basement Terranes

In Chapter 5 we looked at the origins of New Zealand's basement terranes, and in Chapter 7 we encountered two specific examples: the Murihiku Terrane and the Torlesse Terrane (broadly speaking, the latter comprises most of our greywacke rocks; they can be subdivided, but we don't need to concern ourselves about that in this chapter). In Chapter 15 we described greywacke's metamorphosed equivalent, schist, which we meet again here, along with some of the other terranes. These are covered below in order from northeast to southwest, starting at Dunedin (Fig. 16.1 A and Box 16.1).

Otago Schist

Schist is not an original terrane, but a metamorphic rock derived from greywacke. In the case of Otago Schist, it resulted from the collision (at the Gondwana margin) between the Caples and Torlesse greywacke terranes. Otago Schist crops out over wide tracts of inland country, but is seen at the coast only between Dunedin and the Kaitangata Coalfield by the mouth of the Clutha River.

Schist is a strongly layered rock (Boxes 15.3 A, B), and the layering (schistosity) varies in angle and direction of dip from place to place. It is also strongly jointed, like most rocks, hence landforms on schist tend to be of the scarp and dip-slope variety (Box 12.6) and box-shaped. Where dips are steep, bands of harder schist form ribs up the hillsides. Across the region, look for schistosity layering; for lustrous white mica grains lying on the schistosity and reflecting the sun; for spherical red garnet crystals in places; and for white quartz veins cutting across the schistosity, these crumpled in places.

Fig. 16.1 A shows a line separating schist of Torlesse greywacke origin from schist derived from Caples greywacke. Don't expect to fall over it while walking, however, as it's a subtle distinction based on the chemistry of the schist, which in turn is derived from the different chemistry of Caples and Torlesse greywackes.

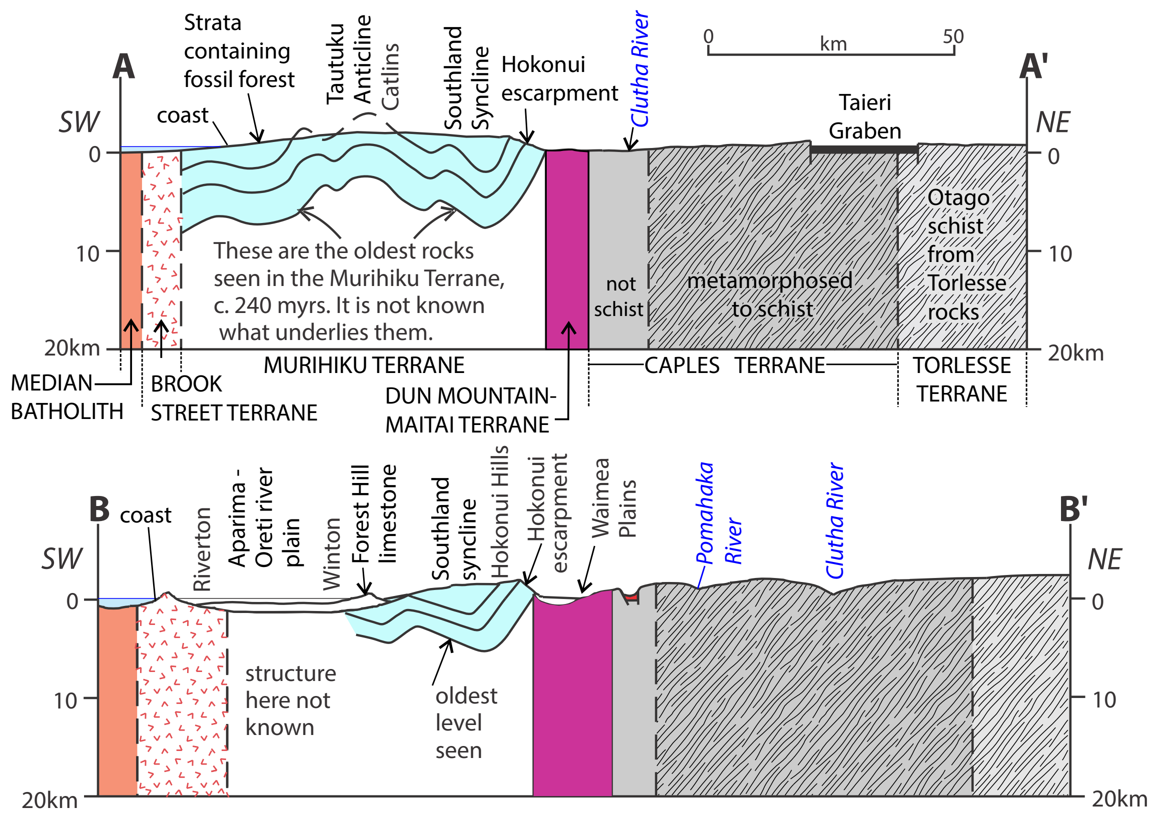
Caples Terrane

Caples rocks are a typical greywacke association (Boxes 6.2 A–C, Fig. 6.7). The reason they are differentiated from the more extensive Torlesse greywacke is because Caples sandstones contain more minerals and rock fragments of volcanic derivation, and fewer quartz grains.

Caples greywacke rocks are not readily accessible, except in the hills immediately north of Balclutha. They stay remote as they track northwestward and then northward to the Humboldt Mountains on the west side of Lake Wakatipu.

Box 16.1. Schematic cross sections across Southland.

Locations marked on the map (Fig. 16.1 A). Structure shown within Murihiku Terrane is real. Structure within other terranes is generally complex. Note that contact between terranes are steep, but are probably not as simple as shown here. Likewise we do not know what is happening in the lower levels, except that it will certainly not be as simple as shown here. Continental crust in Southland is between 25 and 30 km thick.



Dun Mountain-Maitai Terrane

The Dun Mountain part of this name refers to the olivine/serpentine rocks of Dun Mountain and Red Hills in Nelson, and also the Red Hills of western Otago. These are the same belt of rocks, now separated by 480 km of sideways movement on the Alpine Fault (Box 5.3). The Dun Mountain is an obducted terrane, comprising part of the oceanic lithosphere of the old Phoenix Plate, which for some reason was pushed up onto Gondwana instead of diving down under it, as was normally the case.

The Maitai part of the name (also derived from Nelson, in this case the Maitai River) refers to associated marine sedimentary rocks dating from 300–200 myr (Permian and Triassic periods). The combined terrane is continuous from the Alpine Fault in western Otago to the coast at Balclutha, but the proportion of serpentine and associated igneous rocks dwindles eastwards, and they are sparse east of Lumsden. As a consequence, the terrane forms low ground adjacent to the prominent Hokonui Escarpment (Fig. 16.1 A and Box 16.1). The foot of this escarpment marks the tectonic contact (terrane suture) between the Murihiku and Dun Mountain-Maitai terranes.

Murihiku Terrane

The Murihiku Terrane originated as the sedimentary filling of a forearc basin associated with the subduction zone at the Gondwana margin, and ranges in age from 240 myr to 150 myr (Mid Triassic to Late Jurassic periods). The rocks are sedimentary, but consist largely of volcanic materials from the adjacent volcanic arc. We do not know the exact location of this forearc basin, and the original parent arc itself is not preserved either, except perhaps as part of the Median Batholith (see below).

Today, the terrane forms a large tract of Southland (Fig. 16.1 A), and like the Dun Mountain-Maitai Terrane is also found in Nelson, as well as in western North Island (Box 5.3). In Southland, it narrows markedly towards the northwest, as it approaches the Alpine Fault and is smeared out by shear due to the fault.

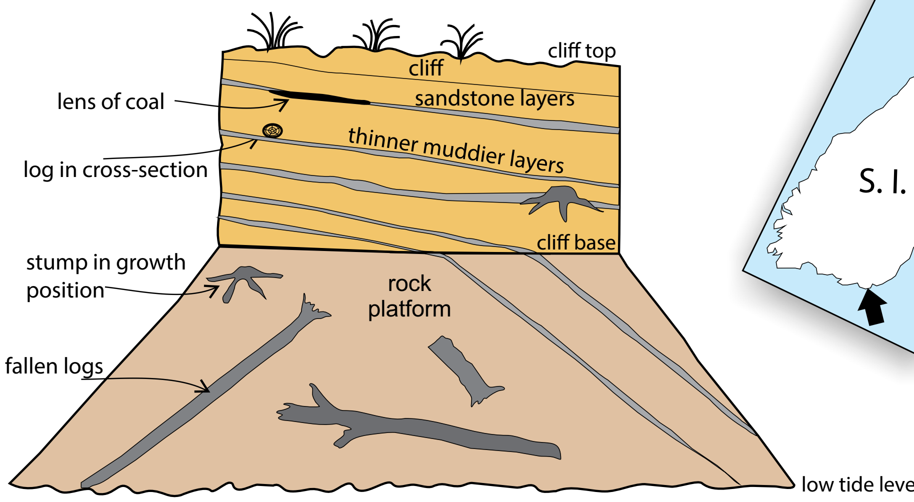
The main features of Murihiku rocks are a great thickness (more than 10 km) of well-stratified mudstones, sandstones and conglomerates, and huge fold structures (Box 6.6 A). The strata have been crumpled into anticlines and synclines that are several kilometres wide and high – the folds shown in the cross sections in Box 16.1 are the actual folds.

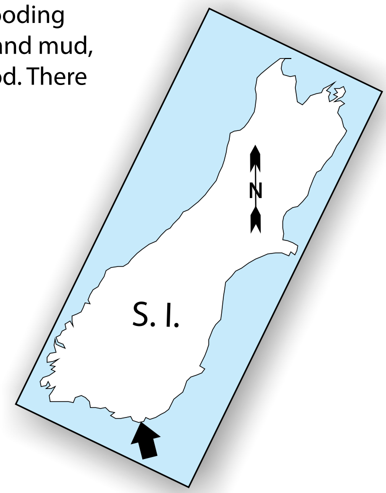
Murihiku rocks make up the Catlins, Hokonui Hills, Taringatura Hills and North Range. Look for bluff-forming sandstone and conglomerate beds, and well-developed scarp and dip-slope profiles (Box 12.6) throughout the area of outcrop. With care, you will be able to figure out some of the folds, from the differing directions in which the scarp slopes face (e.g. along SH9 between Clinton and Mataura), or from the differing dips visible in coastal cliffs. There are many striking cliff exposures on the southeast coast between Nugget Point and Slope Point (accessed from SH92), and fossils of both shellfish and plants can be found in places. The best-known of these is the fossilised forest at Curio Bay, Waikawa (Box 16.2).

Fig. 16.2. Strike ridges of more erosion-resistant sandstone beds within the Murihiku sedimentary sequence that forms the Hokonui Hills, northwest of Gore, Southland. Here the strata dip northeast (left) under the Waimea Plains. Photographer Lloyd Homer, GNS Science.

Box 16.2. Fossil forest, Curio Bay, Waikawa, Southland.

Origin of the forest — it grew on a sandy river delta plain. Continual flooding and subsidence of the area allowed preservation of the layers of sand and mud, and waterlogging of the sediment excluded air and preserved the wood. There is actually a succession of forests.





Look for small plant fragments in the form of black carbon blotches. Leaf shapes may be discernible.

Age of the forest — Jurassic Period, about 170 million years ago.
Type of trees — there were no flowering plants or grasses at this time. The trees were gymnosperms, both primitive conifers related to the living kauri; and podocarps, related to the living totara and rimu. There were also tree ferns.
Other localities — similar forests are preserved in several other places in Southland — e.g. Hokonui Hills, Mokoia, Mataura Falls and Owaka River. If you visit Slope Point, the southernmost point of the South Island, just west of Curio Bay, look for rare plant fossils around the light. The sediments are pebbly at Slope Point. The Curio Bay forest grew in a location close to the Antarctic Circle. The world had an equable climate at that time, without ice caps at the poles. In other places, forests and a reptile fauna flourished at near-polar locations, and the biodynamics of such forests are a matter of great biological interest.

Brook Street Terrane

These rocks form the Takitimu Mountains, and extend to Riverton and Bluff. They are rocks of an old oceanic volcanic arc, of basaltic composition and comprising a wide range of lavas, dikes, volcanic breccias and sandstones, along with their deeper intrusive equivalents ('granites'). The terrane comprises mostly lavas and sediments in the Takitimus, the equivalent gabbroic plutonic rocks at Bluff (seen on the coastal walkway) and fine-grained diorites at Riverton Rocks. Their age is around 250 myr (Permian to Triassic periods). The name Brook Street is again derived from Nelson, where these rocks also occur, but on the opposite side of the Alpine Fault.

This juxtaposition of an old arc (Brook Street Terrane) with an old forearc basin (Murihiku Terrane) is what would be expected from their original relationship. But they do not, in fact, fit together – their ages, chemistry and original contexts are different. This is one of the kinds of situation that gave rise to the tectonostratigraphic terrane theory (Chapter 5). The Brook Street Terrane was added to the Gondwana margin before the Murihiku, although we have yet to figure out just where they both originated.

Median Batholith

A batholith is a body of plutonic igneous rock, its name derived from the Greek words *bathos* (meaning 'depth') and *lithos* (meaning 'rock'). To drive the point home, the adjective plutonic and noun pluton (itself a synonym of batholith) are derived from Pluto, the Greek god of the underworld. Whichever term you use, these are bodies of magma that were not erupted, but cooled slowly at depth, allowing large crystals to grow. They are the roots of old volcanic/magmatic arcs, and can be very large, e.g. the Coast Range Batholith, which runs south from Alaska through British Columbia to Washington state. The rock name 'granite' is used as a catch-all term for the wide variety of different plutonic rocks (they differ because of widely varying chemical composition, which is reflected in the different mineral associations we see).

In the case of the Median Batholith (formerly called the Median Tectonic Zone), it comprises the roots of the magmatic/volcanic arc that resulted from subduction of the old Phoenix Plate underneath Gondwana, around 380–100 myr ago (between the Devonian and Cretaceous periods). A remarkable feature of the Median Batholith is that it is long and relatively narrow, implying that the magmatic/volcanic arc stayed in the same location for some 280 myr, while the Gondwana margin widened by accumulating terrane after terrane (Chapter 5). As shown diagrammatically in Boxes 5.1, 5.2, 5.3, for this to have happened the tilt of the subducting plate must have lessened each time a new terrane was added to the Gondwana margin, in order to keep the 100 km depth contour in the same location. There would, of course, have been an extensive volcanic superstructure to the batholiths. A few fragments are preserved, but most of it has been eroded away in the 105 myr since activity ceased. That same erosion has exposed the batholiths – a number of separate ones make up the terrane.

The Median Batholith is part of the second half of the subduction history at the Gondwana margin. That subduction and terrane accretion story is explored more fully in Chapters 20 and 21.

In Southland, the Median Batholith is seen in the western Longwood Range, where it comprises 'granites' of wide-ranging chemical composition – from granitic (>70% silica) through dioritic (~60% silica) to gabbroic (~50% silica) (Box 8.1). Dates obtained from the batholith in Southland range between 290 myr and 150 myr (Permian–Jurassic period). Easily seen examples of these rocks are fine-grained granite at Oraka Point (Colac Bay), diorite at Wakaputa Point and gabbro at Ruahine Bay.

Cover Strata

As is the case all around the country, basement (Gondwanan) rocks are overlain by younger cover strata. In Southland, these cover strata are closely similar to those of Canterbury and Otago (Chapter 15), although with a couple of differences. One is that there are no volcanic rocks in Southland. The other is that in Southland the regressive strata, which record the retreat of the sea as central New Zealand began to rise again 25 myr ago, include a development of younger coal measures – a lignite field near Gore – as well as two additional limestones.

In Fig. 16.1 A you will see three clusters of cover strata: eastern, central and western. The basal coal measures are thickest and most extensive in the east around Kaitangata, which has been an active coalfield for more than a century. As at Dunedin and Shag Point, these coal measures developed early in the marine transgression (Chapter 4), around 70–80 myr ago.

In the western cluster of cover strata outcrops, the old coalfields of Ohai and Nightcaps, south of the Takitimu Mountains, are also famous. In 2009, Solid Energy announced that the Ohai mine would close as all the recoverable coal had been extracted.

Limestone cover strata are conspicuous in many places across the region, forming ridges and bluffs. There are limestones of three different ages: the countrywide one aged around 30 myr (Oligocene period); and two younger ones aged 20 myr and 15 myr (Early and Middle Miocene). Cross-bedding and scour and fill structures (Fig. 16.3), denoting shallow marine currents, can be seen in some cliffs, and in some quarries brachiopod fossils (lamp shells) are abundant. Limestones of the middle group are found around Winton, Limehills and Browns.

Well-known limestones around Clifden, in the far southwest, lie just west of Fig. 16.1 A, but are still part of the Southland story. They are the region's youngest limestones. From the historic old suspension bridge at Clifden on SH99, you can see the local sequence of strata in the riverbank and bluff around the modern bridge over the Waiau River. This locality is important in the historic development of the New Zealand geological timescale (Fig. 1.2), because it comprises the type or reference section for four of New Zealand's fossil-based stratigraphic time stages: the Altonian, Clifdenian, Lillburnian and Waiuan, covering the period approximately 20–11 myr ago, and named after local rivers, streams and settlements. The rocks here dip to the northwest (upstream), to pass beneath the much thicker strata of the Waiau Valley (Chapter 17).

The central cluster of cover strata indicated in Fig. 16.1 A is centred on Gore. As noted earlier, it includes lignite deposits laid down as the sea began to regress 25 myr ago.

Fig. 16.3. Fluted limestone of Miocene age is prominent on hillsides around Forest Hill, 5 km southeast of Winton, Southland. Cross-bedding is visible at this location on the unfluted lower portions of each limestone block. Photographer Egon Eberle.

Gravels, Alluvium and River Patterns

The course of a big river is the result of a long and interesting history. Unfortunately, that history is usually poorly recorded in the form of alluvial deposits – valleys are all about erosion, and temporary storage of sediment in terraces and floodplains. Southland is crossed by three big rivers, the Aparima, Oreti and Mataura. All of them run from north to south across the geological structure, cutting across several basement terranes and their associated ridges. This can be attributed, in part at least, to the presence of structural depressions that cross the dominant trend and have allowed the preservation of cover strata, as can be seen in Fig. 16.1 A. The other part of the explanation may be superposition – the river courses were originally established on cover strata sloping south off the rising mountains, and while these strata have now largely been removed, the river courses have been retained.

The Clutha River on the other hand, along with its tributary the Pomahaka, does not cut a gorge through the Hokonui Escarpment, despite being New Zealand's most voluminous, and therefore most energetic, river. Instead, when it emerges from the schist terrain it follows the lower ground on the Dun Mountain-Maitai and Caples terranes to Balclutha; this course may reflect the lack of a structural depression across the Murihiku Terrane at this point.

As elsewhere in the South Island, voluminous gravel production has accompanied the rise of mountains and foothills. Invercargill sits on an extensive river plain. The location of the plain is determined by a large structural depression, which has allowed both the preservation of cover strata and the accumulation of a large volume of alluvial sediment brought in by the Aparima, Oreti and Mataura rivers. Because of the proximity of the area to glaciated mountains to the north and west, loess (Chapter 15) forms a layer a metre or more thick over much of the plain. It contributes significantly to the fertile soils for which Southland is renowned.

Stewart Island/Rakiura

Stewart Island/Rakiura (Fig. 16.1 B) is composed entirely of rocks of the Median Batholith. It falls into three parts: the northern Anglem Complex, which contains a selection of the older plutonic rocks of the batholith (older than 125 myr); the central strip, which is an infaulted remnant of the volcanic superstructure associated with the batholith; and the southern half of the island, which comprises the younger granites of the batholith (130–105 myr old). The latter are known as the Separation Point suite, named from their presence at the far northern end of the South Island, on the opposite side of the Alpine Fault.

Geologically speaking, Stewart Island/Rakiura is a continuation of the eastern half of Fiordland (Chapter 17), with one interesting difference. In Fiordland, the Separation Point granites (which make up Mt Titiroa) occur to the east of the older batholiths, whereas by the time we reach Stewart Island/Rakiura they have changed sides.

Chapter 17
Fiordland and Western Otago

Fiordland is one of New Zealand's most spectacular places, for its remoteness, its magnificent rocks and its wonderful glaciated terrain. It is flanked on the east by the equally wild and remote western Otago, and the two areas are separated by the unique Moonlight Tectonic Zone (MTZ).

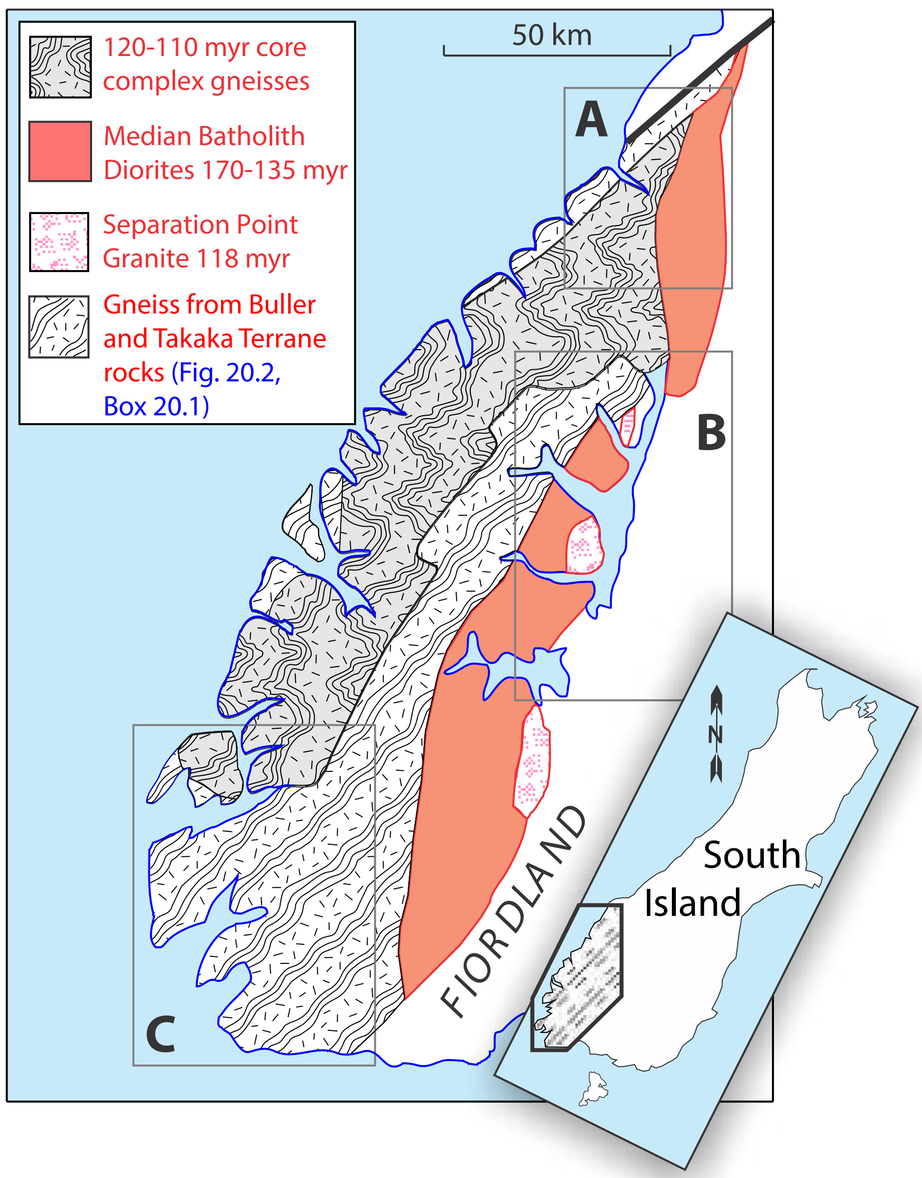


Fig. 17.1. Simplified geology map of Fiordland.

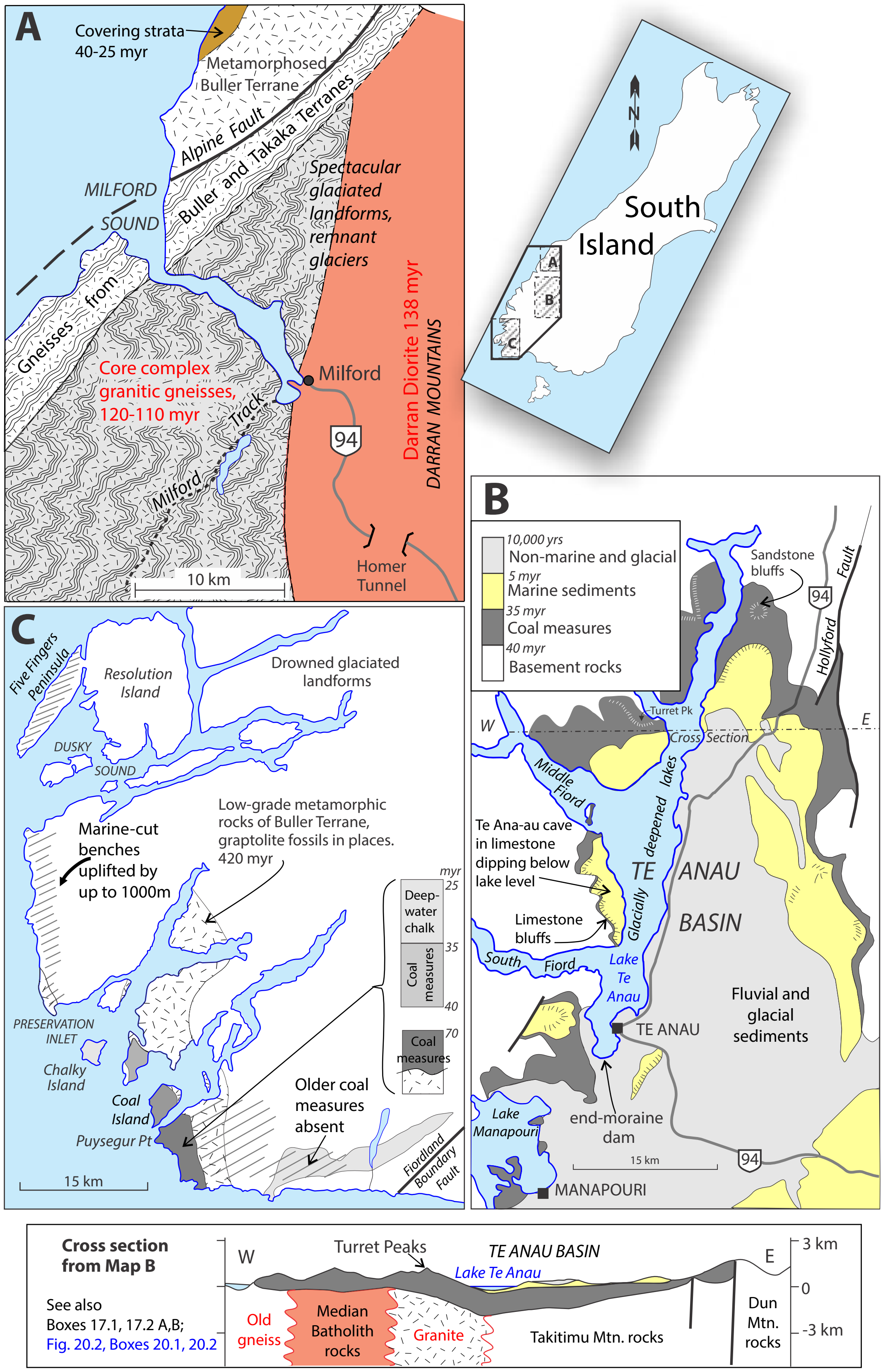


Fig. 17.2. Detailed geology maps of three areas of Fiordland (see inset map for locations).

Basement Rocks of Fiordland

The basement rocks of Fiordland are a direct continuation of those of northwest Nelson and northern Westland, but have been displaced 480 km by the Alpine Fault (Fig. 20.2). They differ in that they have been uplifted further than the Nelson/Westland rocks, and therefore we see rocks from deeper crustal levels that are much more severely metamorphosed than their equivalents in Nelson/Westland. The comparison with Nelson/Westland has helped considerably in understanding the Fiordland rocks, because indications of original provenance in strongly metamorphosed rocks are not easy to tease out.

The Fiordland rocks comprise former sedimentary and volcanic rocks of the Buller and Takaka terranes of northwest Nelson (510–400 myr old), which were intruded by the Karamea Granite 375 myr ago (Chapter 20). These rocks are now mostly metamorphosed to high-grade gneisses (Fig. 17.1 for a distribution map of these rocks, which are largely preserved in a central and southwestern strip). In southwestern Fiordland, however, the metamorphism is locally of a lower grade because the rocks were not so deeply buried. There has also been less later uplift in this region, so the higher levels of rocks have not been eroded away, as they have elsewhere in Fiordland, and we can still recognise the original rocks. Graptolite fossils typical of the Buller Terrane (Chapter 20) are preserved in some remote places here (Fig. 17.2).

The basement rocks of eastern and western Fiordland comprise deep-crustal gneisses and associated granite bodies intruded between 500 and 100 myr (Cambrian to Cretaceous). They are essentially rocks of the lower continental crust, which were uplifted around 100 myr ago during the formation of an interesting structure known as a metamorphic core complex. A closely related metamorphic core complex forms the Paparoa Range in Westland, which is both and easier to understand and more completely preserved (Chapter 20 and Fig. 20.1 A, Box 20.2). The Fiordland core complex is not accessible from a road, and also lacks the basin and range superstructure that we can see in Westland.

In a nutshell, a core complex is the extreme response to stretching of continental crust as a new spreading plate boundary is inaugurated, or when an existing spreading centre is taken underneath a continent by subduction (as happened when the Pacific spreading ridge was subducted underneath western North America not too long ago). Heat from the mantle uprising causes the crust to rise, a separation surface develops in the middle of the crust, and the upper half of the crust slides away under gravity to either side. The lower crustal metamorphic rocks then rise to fill the gap, forming a metamorphic core complex. If stretching continues, the continental crust splits and new oceanic lithosphere begins to form at a new oceanic spreading plate boundary.

The stretching event that formed our two metamorphic core complexes was the one that preceded our separation from Gondwana. It has manifestations all around the country, in the form of old rift valleys and an extensive passive continental margin, and ultimately, in the existence of the Tasman Sea. In our two core complexes, the stretching process ceased when all the spreading 'effort' became concentrated in the Tasman spreading centre (Chapter 4).

Unlike the even older Karamea Arc, whose volcanic superstructure has been completely destroyed (Chapter 20), we do have remnants of the volcanic counterparts of the Median Batholith

(Chapter 16). One of these, the Paterson Group, is found in northern Stewart Island/Rakiura. Another, oddly enough, is in Fiordland. 'Oddly' because Fiordland has been uplifted more than most other locations, to the extent that across much of its western half we see mid-crustal rocks, a fact that would lead us to expect that surface-level volcanic rocks would have long gone. However, uplift is clearly less in eastern Fiordland, where the volcanic Loch Burn Formation is found. Both it and the Paterson Group have been preserved from erosion in long, narrow, down-faulted strips, flanked by different units of the Median Batholith, and both are dated at around 150 myr.

Unfortunately, there are no roadside exposures of gneisses (strongly banded high-grade metamorphic rocks; see above) anywhere in Fiordland. They can, however, be seen from various walking tracks, including the famous Milford Track, and, at a distance, from boat trips on the various lakes and fjords. Taking a boat trip is a must at any rate, if only for the superb glaciated terrain (see below).

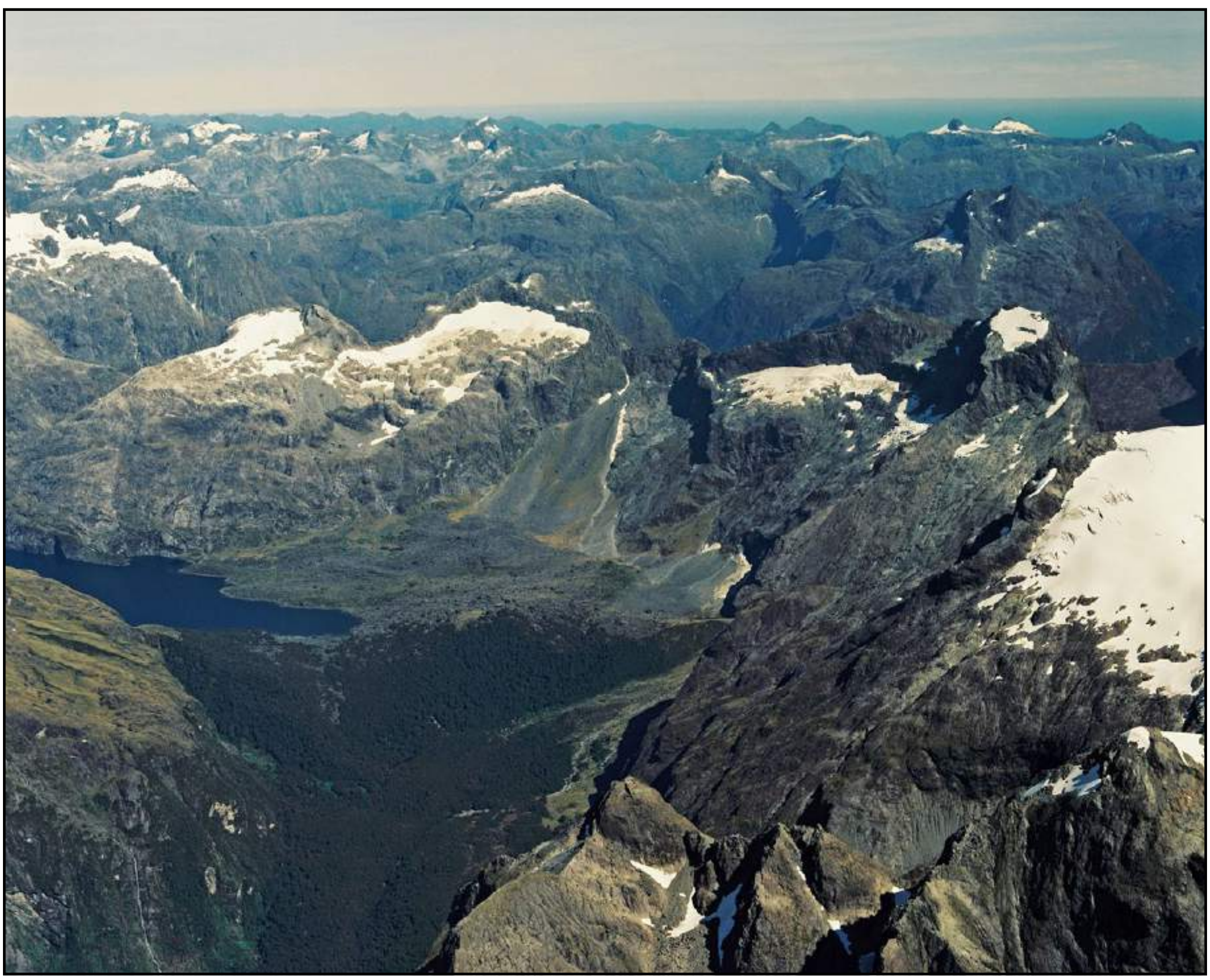


Fig. 17.3. Fiordland is characterised by its glaciated mountain landscape, like the Darran Mountains, east of Milford Sound. Lake Adelaide in the foreground.
Photographer Lloyd Homer, GNS Science.

Basement Rocks of Western Otago

This is the rugged and remote country that lies between Lake Wakatipu and the road to Milford Sound (SH94), and extends north of the road, through the Olivine Range and the Red Hills to abut the Alpine Fault south of Jackson Bay. It includes Mount Aspiring National Park, and is traversed only by walking tracks, notably the Rees–Dart, Routeburn and Hollyford tracks.

In Chapter 16, we noted that the six basement terranes that have wide outcrops in Southland swing to a northerly trend in the west of the region and narrow as they approach the Alpine Fault – or even vanish completely in places in the case of the Murihiku Terrane. The rocks have, in effect, been smeared out by the shear stress created by the two sides of the Alpine Fault pushing against each other as well as sliding sideways. In western Otago, east of the Te Anau Basin and SH94, we see the Brook Street, Murihiku and Dun Mountain-Maitai terranes at their most smeared. The geological map in Box 17.1 shows a highly simplified version of affairs, but in it the Brook Street Terrane of old volcanic rocks can be seen narrowing northwards. Rocks along SH94 in the Eglinton Valley belong to this terrane.

The most distinctive feature of the smeared zone occurs at its northern end, where, squeezed between the Olivine Range and the Alpine Fault, there are two blocks of dunite, the olivine-rich rock characteristic of the Dun Mountain Terrane (Chapter 22). The rich red-brown weathering colour of the dunite, along with a lack of bush cover, characterises the spectacular Red Hills Range (Fig. 3.2). Not only are they spectacular, but the Red Hills were a vital early clue to the scale of displacement on the Alpine Fault, because they have an exact matching counterpart on the opposite side of the fault in Nelson, 480 km distant.

Most of the remote country lying east of the narrow belt of Dun Mountain-Maitai, Murihiku and Brook Street terranes, between it and Lake Wakatipu, is made of schist and greywacke belonging to the Caples Terrane. On the western side of the belt, the iconic Darran Mountains (traversed by SH94) are made of diorite and are part of the Median Batholith. The Milford Tunnel cuts through the Darran Diorite, and the rock is well exposed around the eastern tunnel entrance; individual crystals are just big enough to see with the naked eye. Milford Sound settlement is also located on Darran Diorite, which is dated at 168–131 myr.

Cover Strata

Like most of the rest of New Zealand, Fiordland was submerged during the long, quiet interval between 100 myr and 25 myr ago, coinciding with phases two and three of the country's development. The same succession of non-marine coal measure strata followed by sandstones and marine limestones occurs here as elsewhere (Chapter 4), but there are three interesting differences. First, in the southwest there are two groups of coal-measure strata, around 70 myr and 40–35 myr old. This is the same as at Greymouth and the Taranaki Basin, reflecting the fact that, before the plate tectonic movements of the last 45 myr, southwest Fiordland was adjacent to Westland (Box 17.1), and, as noted above, was in the zone of rifting 100–80 myr ago.

Second, the limestones in the southwest are deep-water chalks, similar to the Amuri Limestone of Marlborough (Chapter 15), which implies that there was a deep marine basin adjacent to the southwest corner of Fiordland.

Fig. 17.4. Sandstone and mudstone turbidite cover beds from Coal Island, southwest Fiordland, were deposited in a deepwater rift lake some 100 million years ago (Cretaceous). Photographer Jon Lindqvist.

Third, in some places adjacent to Lakes Manapouri and Te Anau, these rocks are much thicker than usual. This fact links to the large area of cover strata located around Lake Te Anau, which are there because of the sedimentary basins of the Moonlight Tectonic Zone (MTZ, see below), active 45–25 myr ago. Outside the MTZ, as in the country between South and Middle Fiords of Lake Te Anau (the Murchison Mountains), the cover strata are thin, as usual, and contain the countrywide 30 myr limestone bed. This limestone bed dips eastward off the mountains, to pass under the lake, and hosts the famous Te Ana-au Glowworm Caves (Fig. 17.2).

The Moonlight Tectonic Zone

A tectonic zone is a tract of country, usually elongate in shape, that is characterised by a particular style of tectonic deformation. The MTZ (Box 17.1) lies between Southland and Fiordland, and at the present time comprises a line of three structural and sedimentary basins. From south to north, these are the Solander (offshore, between Stewart/Rakiura and Solander islands), Waiau and Te Anau basins. The large coastal indentation of Te Waewae Bay reflects the influence of the Waiau Basin. The MTZ also includes the Moonlight Fault, which crosses Lake Wakatipu at Bobs Cove, where its fault angle contains a steeply dipping sliver of young cover strata, the sole remnant of the cover strata that once covered all of the Wakatipu region.

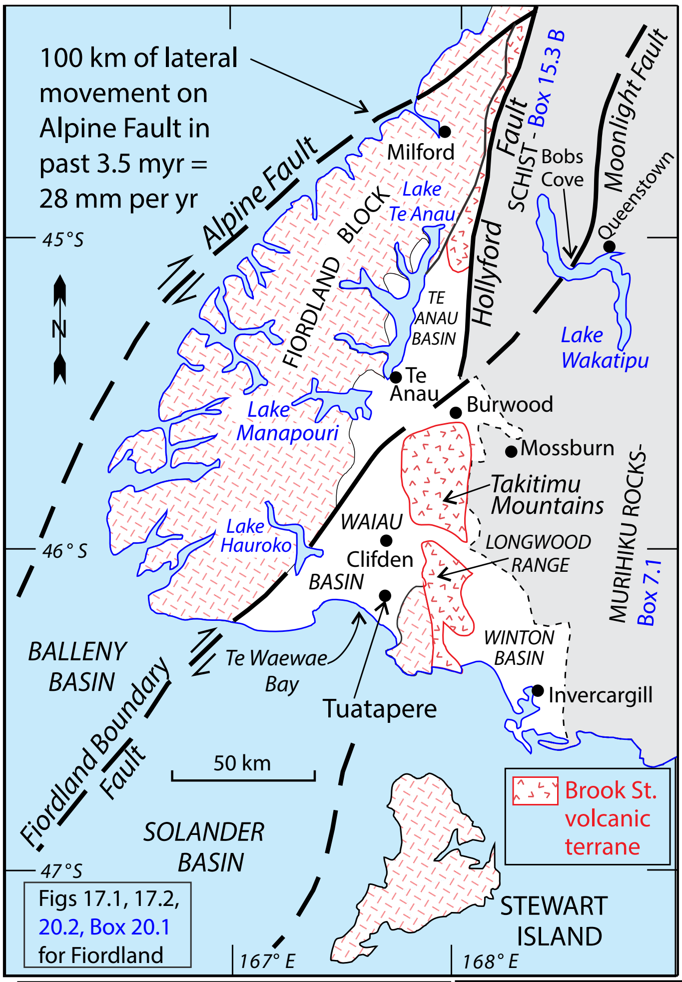
The MTZ is sandwiched between the tectonic blocks of Fiordland and Southland/western Otago, and is cut off by the Alpine Fault in the north. Its story differs from that of the rest of the country, telling of a unique interval lasting 20 myr (45–25 myr ago), during which this part of New Zealand was the site of a divergent or spreading plate boundary between the newly established Pacific and Australian plates. In Chapter 6 we alluded to the same plate boundary in the north of the country, where it was inferred to be a very slow-moving subduction zone. Boxes 5.5 A–D tell the full history, and in particular explains how the spreading plate boundary propagated northwards into continental crust, from oceanic crust to the south, and changed further to the north into the slowly subducting plate boundary. This was all on account of the location of the pole of rotation around which the Pacific Plate was rotating in an anticlockwise direction with respect to the Australian Plate.

At that time, the pole of rotation was located in central New Zealand. The consequence of this was that the anticlockwise rotation of the Pacific Plate moved it and the Australian Plate together to the north of the pole, and apart to the south of the pole. Within New Zealand, the movement was slow in both cases, because of nearness to the pole. The pole subsequently moved progressively to the southeast (Boxes 5.5 A–D), with the result that the nature of the Pacific–Australian plate boundary changed, dramatically, around 25 myr ago.

Between 45 myr and 25 myr ago, while most of New Zealand slept and subsided gently beneath the sea, the MTZ was the active spreading plate boundary within New Zealand continental crust. This plate divergence opened up a line of deep, localised marine basins, which then did what such basins normally do – they attracted sediment. This commonly arrived by way of sediment gravity flows to form flysch deposits comprising turbidites (Boxes 6.5 A, B).

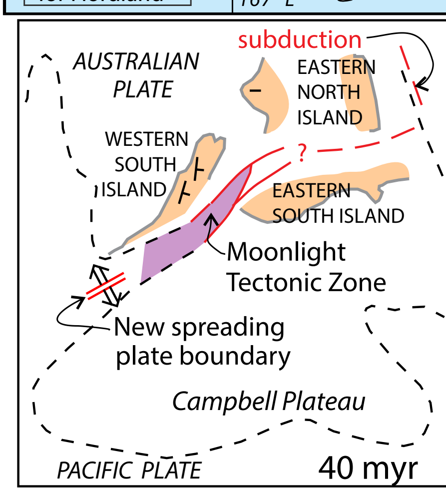
Fig. 17.5. Oblique view northwards up the axis of the Moonlight Tectonic Zone from Te Waewae Bay on the south coast through the Waiau and Te Anau sedimentary basins. The zone divides the tectonic blocks of Fiordland (left) from Southland-west Otago (right). Photo courtesy of Google Earth.

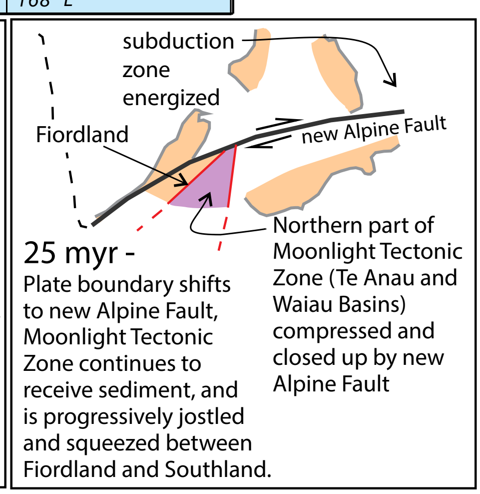
Box 17.1. Moonlight Tectonic Zone (MTZ) – Te Waewae Bay to Te Anau.



Main facts —

1. The Moonlight Tectonic Zone (MTZ) (Te Waewae Bay-Te Anau-Bobs Cove -Moonlight Fault) is our best record of the first 20 myr of the Pacific-Australian plate boundary.
2. From 45-25 myr it was a spreading boundary, opening and forming sedimentary basins.
3. At 25 myr the plate boundary changed location to the Alpine Fault, and changed nature to a combination of compression and sideways movement. Box 5.5 B explains why.
4. From 25 myr, northern MTZ has been progressively squeezed and closed, while mountains have risen and jostled for position around it.
5. Southern MTZ has received much sediment from both sides since 25 myr.
6. Refer back to Chapters 3 and 4.





Although this process continued for 20 myr, the spreading was slow. It did not open a wide oceanic basin within the continent, or develop a metamorphic core complex as happened in Fiordland 100 myr ago (see above). The reason for this is twofold. First, as noted, the area lay close to the pole of rotation, so that the quantum of opening was small. Second, the southeastwards migration of the pole of rotation soon converted the pure spreading motion into a more complicated combination of spreading and transform movement (Boxes 5.5 A–D). Geologists call this a transtensional opening.

Further south, however, further away from the pole of rotation, spreading was faster. The result was the formation of the Emerald Basin of new oceanic crust that now lies to the southwest of Stewart Island/Rakiura.

The dramatic energising of the plate boundary 25 myr ago affected the whole country. In particular, the Alpine Fault began to form in western South Island, but it did not follow the MTZ. Instead, it cut across the northern part of it (Box 17.1). The effect was to reverse the opening process in the northern part of the MTZ, although the sedimentary basins were not immediately destroyed. As relief increased, the dominant type of sediment changed to sand and gravel, and the basins were slowly squeezed.

As with the rest of the South Island, the main phase of compression and uplift of mountains here did not begin until 5 myr ago. As a result, there is a good record of sedimentation in the MTZ (mostly in basins that were still connected to the sea) for the period 25–5 myr ago. From 5 myr onwards, however, the basins have been progressively squeezed shut, starting from the north; since then, gravel-dominated sedimentation from rising mountains has been the story. Te Anau and Waiau basins are now folded and are receiving glacial, river and lake sediments. Solander Basin is still open and still marine, with Te Waewae Bay indicating its presence at the coast. Everything north of Te Anau Basin has been squeezed shut, uplifted and eroded away, except for remnant fault-angle slivers on the Moonlight Fault at Bobs Cove on Lake Wakatipu, and on the Hollyford Fault north of SH94.

As noted above, in places along the eastern margin of the Fiordland block the 45–25 myr old sediments display the usual character of the rest of the country (i.e. they are thin and are dominated by limestone), indicating that the western edge of the MTZ coincided closely with the present-day Fiordland Boundary Fault.

The Waiau River

The Waiau is a fine alpine river with a catchment that covers a large area of southwestern New Zealand. It drains Lakes Manapouri and Te Anau, is joined by the Mararoa River (which itself drains western Otago), and flows south along the southern MTZ to reach the sea at Te Waewae Bay (Box 17.1). Its flow was reduced when the Manapouri Power Scheme diverted water from the lake to Doubtful Sound. The gravel banks along the river bed are always worth a fossick because of the great variety of handsome rocks. These come mostly from the Fiordland portion of the Median Batholith and are 'granites' of intermediate composition (silica content 60–70%). They are generally a composition in black and white, the white crystals being plagioclase feldspar and the black ones either amphibole (e.g. hornblende) or pyroxene (e.g. augite).

Banding (signifying the metamorphic rock gneiss; Box 15.3 B) may be present in other stones. These rocks come from further into Fiordland, and are derived from the old Buller and Takaka terrane rocks (Fig. 17.1). Veins of white quartz and feldspar are common in all these pebbles. Granite cobbles from Mt Titiroa are lighter in colour, containing white feldspar, glassy quartz and small black glistening micas; these are genuine high-silica granite. Other rocks in the river gravels come from the schist mountains to the northeast, and the old volcanic Takitimu Range to the east.

In fact, one of the factors that has helped geologists to unravel the highly complicated history of the MTZ has been the wide range of distinctive rock types found in the old gravels along the Waiau, rocks that leave little doubt about their provenance. Elsewhere in New Zealand, the rocks found tend to be greywacke, greywacke or greywacke, making it generally impossible to distinguish one source from another, but here there is an embarrassment of riches, rock-wise.

The Uplift of Fiordland

The fact that cover strata are preserved only around the edges of Fiordland is because of the tectonic uplift that is presently occurring here. This uplift is evident in the remains of extensive marine-cut terrace surfaces around the southwestern corner at heights up to 1000 m. Recent work at the southwest corner of Fiordland has dated one such terrace, 65 m above sea-level, at around 110,000 years, which is the time of the last interglacial high sea-level before the present-day interglacial high. This indicates an average uplift rate here of 0.5 mm/year.

The method used to date this terrace is interesting. It makes use of the fact that energetic cosmic rays from space interact with rock exposed at the earth's surface to form cosmogenic nuclides. The rays penetrate rock, and can form nuclides at up to 100 m depth, although the production rate varies with latitude and altitude. The nuclides themselves are radioactive isotopes of various common elements, such as helium-3, beryllium-10, carbon-14 (that's the one used in radiocarbon dating), neon-21, aluminium-26 and chlorine-36. Like all radioactive elements, these nuclides decay with time, each one at a different rate, a fact that can be made use of in some other dating scenarios. In this case, however, the time that the terrace boulders have been exposed to cosmic rays is much less than the decay times of beryllium-10 and aluminium-26, so the dating technique simply measures the quantity of the two nuclides in the surface layer of the boulders and uses that as a clock, given that the rate of their production is known. Having two independent clocks means that if they give the same age we can be confident that the 'apparent exposure age', as it is called, is probably correct.

The uplift in southwest Fiordland is much more recent than that associated with the metamorphic core complex, which took place around 100 myr ago. As a consequence of these successive uplifts, Fiordland has a strong positive gravity anomaly (a stronger-than-average gravitational pull) owing to the presence of heavy lower-crust rocks at the surface. When walking the Milford Track you and your pack are fractionally heavier than you normally are!

Fig. 17.6. Much of southwest Fiordland is currently being pushed upwards as evidenced by the presence of marine-cut terrace surfaces well above above sea level. In this photograph there are a series of uplifted terraces between Gold Burn and Preservation Inlet, up to 400 m high. Photographer Lloyd Homer, GNS Science.

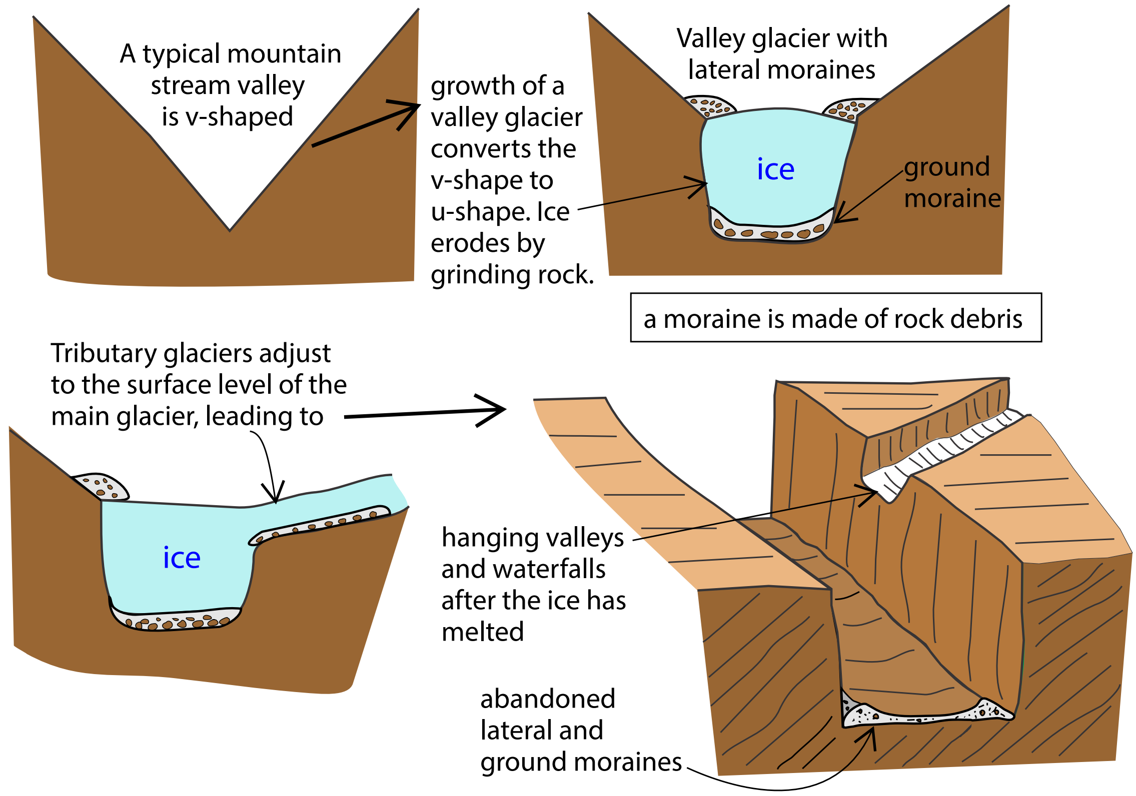
Glacial History and Landforms

The combination of height, hard homogeneous rocks and extreme rainfall (10 m-plus/year) in Fiordland gives rise to spectacular steep landforms, through the intermediary of glaciation. There have been many glaciations in the past few million years, the most recent climaxing 20,000 years ago and retreating between 20,000 and 10,000 years ago. Very high snowfall in Fiordland gave rise to thick, fast-moving glaciers, which spread radially outwards onto the continental shelf (global sea-level was at a glacial low of −130 m). In the Te Anau Basin, these glaciers joined with others originating in mountains to the north and east, and together they moved south.

The results today, after the retreat of the glaciers (there are still some tiny residual glaciers in places), are well-preserved glacial erosional landforms throughout Fiordland. These include cirques and, of course, the fjords, where hanging valleys produce spectacular waterfalls, as well as the glacially excavated lakes of Manapouri and Te Anau (Boxes 17.2 A, B). They also include glacial accumulation landforms such as moraines (Box 19.4) throughout the Manapouri–Te Anau basin. Lake Te Anau itself is held in by the end moraine left behind by the final phase of the glacier that carved out the lake bed as it was retreating.

Box 17.2 A. Fiords and U-shaped valleys and cirques.

A notable landmark in the history of Science was the mid-nineteenth century recognition, by Alexander Agassiz, that the mountains of Scotland had been glaciated, even though there are no glaciers there now. One of his lines of evidence was U-shaped valleys (see Box 19.4 for others).



Formerly glaciated terrains in NZ — include Mt Ruapehu and the higher North Island mountains south of Mt Ruapehu (glaciers survive on Mt. Ruapehu); all the South Island mountains, where glaciers survive in the Mt Aspiring and Mt Cook regions (Box 19.4); Fiordland and Stewart Island. U-shaped valleys are best seen in Fiordland — e.g. the Milford Sound road SH 94. They are less clear in the Southern Alps, because of the highly fractured nature of the greywacke rock there, and the rapid erosion — there has been 100 metres of uplift since the end of the last glaciation 10,000 years ago. However, glacial moraine landforms are superbly preserved in the Alps and on the West Coast following the recent retreat of the glaciers.

Fiords are U-shaped valleys now occupied by the sea. How can this be?

1. During glaciation (the last one climaxed 20,000 years ago) world sea level falls by about 130 metres (because of the large quantity of water locked up in ice sheets).
2. Glaciers can erode to below sea-level, because ice does not float until it is nine-tenths submerged — more than nine- tenths if it is heavily loaded with rock.

Fiords commonly have a shallow sill at their entrance. The sill may be a terminal moraine, or made of bedrock preserved from erosion under the thinning terminus of the glacier.

Box 17.2 B. Cirque landforms.

Cirque landforms

Cirques are common landforms in Fiordland. In the Southern Alps the cirques may still contain ice, or they may be ice-free.

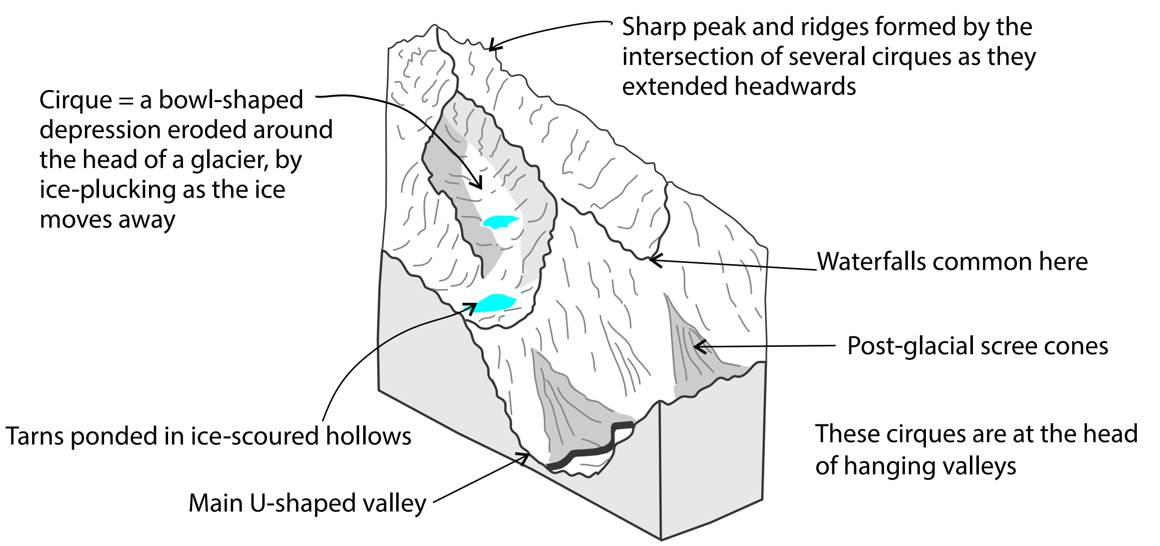
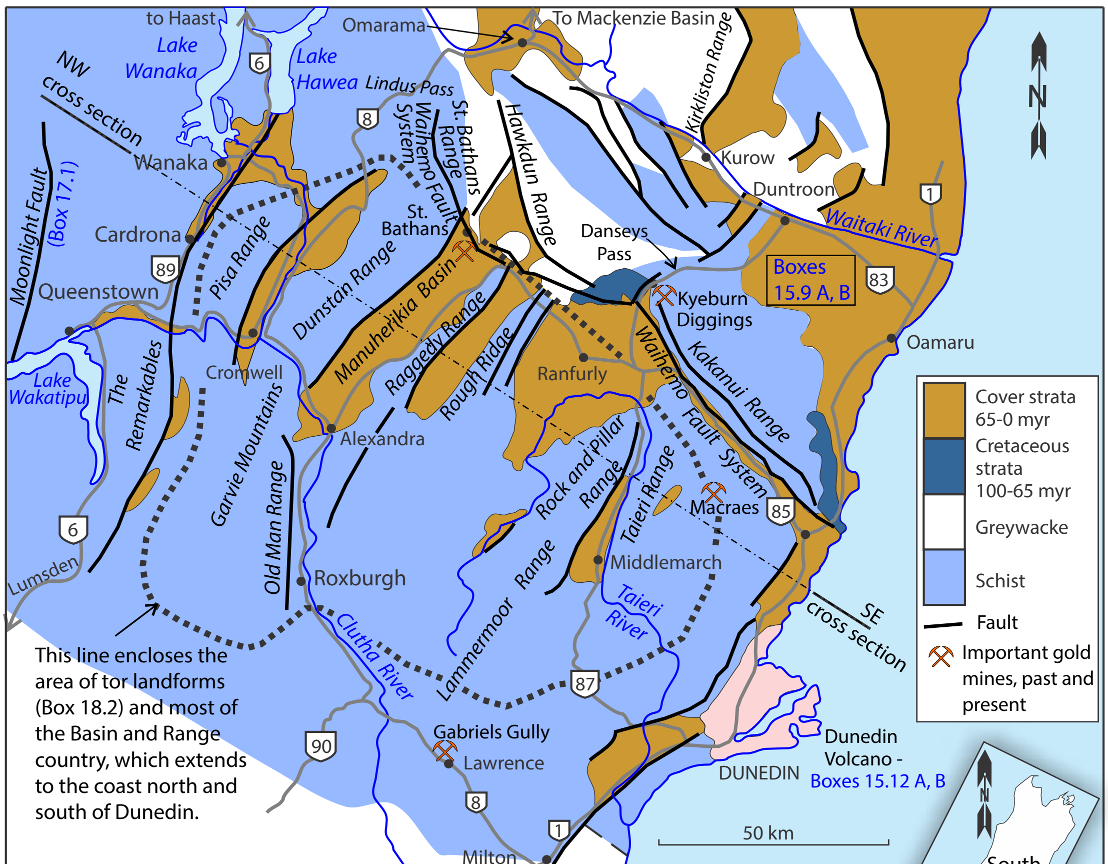


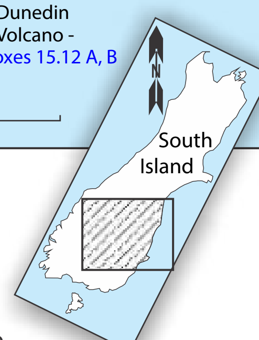
Fig. 17.7. A spectacular U-shaped, glacier-carved hanging valley with waterfall discharging into the larger glacier-carved valley of Doubtful Sound, Fiordland. Photographer Bruce Hayward.

Chapter 18

Central and North Otago and South Canterbury

Box 18.1. Central Otago – Basin and Range Country.

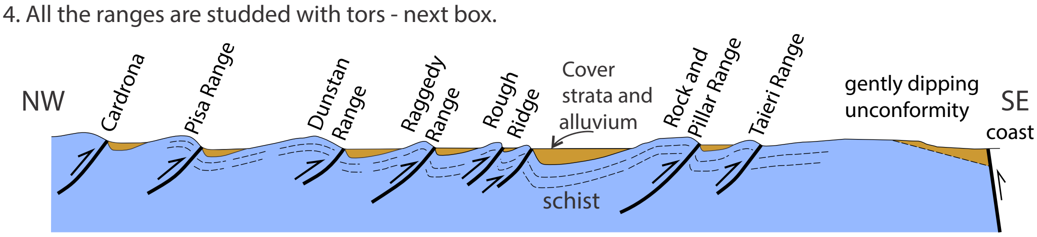




Diagrammatic cross-section, not to scale, across the Basin and Range Province.

Note-

1. The basins are mostly one-sided fault-angle or half-graben structures, controlled by near-vertical reverse faults (Box 6.6 C for fault types).
2. The ranges are upfaulted blocks of schist, which in some cases are folded broadly into anticlinal (arch) shapes as shown by the disposition of the Otago peneplain.
3. Cover strata which once covered the whole area are now preserved only in the basins, where they are largely concealed by recent alluvium. They include significant lignite reserves.
4. All the ranges are studded with tors - next box.



In Chapter 15, we visited the basin and range province of North Canterbury, centred on Culverden. Here, in Central and North Otago and South Canterbury, there is a much bigger basin and range province. As with the North Canterbury one, the southern province is actively forming under the influence of the same tectonic driver – the southwesterly motion of the Pacific Plate, pressing obliquely against the Australian Plate at the Alpine Fault. This is generating an overall regime of compression from the northeast, pushing against a backstop to the southwest. As with the whole of the South Island, most of the action has taken place over the past 5 myr, since compression began at the Alpine Fault plate boundary. The province falls into two contiguous but contrasting parts: Central Otago, with a basin and range landscape oriented northeast–southwest; and North Otago and South Canterbury, also a basin and range province, but with ranges trending northwest–southeast, at right angles to those of Central Otago.

Central Otago

Central Otago is a unique and highly distinctive region of New Zealand. The dry, continental climate results from its location in the rain shadow of the widest part of the Southern Alps, and its great distance from the east and south coasts. The broad, northeast-trending ranges and intervening basins here form a distinctive scenic and geological landscape. The schist rock that underlies the whole area imparts its own ambience, stemming largely from the fact that the planes of schistosity (Boxes 15.3 A, B) dip at lower angles here than elsewhere, which is reflected in the landforms. The area retains large tracts of the Otago peneplain, most of which is studded with hundreds of rock tors. Finally, as a spin-off from that combination of factors, the area was rich in alluvial gold, and still has a bulk-rock schist gold mine at Macraes Flat.

The Central Otago Basin and Range Province

Basin and range is a term used to describe a landscape composed of parallel ridges and basins that are fault-controlled. As shown in the cross section in Box 18.1, the northeast-trending ranges of Central Otago are uplifted on their southeastern sides along steeply dipping reverse faults. The ranges all have the same asymmetry, with range tops closer to the southeastern fault margin, and northwestern slopes forming a dip-slope of the Otago peneplain that passes beneath cover strata. Older cover strata and more recent alluvium are preserved in the fault-angle depressions (the basins).

Reverse faults imply an overall shortening and thickening across the area (Box 6.6 C). This is exactly what is going on at present as part of the general compression across the South Island, where Canterbury and Otago crust (Pacific Plate) is being pushed against West Coast crust (Australian Plate) across the Alpine Fault, causing uplift of the Southern Alps (Box 19.1 A). However, Central Otago is sufficiently distant from the Alpine Fault that uplift has not destroyed the region's peneplain surface (see below), as it has further west – hence the broad, planar character of the ranges.

From the northeast trend of the faults and ranges, one might assume that compression would be directed at right angles to the faults, along a northwest–southeast line. In fact, this isn't the case. Distortion of survey triangulation networks between the 1860s and late twentieth century shows that compression is actually being directed along a NE–SW to ENE–WSW band – that is, nearly parallel to the faults (and to the Alpine Fault as well). This alignment of tectonic stress is actually what would be expected from the southwesterly motion of the Pacific Plate, and in theory it should cause some sideways movement on these faults, although strangely there is no sign of this.

The range structures are large in scale and geologically young – less than 5 myr (the same age range as the Southern Alps). The schist rocks making up the ranges are themselves much older, and contain small-scale structures dating from the formation of these rocks around 180 myr ago (Boxes 15.3 A, B). The main structure within schist is schistose layering, which shows up clearly in the landscape wherever schist forms the surface rock. In most places within the outcrop of the Otago Schist, the layering is inclined at 40° or more to the land surface, but it so happens that in the basin and range area of central Otago it is nearly parallel to the surface (and hence to the peneplain; see below). This arrangement leads to a distinctive array of landforms, showing up particularly in the numerous tors (see next page).

Fig. 18.1. Looking north east from the Old Woman Range through some of the basin and range topography of central Otago. In the centre is Lake Dunstan in the upper Clutha basin and on either side is the Pisa Range (west, left) and the Dunstan Range (east, right). Photographer Nick Mortimer.

The Central Otago peneplain

This is part of the virtually planar surface (the 'pene-' prefix derives from the Latin *paene*, meaning 'almost') that was eroded across most of the older rocks of New Zealand during the long tectonic quiet period of phase three of the country's formation 100–25 myr ago (Chapter 4, especially Box 4.1). The erosion surface was eventually submerged and covered by marine strata in most places. Since the present tectonically noisy period began, 25 myr ago, New Zealand has been folded, faulted and uplifted to varying extents. The erosion surface is still buried in some places, has been uplifted and totally destroyed by erosion in others, and remains as a distinctive part of the landscape, after the cover strata have been removed by erosion, in yet other places. Central Otago has by far the largest such remnants of the peneplain, and they form the present-day surface of much of the schist country (Fig. 18.2).

You can recognise the peneplain surface by mentally filling in the small stream valleys that have cut into it, and linking the planar surfaces that are preserved on the intervening minor ridges. The word planar is used, rather than flat or level, because the surface is rarely horizontal – it has normally been tilted to angles of a few degrees from horizontal. Larger rivers, like the Taieri, have cut deep valleys into the peneplain (both the gorge and peneplain are seen spectacularly on the Taieri Gorge Railway excursion from Dunedin to Middlemarch). Most commonly, you will recognise the peneplain sloping toward the basins from the southeast side, and diving underneath the young sediments in the basin.

Note also that the peneplain is arched up over some of the ranges. This indicates that the rocks have been folded into broad anticlines in addition to being raised along reverse faults.

Fig. 18.2. The Central Otago Peneplain has been uplifted, folded and in many places stripped of its soft overlying cover beds to exhume its former flat surface in many parts of Otago. Here it is seen forming the surface of the Dunstan Mountains viewed from near Luggate to the north. Photographer Bruce Hayward.

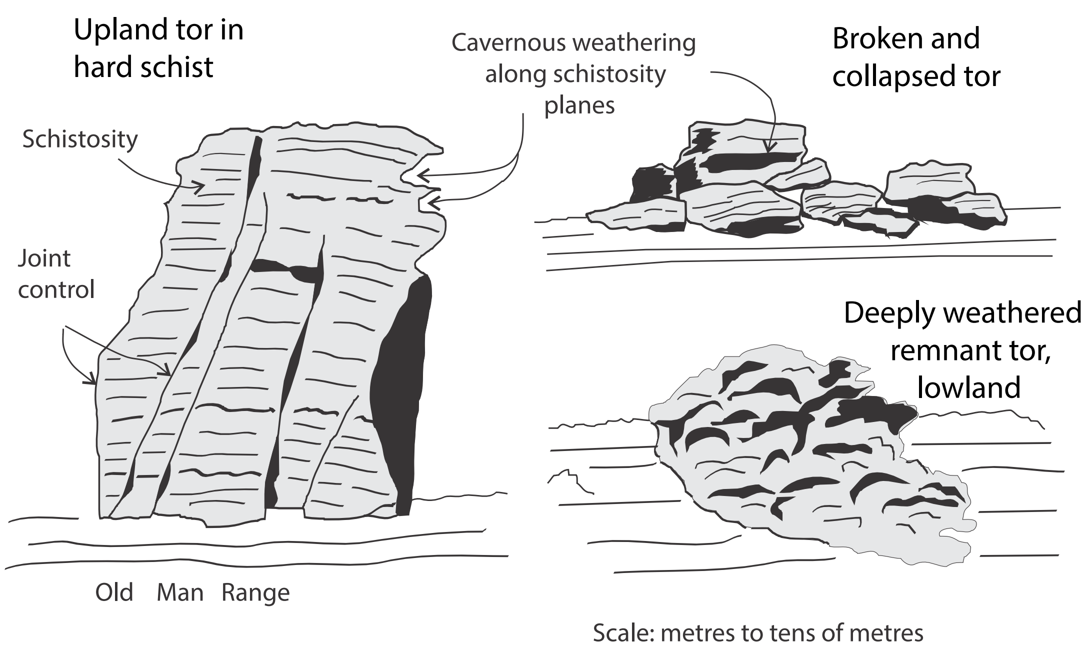
Tors

There is debate among geologists about how closely the peneplain surface we see now approximates to the surface that formed 100–25 myr ago. Undoubtedly, there has been a degree of degradation and downwasting of the surface since it was exposed. Some workers interpret the tors, or rock pillars (Box 18.2 and Fig. 18.3), found on the peneplain as being remnants of unweathered schist surviving from weathering below the level of the peneplain, and subsequently exposed by recent removal of the weathered material. This would make the peneplain that we see today a modified version of the original.

The precise mode of formation of the tors, and why they are present only in a clearly defined area, is not understood. They appear to be old – perhaps more than 1 myr. There is a relationship between the degree of weathering of tors and the parent schist – upland tors do not occur where the schist is deeply weathered – and there may be a relationship with localised silification (hardening) of schist. Similar landforms occur in granite country, e.g. on upland surfaces of Stewart Island/Rakiura.

Box 18.2. Tors.

A tor is a rock pillar, or a stack of loose blocks of rock, that stands above the surrounding terrain. Tors occur in many places (e.g. granite tors on Dartmoor, England) but are particularly abundant throughout a large part (though not all) of the schist country of Central Otago (Box 18.1).



Three examples of schist tors in Central Otago, rising above the Otago Peneplain. Note the dual control by widely spaced, steep joint surfaces, and closely spaced schistosity planes which are inclined at different angles and directions in different places.

Fig. 18.3. 3–5 m tors made of schist, in the Dunstan Range, central Otago. Photographer Nick Mortimer.

Effects of climate and drainage

One consequence of the dry climate in Central Otago is that streams are more widely spaced than in most of New Zealand, which in turn has aided the preservation of the peneplain. It has also made it possible to use stream patterns to analyse the way in which the ranges are still growing. For example, Rough Ridge west of Ranfurly is in two parts: North and South. South Rough Ridge has disrupted and diverted streams that once flowed smoothly eastwards from North Rough Ridge, indicating that North Rough Ridge is older and now inactive, while South Rough Ridge is younger and is still growing and extending northwards.

As in other parts of New Zealand, the major rivers (in this case the Clutha and Taieri) pre-date the modern topography and have incised their valleys into the growing ranges, forming gorges. Locally they flow along the basins. The Clutha, New Zealand's most voluminous river, gets much of its water from Lakes Wakatipu, Wanaka and Hawea, and from the wetter ranges of western Otago. However, in order to gather up a decent quantity of water in drier Central Otago, the Taieri has to make a great loop through the eastern half of the basin and range country.

Glaciation

Central Otago was not covered by ice sheets during the glacial periods (there wasn't enough snowfall), but it was strongly affected by very cold temperatures and waterlogged ground. During several glaciations (the last one climaxing 20,000 years ago), frost-sorting of stones into circles and stripes by repeated freezing and thawing took place on the ranges. On the slopes, solifluction occurred, which is glacier-like creep of waterlogged and semi-frozen rock debris into lobe-shaped masses or lumpy slope-parallel ledges. The latter can be seen from the Danseys Pass road, and the former in the Kawerau Gorge and on the Old Man and Dunstan ranges. In the gorges of Central Otago, solifluction deposits are closely associated with landslides (see below).

Landslides

Surveys of potential hydroelectricity dam sites in the Clutha and Kawerau gorges in the 1980s revealed a much greater prevalence of landslipping in the schist terrain than had been previously recognised. In particular, investigations accompanying the design and building of the Clyde Dam, on the Clutha River between Alexandra and Cromwell, revealed a disturbing number of deep-seated landslides on the slopes above the dam and Lake Dunstan. Extensive and expensive remedial works were necessary to ensure the safety of the construction. Other potential dam sites were ruled out by related geotechnical problems, including previously unrecognised active fault traces.

Old landslides can be recognised by their generally lower relief and finer, lumpy surface texture, compared with the surrounding schist terrain. They range up to 1 km or more in width, though most are tens to hundreds of metres wide, and of course they occur on slopes and tend to funnel downslope into stream valleys (Boxes 12.5 A–C).

Otago gold

Gold-mining began in Otago with Gabriel Read's 1861 discovery near Lawrence (Gabriels Gully). Subsequently, hundreds of gold discoveries were made, throughout Central Otago and the basin and range country. The great majority of these were alluvial gold – tiny grains of gold and rare nuggets in alluvial gravels – obtained by panning and sluicing. The origin of nuggets is described below.

The story of Otago gold has many points of crossover with the geological development of the region. The ultimate source of gold is quartz veins in schist. The veins form from circulation of hot water through cracks, late in the process of schist metamorphism, and contain concentrations of rare elements sweated out of the schist (tungsten in the ore scheelite is one that has been mined in western Otago, e.g. around Glenorchy). Few veins are big enough or contain enough gold to sustain an underground quartz-mining operation, but at Macraes Flat at present a large and successful open-cast mine and deep underground mine that extends below sea level is extracting gold from a bulk-rock low-grade ore comprising a mixture of schist and quartz veins, and was extending activities into an underground operation.

Gold released by erosion of schist has accumulated in alluvial sediments. Gold is very heavy, and is found in small grains (less than 1 mm) that get trapped between larger rocks and sand grains. It is typically concentrated at or near the base of alluvial gravels, forming what the gold-miners called 'leads'. The accumulation of gold in the Central Otago alluvial gravels occurred during five different geological periods:

1. Around 100 myr ago (Cretaceous period), when the schist was first exposed. At this stage New Zealand was mountainous and wet, and the long tectonic quiet period (Chapter 4) was just beginning.
2. During the initial phases of the subsidence and submergence of New Zealand when coal measures were forming, around 70–35 myr ago (Chapter 4).
3. In the early phases of tectonic revitalisation beginning 25 myr ago, when the first new land areas emerged (Chapter 3).
4. Around 5 myr ago, when the main phase of mountain-building began throughout the South Island (Chapter 3).
5. During glaciations and deglaciations.

Older gravels tend to be composed mainly of the more resistant quartz pebbles, while the youngest ones consist largely of schist and greywacke rock fragments. Each successive round of production of alluvial sediment has tapped both raw schist and older alluvial sediments. Gold can thus be recycled from an older alluvial deposit to a younger one, while gold-miners had the potential to mine alluvial gold from deposits of several different ages in one location. For example, around Lawrence much gold was obtained from gravel deposits of the oldest of the five episodes, and at Gabriels Gully the large blocks of gold-bearing conglomerate that you see when you visit are of the same age (these blocks were too strongly cemented for the miners to break down). Gold was also obtained from recent alluvium deposits at Gabriels Gully.

Gold nuggets

Surprisingly, gold is not insoluble in water. Gold nuggets do not come from quartz veins, but grow in the alluvial sediment through precipitation from water onto existing grains of gold – or even onto gold coins. As Central Otago has moved through this succession of five periods of gold accumulation, the climate has become progressively more arid. A dry climate equals less water, which in turn equals a greater concentration of dissolved chemicals – principally lime from the increasingly alkaline soils. This has led to more solution and precipitation of gold as time has gone by, and as a result gold grains and nuggets tend to be bigger in younger deposits.

Fossils from cover strata in the basins

Because of the dominance of broad schist ranges, we do not normally associate Central Otago with fossils. However, some of the most interesting fossils in New Zealand have been found here in recent years, and their origin stems in part from the on-again, off-again nature of old marine inundation. There is a lot in this book about the transgressive, or onlapping, sequence of strata that resulted from the gradual subsidence of Zealandia 100–25 myr ago, beginning with swamp coal measures and culminating in the 30 myr-old limestone that is found in many places (Chapter 4). When the subsidence motor went into reverse 25 myr ago and the land started to rise again, the sea, of necessity, had to retreat or regress, leading in some places to a regressive, or offlapping, sequence of strata that overlies the older transgressive sequence. This sequence follows a reverse order, starting with marine strata and moving to non-marine strata where circumstances permitted. The onlap-offlap cycle is shown diagrammatically in Box 15.9 A.

In Central Otago few transgressive strata are exposed and the extent of marine inundation is uncertain. By 19 myr ago a very large lake, or complex of lakes, existed here. This persisted for at least 3 myr, and is known as Lake Manuherikia, after the Manuherikia River. The sediments accumulating in this lake ranged from gravels to muds, and included swamp peats that are now low-grade coals and lignites (low grade because they have never been buried deeply). These are the Otago-Southland Lignites that are often touted as a future energy source. As a consequence of the basin and range faulting and erosion of the last 5 myr, lake deposits are now preserved in widely scattered outcrops.

More interestingly, there are unique vertebrate fossils in the lake beds. In addition to the fish bones and snail shells that we would expect to find, there are bones and eggshell fragments of at least 25 species of bird, as well as bones of lizards and a variety of bats. Of particular interest are freshwater crocodiles and the earliest record of tuatara fossils in New Zealand. There are also plentiful plant fossils. These fossil-rich beds are under intense investigation at the time of writing – who knows what further surprises await us.

All this serves to illustrate the patchy and capricious nature of the fossil record, especially of non-marine and terrestrial organisms. A few years ago, no one would have predicted finding crocodiles or bats in our fossil record. Part of the problem is that bones and shells dissolve in acidic water, and the majority of our non-marine sediments occur in coal-measure sequences that formed in acidic swamp environments. Lake Manuherikia was unusual in that, in the dry and warm climate of the time, its waters contained plenty of lime and hence were alkaline.

Signs of mountain-building activity

Mountains – even quite low ones – shed gravel, which is gathered up by adjacent basins. Gravel and other deposits laid down early on in the current mountain-building cycle (i.e. in the last 5 myr) have themselves been subjected to ongoing tectonic deformation and weathering. There are extensive deposits of these older conglomerates in the Otago basins, in which the original bedding is tilted at varying angles. They are collectively called the Maniototo Conglomerate, and an example can be seen from SH85 at Kyeburn, on the bank of the Kye Burn.

Another well-known example, promoted locally as a tourist attraction, occurs at Omarama, on the southern fringe of the Mackenzie Basin. The colourful Clay Cliffs occur just to the north of the town, and form a striking badlands topography. These older, mountain-derived gravels, sands and muds, now weathered to reds, browns and yellows, dip at an angle of 35°.

Volcanoes

While the emphasis in this chapter is on tectonic action, there has also been notable volcanic activity in the not-too-distant geological past. In Chapter 15, when discussing the cover strata seen around Canterbury and North Otago, we noted that they contain a peppering of basaltic volcanic rocks. These basalts are quite independent of the big Banks Peninsula and Dunedin volcanoes, they occur throughout the region, and they are of widely varying age. They were also found offshore in the Endeavour and Galleon oil exploration wells (Box 15.7), aged around 35 myr and 55 myr, respectively.

Further offshore still, out on the Campbell Plateau, the three main subantarctic islands – Antipodes, Auckland and Campbell – are all remnants of relatively young, large basaltic volcanoes. The origin of these widely scattered basaltic volcanoes, with their varying ages, is not known. Although far apart, they are confined to a well-defined area. The basalt originates in the mantle, and there is no pattern to the ages that would suggest either a fixed source or one that is regularly moving.

In Central and North Otago, the basalts are aged around 5 myr (compared with 30–35 myr around Oamaru and 1.5 myr at Timaru). They extend from the coast for 80 km inland, and have a north–south extent of 65 km. There are many small, isolated exposures, the biggest area being around Kyeburn. None of the original volcanic landforms is preserved, and much of the original extent has been lost to erosion. Around Palmerston, the basalts form remnant cappings on conical hills, while Palmerston church is built of brown basalt garnished with white Oamaru limestone. Brown-weathering rock and soil is the hallmark of the volcanics.

The western lakes – Wakatipu, Wanaka and Hawea

As Box 18.1 shows, the most westerly basin and range pair in Central Otago are the Cardrona Valley and Pisa Range. Further west, the uplift associated with the Southern Alps begins to dominate, and basin and range tectonics cease. The slope of the schistosity planes becomes more variable, giving rise to more box-like and scarp and dip profile landforms, e.g. Mt Iron at Wanaka (Box 12.6).

Greater altitude, and proximity to the Southern Alps, bring us into the realm of alpine glaciation. In this zone we find the three large, glacially scoured lakes – Wakatipu, Wanaka and Hawea – surrounded by glacially eroded landforms like those illustrated in Box 19.4. Each lake occupies an ice-scoured U-shaped valley, and is dammed by the terminal moraine of the last glacier to occupy the site. The great depth of these lakes is a testimony to the phenomenal grinding power of moving ice.

Lake Wakatipu is 380 m deep, its deepest part being 70 m below sea-level. This is a good illustration of the fact that, unlike rivers, moving ice is not constrained by any base level. The rock sill at the southern end of the lake is covered by about 75 m of glacial moraine, making the rock surface there much higher than in the deepest part of the lake. This is typical of glacial lakes: ice has maximum grinding power where it is thickest, not near the glacier terminus, where it is thin. Unusually, Lake Wakatipu does not overflow its terminal moraine, but instead overflows into the Kawarau River halfway along the lake. The limited capacity of the outflow gorge occasionally leads to prolonged flooding of the lake shores (including Queenstown), when inflow from the mountains to the west exceeds the outflow capacity.

Queenstown, on the northern shore of Lake Wakatipu, is a major tourist centre. All of the mountains visible from here are made of glacially fretted schist, and, as noted above, the local angle and direction of the schistosity has a strong effect on microtopography. The Remarkables owe their distinctive appearance to a regularly spaced system of parallel streams cutting into the steep lakeward face.

The deepest parts of Lakes Wanaka and Hawea are also below sea-level. Wanaka town gives access to a large area of schist backcountry, including Mount Aspiring National Park, via the Matukituki Valley. Virtually the only view from a road of Mt Aspiring itself is from Glendhu Bay on Lake Wanaka. The mountain is known as 'New Zealand's Matterhorn', its pointed 'horn' resulting from deep glacial excavation on all flanks (Boxes 17.2 B and 19.4).

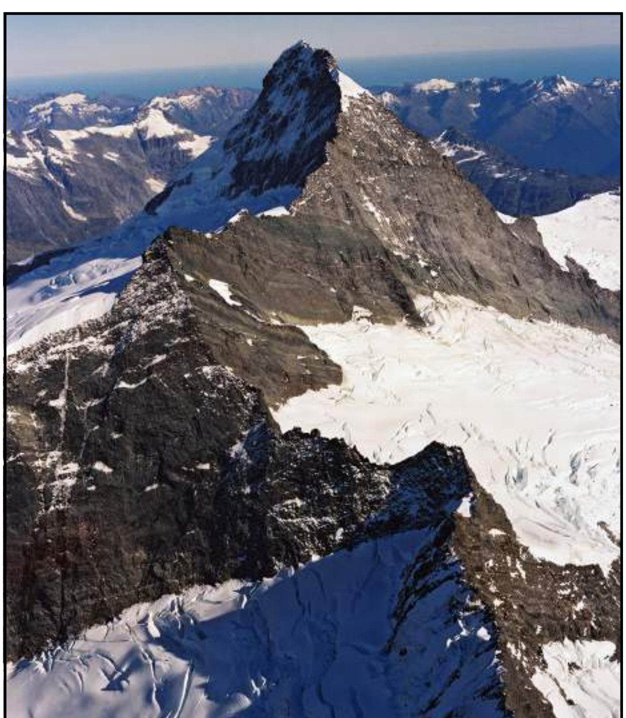


Fig. 18.4. The pointed 'horn' shape of Mt Aspiring is a result of deep excavations by glaciers that were present all around it during the last glaciation.
Photographer Lloyd Homer, GNS Science.

The Clutha River

Alexandra is the main town of Central Otago, and lies in the centre of the basin and range province. Despite being one of the driest towns in New Zealand, it was troubled around the turn of the millennium by prolonged flooding on the Clutha River. The floods were fed by the large western mountain catchment of the Clutha, and the water was retained around Alexandra by a build-up of sediment in the bed of the river. The build-up is related to the slowing of the river as it enters artificial Lake Roxburgh below Alexandra, part of a hydroelectricity scheme. The situation is one of the unexpected consequences that often occur with large engineering projects, and the flooding could recur unless the sediment build-up is removed.

The North Otago and South Canterbury Basin and Range Province

We noted earlier that the iconic Central Otago landscape of northeast-trending ranges and basins, so beloved of landscape painters, is geologically a bit strange. That is because, contrary to all reasonable expectations, the ranges lie nearly parallel to the regional tectonic stress direction instead of at a high angle to it. Well, that strangeness disappears abruptly when we pass into North Otago at the Waihemo Fault (Box 18.1).

The Waihemo Fault (more properly called the Waihemo Fault System, because it comprises a number of parallel and overlapping faults) extends for more than 120 km in a northwest–southeast direction, at right angles to the Central Otago basin and range fault systems. All of the latter terminate at the Waihemo Fault. North of the fault is another basin and range province, measuring about 120 sq km, in which the dominant trend is northwest–southeast, like the fault. In other words, the Waihemo Fault marks the boundary between two distinctive tectonic provinces, each responding to the same broad tectonic stress field, but in different ways. The deep crustal rocks of Canterbury are thinner and stronger than the schist crust of Otago, leading to these major differences in surface topography. The Waihemo Fault is part of this important crustal boundary.

Along the way, the Waihemo Fault passes close to the quintessential Otago basin towns of (from northwest to southeast) St Bathans, Idaburn, Wedderburn, Naseby, Ranfurly and Kyeburn, and then follows the valley of the Shag River to Palmerston. Strictly, of course, the Shag River valley follows the downthrown fault angle of the Waihemo Fault, and SH85 follows them both.

Waihemo Fault is a family of steeply dipping reverse faults, with the northeast side being pushed up and over the southwest side, as would be expected from the southwesterly movement of the Pacific Plate. The resulting ranges, on the northeast side of the fault-angle depression, include the St Bathans Range, the Ewe and Hawkdun ranges, the Kakanui Mountains and the Horse Range. All of them follow the northwest–southeast trend.

The North Otago and South Canterbury basin and range province is neatly bisected by the Waitaki River, one of New Zealand's great mountain-fed gravel-bed braided rivers. It, too, follows a northwest–southeast family of faults, the Waitaki Fault System. The broad fault-angle depression on the southwest side of the fault has helped to preserve the extensive area of cover strata that extends inland from Oamaru (Boxes 15.9 A, B).

North of the Waitaki River, the pattern of faults and ranges gets messy. There are two reasons for this. The first is that the northeast trend of the Central Otago basin and range system is not completely lost at the Waihemo Fault, and so the geological map north of the Waihemo Fault shows a chequerboard pattern of faults with northwest and northeast trends. The Waihemo Fault itself has a branch fault that curves around to the northeast at Kyeburn to form the Danseys Pass Fault. This fault abuts the Waitaki Fault at Otekaieke, and gives rise to Danseys Pass, the only road connection through the ranges separating the St Bathans–Shag River fault-angle depression (along the Waihemo Fault) from the Waitaki Valley.

Danseys Pass is a splendid mountain road giving great views of schist country, along with lumpy hillside ledges resulting from solifluction creep during the last glaciation (see above). The road itself is coated with locally obtained stream gravel that is quite different from the usual New Zealand road 'metal' of crushed greywacke or lava stones. Here, the dominant stones are stream-rounded white quartz pebbles.

Fig. 18.5. Danseys Pass, between Naseby and the Waitaki Valley, shows the boundary between the Canterbury and Otago schist country.
Photographer Lloyd Homer, GNS Science.

Immediately north of the Waitaki River/Waitaki Fault, directly opposite Kurow township, the valley of the Hakataramea River is a northeast-trending fault-angle basin exactly like those of Central Otago. It is separated from the Kirkliston Range to the northwest by the Kirkliston Fault, a steeply dipping reverse fault. Towards its northern end, the Hakataramea Basin bends towards the north, and forms a narrow pass separating the north-trending Grampian Mountains from the similarly trending Rollesby and Dalgety ranges and the Hunter Hills.

The last remaining basin and range pair of the province, also north-trending, is the Fairlie Basin and the low range of hills on its east side, separating the basin from the south end of the Canterbury Plains.

The second reason for the somewhat untidy pattern of basins and ranges in North Otago and South Canterbury results from the shearing/smearing effect of the sideways motion on the Alpine Fault. We saw in Southland (Chapter 16, but best seen in Box 5.3) that all the belts, or terranes, of basement rock swing round from a northwesterly to a northerly trend, and narrow, as they approach the fault – they are smeared by drag on the fault as the Pacific Plate slides towards the southwest. The effect continues to the north, and affects the basins and ranges of North Otago and South Canterbury. As they approach the Alpine Fault, these landforms are similarly dragged and smeared round to a northerly or even northeasterly trend, and narrowed. Ranges with this trend include, from the west, the Young Range (which lies east of Haast Pass), and the Barrier, Ohau, Diadem and Ben Ohau ranges. These ranges merge into the Southern Alps at the Aoraki/Mount Cook National Park.

To sum up the Central Otago–North Otago–South Canterbury basin and range province, we see that, within a simple overall setting provided by the southwesterly motion of the Pacific Plate (around 38 mm/year), the basement rocks are responding in quite complicated ways. In part, this is because there is an inherited component in the fault pattern; in other words, the compressive tectonics of the past 5 myr did not start with a clean slate.

Evidence for that inheritance can be seen on geological maps in the way that the various faults produce a chequerboard pattern of schist and greywacke blocks. Schist blocks are uplifted relative to greywacke blocks (schist, being the more metamorphosed equivalent of greywacke, lies below greywacke), but that up and down fault movement took place well before the erosion during the geological quiet period 100–25 myr ago. In the present-day cycle of tectonic activity, some – but not all – of the old faults have been reactivated.

Chapter 19

The Alpine Fault and the Southern Alps

Fig. 19.1. The Southern Alps extend for much of the length of the South Island and are margined to the northwest by the Alpine Fault. Image from NASA, credit: Jeff Schmaltz, MODIS Rapid Response Team, NASA/GSFC.
https://eoimages.gsfc.nasa.gov/images/imagerecords/67000/67355/NewZealand.A2003192.2235.1km.jpg
See also the Google Earth image in Fig. 2.3.

The South Island's Alpine Fault is one of the geological wonders of the world. It forms the boundary between the Pacific and Australian plates, where that boundary passes through continental crust and where, coincidentally, it is nearly parallel to the anticlockwise rotational movement of the Pacific Plate (Chapter 3). The latter determines that it is a right-handed (dextral) sideways-moving (transcurrent) fault. The fault is clearly visible from space (Fig. 2.3) but is not nearly so easy to see on the ground. In fact, it was not recognised as a continuous structure until 1948. It is now mapped for 600 km on land, from the mouth of Milford Sound to Cook Strait. There is more of the fault offshore at both ends, but the question of how it merges into other plate boundary structures under the sea is still to be fully determined.

Of the many different manifestations of the Pacific–Australian plate boundary as it passes through New Zealand, the Alpine Fault/Southern Alps pair is the most striking. The precise location of the pole of rotation of the Pacific Plate determines that the southwestward movement of Pacific Plate crust along the fault is not exactly parallel to it, but moves towards it at an angle that varies along the fault, between 0° in Marlborough and a maximum of 12° in the vicinity of Aoraki/Mt Cook, the highest part of the Southern Alps. That oblique approach, of Pacific Plate continental crust towards the backstop of Australian Plate continental crust on the west side of the Alpine Fault, creates compression and squeezes the rock of the Pacific Plate. Rock of continental crust composition (in this case greywacke and schist 2.7 times heavier than water) is lighter than the rock of the earth's mantle beneath (more than three times as heavy as water) and therefore cannot go downwards when it is squeezed – it is too buoyant. It cannot go sideways, either, because there is an infinitely long extent of similar rock on three sides. The only possible response to the compression is for the rock to go upwards and form mountains, which is why the Alpine Fault and Southern Alps are considered a pair.

Calculating Rates of Movement

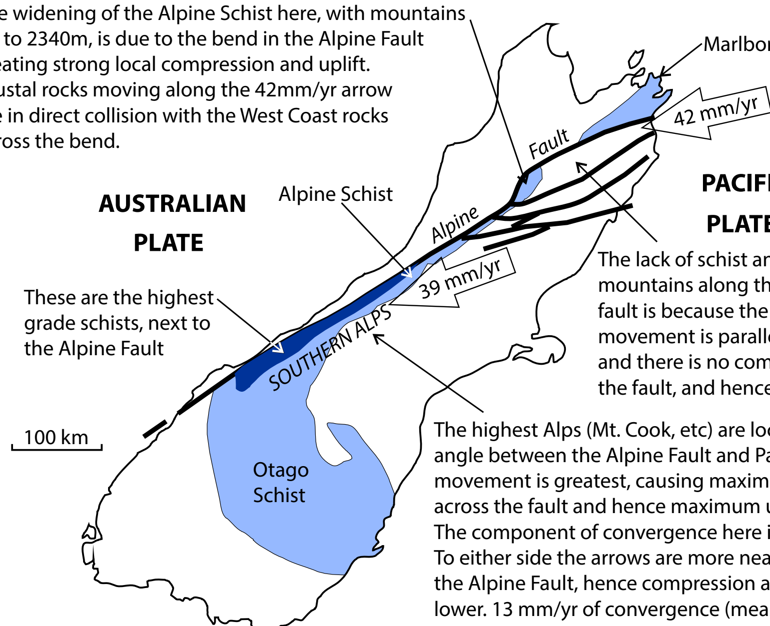
We noted in Chapter 3 that the Pacific Plate is rotating anticlockwise at 1°/myr around a pole of rotation that is presently located at latitude 60°S, longitude 180°. It has been doing so for 45 myr, and like a rotating record on a turntable (a spherical turntable in this case), the actual speed of movement depends on distance from the rotation pole – the further away, the faster the movement. Thus, Boxes 19.1 A and 19.3 show that the speed of Pacific Plate motion through the South Island varies between 42 mm/year at the northern end and 35 mm/year at the southern end because northern South Island at the fault is further from the pole of rotation than southern South Island.

Those rotation speeds are 'global' values, derived from global measurements and calculations of plate movements. But how do they translate into actual motions of rock particles at, and adjacent to, the plate boundary? That question is addressed below, but first we need to remember that the numbers apply only to the last 5 myr, this being the time the pole of rotation has been in its present position. Prior to that, the rotation was the same, but the different positions of the pole (Box 5.5 C) meant that there was much less compression at the boundary.

Box 19.1 A. Uplift of the Southern Alps along the Alpine Fault plate convergence.

The Southern Alps exist because continental crust of the Pacific Plate is pushed against continental crust of the Australian Plate at the Alpine Fault, causing crustal rocks to rise because they are too buoyant to be subducted downwards.

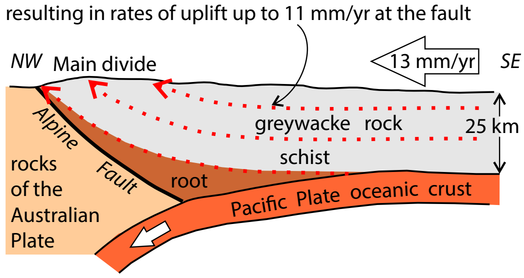
The widening of the Alpine Schist here, with mountains up to 2340m, is due to the bend in the Alpine Fault creating strong local compression and uplift. Crustal rocks moving along the 42mm/yr arrow are in direct collision with the West Coast rocks across the bend.



The lack of schist and high mountains along this part of the fault is because the plate movement is parallel to the fault, and there is no compression across the fault, and hence no uplift.

The highest Alps (Mt. Cook, etc) are located where the angle between the Alpine Fault and Pacific Plate movement is greatest, causing maximum compression across the fault and hence maximum uplift.
The component of convergence here is 13 mm/yr.
To either side the arrows are more nearly parallel to the Alpine Fault, hence compression and uplift are lower. 13 mm/yr of convergence (measured at right angles to the fault) translates to uplift of 10-11 mm/yr — see the diagram.

Individual rock particles follow these paths to the surface, resulting in rates of uplift up to 11 mm/yr at the fault



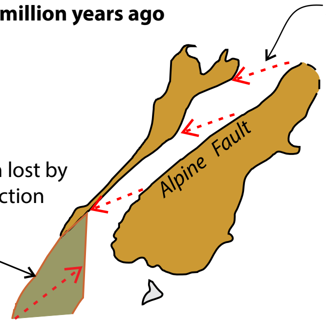
See also Box 19.3 A

See how greywacke reaches the surface east of the main divide, and schist (metamorphosed greywacke from deep crustal levels) west of it. The schist is the hottest rock, so heat flow is highest between the Alpine Fault and the main divide, causing hot springs.

Box 19.1 B. Material lost along the Alpine Fault plate boundary, either by erosion or subduction.

The South Island reconstructed to 6 million years ago

All the rock from this strip has been uplifted in the Alps and destroyed by erosion



The rock in this area has been lost by the opposite process - subduction beneath Fiordland

The Alpine Fault plate boundary, and its transition into oblique subduction zones both to the south and north, is a wonderful natural laboratory of fast-moving, three-dimensional tectonic processes. As such, it attracts a lot of research from the international scientific community of geologists, geophysicists, geomorphologists, marine geologists, radiometric daters and others. Earthquake prediction figures highly in the research, because a super-active plate boundary fault like the Alpine Fault generates earthquakes that are powerful (more than 7.5 on the Richter scale) and geologically frequent (every 200–300 years on each sector of the fault – it doesn't rupture as a single unit).

One result of all this research is that the numbers relating to rates and consequences of plate motion are constantly being revised and refined. A recent set of calculations on the southern sector of the Alpine Fault made use of the displacement of glacial landforms that have been dated at 18,000, 58,000 and 79,000 years. Older landforms have been displaced more than younger, and the right-handed sideways offsets were 435 m, 1240 m and 1850 m, respectively. All of these numbers come with plus or minus error factors, reflecting the fact that landforms and radiometric ages do not lend themselves to super-precise measurement. Instead, the numbers are averages. A simple comparison of distance versus time along the southern sector of the Alpine Fault over the past 79,000 years, using this data set, gives a long-term average speed of 23 mm/year, remembering that the actual displacement occurs in jumps during earthquakes.

Going back to the global value, which says there is 35 mm/year of plate movement in the southern end of the fault, what has happened to the missing 12 mm/year? The answer is that some of it has been distributed among other faults in a zone 80 km wide on the southeast side of the Alpine Fault, where it is causing oblique dextral reverse faulting, while the rest of it is being taken up by pure reverse faulting in faults further away on both sides of the Alpine Fault. Thus, some of that 12 mm/year is causing the pure reverse movement on the basin and range faults of Central Otago that were discussed in Chapter 18.

How much sideways movement?

Determining the amount of sideways displacement on faults is a tricky business. Sideways movement on faults was recognised much later than vertical movement, because when you see a fault exposed in a cliff or quarry only the vertical offset is obvious. If the fault is displacing dipping strata, which is a common situation, and you see an apparent sideways offset in a horizontal exposure like a shore platform, that doesn't mean anything. This is because pure vertical offset on such a fault causes apparent vertical offset in a vertical exposure, and apparent sideways offset of the dipping strata in a horizontal exposure. By the same token, pure sideways motion on such a fault has exactly the same effects (Box 6.6 D). The only thing that might indicate definite sideways movement is the occasional presence of scratches on the fault-plane surface (which isn't always exposed) that record the direction of the last movement. Otherwise, you need a marker that is clearly identifiable on both sides of the fault, and that is either a straight line (for a recent movement) or a vertical structure of some kind.

The first record in the world of definite sideways displacement on a fault was in North Canterbury following the Canterbury earthquake of 1888. A fence on Glynn Wye Station fortuitously crossed the trace of the Hope Fault at right angles, and was offset by 3 m (Fig. 19.2). The movement was purely sideways (right-handed, like the majority of our faults), with no vertical offset.

Shortly afterwards, the 1891 Mino-Owari earthquake in Japan created a near-vertical new scarp 6 m high, and at the same time offset a road sideways (left-handed) by 4 m, thus establishing both sinistral and oblique movement. The famous 1906 San Francisco earthquake also caused well-documented sideways movement on the equally famous plate-boundary San Andreas Fault of California, the Alpine Fault's counterpart on the opposite side of the Pacific. Later observation of structures built across the San Andreas Fault showed that in some places there is continuous creep on the fault, which is not surprising given the continuous movement of plates.

As noted earlier, the Alpine Fault was not recognised as one long, continuous structure until 1948. When this realisation came about, however, its sideways displacement was also determined, because by great good fortune the fault displaces a sequence of vertically dipping basement rock terranes that are geologically distinctive and easily matched. The best marker is the Dun Mountain Terrane of ocean-floor crust. There is a perfect match between the Red Hills Range of southern West Coast and the Red Hills of southeast Nelson (Figs 3.1, 3.2; Box 19.3), which are offset by 480 km. In 1948, this was a revolutionary determination, because before that there was little indication of the huge scale of such phenomena. In fact, much of the progress in geological understanding over the past 150 years has been a gradual realisation of the vast scale on which geological processes operate.

Fig. 19.2. The Hope Fault in North Canterbury ruptured in an earthquake in 1888 producing a sideways displacement of 3 m marked by this offset fence, which was broken by the quake but had been repaired by the time this photo was taken by Alexander McKay. From GNS Science.

Now, remembering that actual offset on the fault itself is less than the plate movement rate in the southwestern part of the Alpine Fault (see above), we should ask whether 480 km is the total offset between the Pacific and Australian plates across the South Island plate boundary. The answer to that question is that it is only about half, because global calculations of plate movements suggest about 1000 km of total displacement. The missing half of the displacement is distributed through rocks and structures in a wide zone on either side of the actual fault.

How much vertical movement?

We've dealt with sideways movement along the Alpine Fault, but what about the vertical movement that creates the Southern Alps? As discussed above, the angle between the approach of the Pacific Plate and the fault is greatest in the vicinity of Aoraki/Mt Cook, the highest point of the Southern Alps – which is no coincidence since the rate of uplift is directly related to the amount of compression. Thus, 39 mm/year of plate speed at this location, approaching at an angle of 12°, converts to compression at right angles to the fault (i.e. shortening) of 13 mm/year. That number then converts to uplift of 10–11 mm/year, because the rock particles are not travelling vertically upwards, but are following an oblique path determined in part by the steep eastwards dip of the fault (Box 19.1 A). Interestingly, exactly the same uplift rate has been calculated from carbon-dating of uplifted terraces in valleys that cross the fault.

Moving along the Alpine Fault, to either side of Aoraki/Mt Cook, the convergence angle declines and so does the rate of uplift – to 6–8 mm/year. The northernmost part of the Alpine Fault – the Wairau Fault, which follows the Wairau River (more correctly, vice versa) from Nelson Lakes National Park to Cook Strait – happens to lie exactly parallel to the plate trajectory. Thus, there is no compression at the fault here, and no mountains adjacent to it on the southeastern side (Chapter 15).

As far as can be determined, the present situation has pertained for the past 5 myr. If you multiply the rate of uplift of 10 mm/year by 5 myr, you get a total uplift of 50 km around Aoraki/Mt Cook. The mountain, of course, is nowhere near that high, being 3.764 km at its summit, give or take a few metres to allow for the top falling off from time to time (see below). The 50 km figure may be a top-end estimate, but at any rate you get the picture about the rate of erosion in high mountains, especially when they sit athwart the weather systems of the Roaring Forties. The fact that many billions of tonnes of rock have been uplifted many kilometres also gives you a good idea of the amount of energy that is expended in plate movement.

Dragging Schist to the Surface

Most of the rock being pushed up into the air along the Alpine Fault is greywacke of Canterbury's Torlesse (Rakaia) Terrane. Greywacke is itself partially metamorphosed by heat and pressure, and it is underlain by (and in northern Otago passes laterally into) its more metamorphosed equivalent, schist. As shown in the cross section in Box 19.1 A , the schist and greywacke that comprise the continental crust of Canterbury are underlain by oceanic crust. A partition takes place to the east of the Alpine Fault, whereby the heavy oceanic crust passes downwards obliquely beneath the

Australian Plate (partial subduction), while the schist and greywacke peel off and are pushed up the interface provided by the Alpine Fault. Thus, the deepest of these layers – schist – emerge at the surface directly adjacent to the fault, while the shallower greywacke layers form the Main Divide and points east.

Schist (Boxes 15.3 A, B) comes in various grades (degrees of severity) of metamorphism, defined by characteristic minerals. Chlorite schist is succeeded by mica schist, followed by garnet schist and finally oligoclase schist. The highest grades of schist, from the deepest levels, reach the surface only next to the Alpine Fault, and then only in the central reaches where uplift is fastest. Because it comes from the greatest depths, schist is relatively hot, and gives rise to a band of elevated heat flow and hot springs close to the fault (Box 9.5).

Bends in the Alpine Fault

In detail, the surface trace of the Alpine Fault can be wavy, because it is a thrust fault that dips steeply to the southeast and intersects valleys and ridges. In broad view, however, it follows a remarkably straight line from Milford Sound to Nelson Lakes, where it veers left around the only big bend in its trace. The apparent reason for this bend is discussed in Chapter 15.

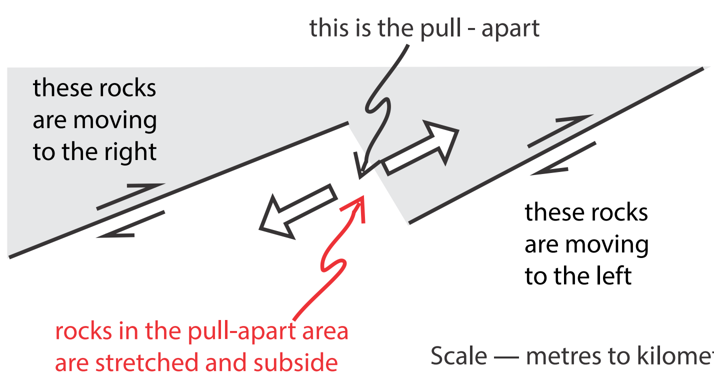
Bends and offsets in sideways-moving faults have interesting consequences. Box 19.2 A explores the case of Hanmer Springs basin, where the right-handed Hope Fault does a right step that is opening up a pull-apart basin in the region of the step. On the Alpine Fault at Nelson Lakes, the converse happens – a right-handed fault does a left step. This leads to opposite sides of the fault pushing together in the vicinity of the big bend. As elsewhere on the fault, compression equals uplift, so it is no coincidence that the widest area of dragged-up schist along the northern part of the fault occurs in this bend. The uplift also makes it easy to say where the Southern Alps begin – with the 2000 m-plus peaks of the St Arnaud, Travers, Mahanga and Ella ranges and the Spenser Mountains in and to the northeast of the bend.

The question of the southern limit of the Southern Alps is less easy to answer. Mt Aspiring (3027 m) and its fellow peaks in the Mount Aspiring National Park are clearly part of the Southern Alps. South and southwestward from there, 2000 m ranges persist down either side of the northern end of Lake Wakatipu, and into the Darran Mountains of northern Fiordland. So perhaps the Southern Alps per se can be said to finish at SH94, the road to Milford Sound. The Alpine Fault passes out to sea at the entrance to Milford Sound, and while the topography of Fiordland is equally grand and impressive as the Southern Alps, it is lower and different (Chapter 17).

Box 19.2 A. Pull-apart basins – Hanmer Springs Basin.

Sedimentary basins — portions of the earth's crust that subside for a period of time — provide accommodation for sediment to accumulate. In the long term they give us coal, oil, gas, sand, aggregate, clay, rock-salt, limestone, phosphate, gypsum, groundwater and other useful materials.

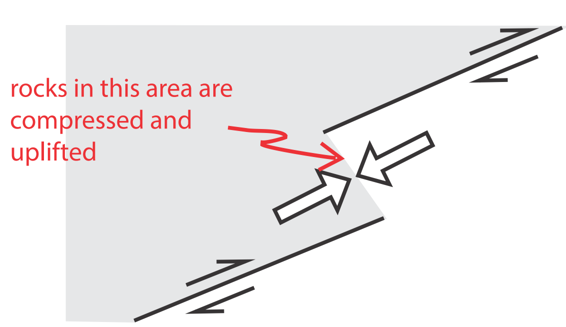
Large-scale basins (e.g Murchison, Canterbury, Great South — Boxes 20.6, 15.7, 15.14 respectively) are tectonically driven by deep-seated forces. Smaller scale basins are generally closely tied to fault offsets — i.e. they are tectonic but relatively superficial. The three main types of fault (Boxes 6.6 C, G) each have up-and-down components of motion which create basins adjacent to the faults called **fault-angle depressions** (e.g. the cross-sections in Box 15.5). Fault-angle depressions do two things — they preserve covering strata from being eroded away, and they can provide space for more sediment to accumulate. Examples of fault and fault-angle basins are the Basin and Range province of Otago (Box 18.1) and the North Canterbury Fold and Thrust Belt (Box 15.6 — reverse faults); and the main Marlborough Faults (Boxes 15.4 A, 15.5 — a combination of sideways and up-and-down motion).
There is one more class of fault-controlled sedimentary basin, called, in full, Strike-slip Pull-apart Basins. *Strike-slip* refers to sideways motion — slip parallel to the strike (direction of a horizontal line drawn on the plane of the fault). It is also known as transcurrent or transform motion.
This is how they work —
This diagram is in map view.



A *right-handed* (dextral) strike-slip fault (relative motion shown by half-arrows) makes a step or sidestepping bend to the *right*.

Scale — metres to kilometres

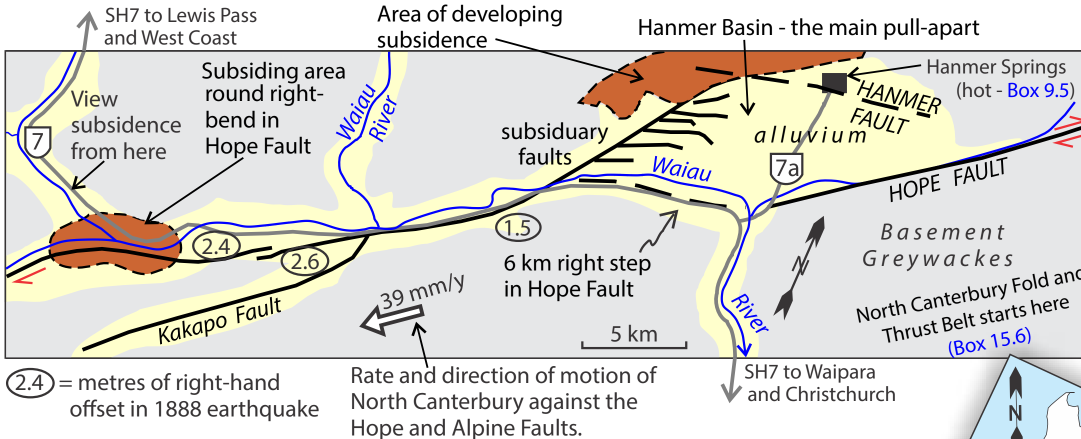
The opposite of the pull-apart basin is the compressional uplift, where a *right-handed* fault makes a *left step* or bend.

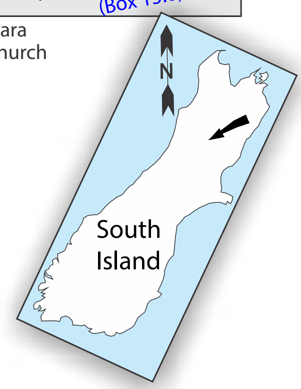


Exactly the same things happen with left-handed faults.
Left-to-left = pull-apart,
left-to-right = compression.

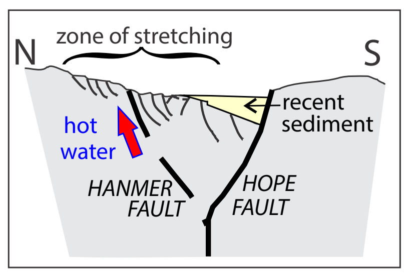
Box 19.2 B. Hope Fault – pull-apart basin and 1888 earthquake.

New Zealand does not presently have any large-scale (tens of kilometres) pull-aparts. The Hanmer Basin is a good example of a medium-scale (kilometres) pull-apart at a right-step on the Hope Fault.





Section across Hanmer Basin, showing the wide zone of stretching. The lack of preserved Tertiary age strata in the basin indicates that they had all been eroded away before the basin formed - i.e. it was an area of high relief before the basin opened during the last one million years (maybe less).



The North Canterbury Earthquake, 1 September 1888, and the Hope Fault

This was one of New Zealand's larger historic earthquakes, with a magnitude estimated at 7.0 to 7.3. It was felt over most of New Zealand, and toppled the spire of the Anglican Cathedral in Christchurch. However, it was centred on the Hope Fault in the area west of Hanmer Springs now traversed by SH7 (see map), and caused movement of the fault with right-handed sideways offsets of up to 2.6 metres — measured by displacement of fences. Scientifically, it was an important earthquake because it was the first one in the world to cause observed and recorded sideways movement. In fact we know now that the much more powerful 1855 Wellington earthquake caused nearly 20 metres of sideways movement on the Wairarapa Fault, but there were no fences or roads in that area at that time, and very few people, so it was not recorded then. Sideways displacements measured in 1888 are shown on the map, and range from 1.5 to 2.6 metres. Older displacements have also been able to be measured and dated (radiocarbon dates) by the offset of natural features such as glacial moraines, enabling us to calculate two things — a long-term rate of movement on the Hope Fault for the past 17,000 years of about 14 mm per year, and a probable return period for earthquakes of the 1888 magnitude of between 90 and 170 years (it will not escape notice that it is already more than 100 years since the last one). A neat thing about this area is that pull-apart basins occur on all scales from tens of metres to the Hanmer Basin. They are very young, and so can be studied by the way they distort the ground surface, i.e. in great detail. The westernmost depression marked on the map occurs at the junction of the Hope and Boyle Rivers, and is caused by a right step of about one kilometre in the Hope Fault. Half of it has been washed away by the rivers, but the remaining southern half can be viewed from SH7 as the road leaves the Hope Fault and climbs northwards towards the Lewis Pass.

How Old is the Alpine Fault?

We've established that the Southern Alps are 5 myr old, but the Alpine Fault itself is much older. Just how much older has been a topic of considerable debate in the 60 years since its recognition. It was originally thought that it may date back to the tectonic deformations that accompanied the accretion of the basement terranes to Gondwana during phase one of New Zealand's formation more than 100 myr ago (Chapter 5). However, evidence like that described in Chapter 17, where the Alpine Fault cuts across an earlier plate boundary structure, the Moonlight Tectonic Zone, has made it definite that the Alpine Fault in its present manifestation has existed for 25 myr following the vitalisation of the plate boundary 25 myr ago (Chapter 3).

Where Can You See the Alpine Fault?

Despite its vast importance and huge scale, the Alpine Fault is very difficult to see on the ground, and it's not hugely impressive when you do see it. Probably the best thing here is to note that the steep western face of the whole Southern Alps is basically the scarp of the fault – degraded by erosion like all fault scarps, but its surface expression nevertheless. Another important thing to note is that the fault doesn't look like the clean, crisp fractures shown in Box 6.6 C. Faults form by incremental growth – generally 1–2 m during each earthquake. Over a long period, the rock adjacent to the fault plane gets ground down into a fractured state – a breccia if the bits are sizeable, or a 'gouge' or 'pug' if the bits are small. This is especially true if the two sides of the fault are pressing together. The Alpine Fault has hundreds of kilometres of movement, and we should expect a strong development of fault gouge, which is exactly what we find – up to a 1 km width of it!

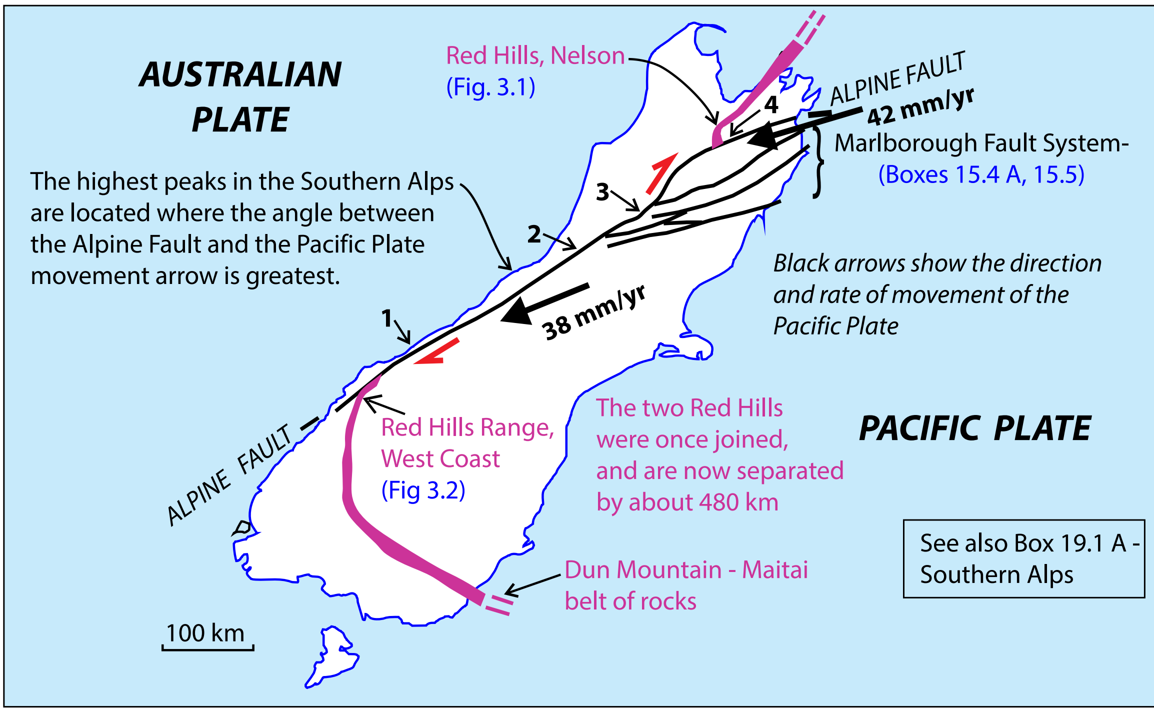
The rock in the fault gouge is ground up to form cataclasite, or in extreme cases the friction generated by fault movement actually melts the rock. The molten rock then solidifies to form natural glass (known as mylonite), which is typically distributed in veins among the cataclasite.

We don't often see the fault gouge because it is soft and easily eroded, and normally forms valleys. The Alpine Fault's gouge can, however, be seen in roadside exposures on the Arthur's Pass highway (SH73) about 5 km west of the turn-off to Lake Brunner, on the big bend underneath the Bald Range above the Taramakau River. As expected, it is soft and 'puggy'.

It is more common to see the surface trace of the most recent offset, recording the most recent earthquake. This will take the form of an escarpment 2–3 m high, as described in Box 13.3 A. Box 19.3 shows four places where the recent trace of Alpine Fault movement can easily be seen. One is at Inchbonnie by the Arthur's Pass highway, close to the exposure of the fault pug described above. From SH73, take the side road to Inchbonnie, just west of Jacksons. Cross the Taramakau River, and about 500 m before the right turn signed to Moana, look for a property on the north side of the road called Stonui. A few metres west of the Stonui driveway, the road descends about 3 m – this is the offset. The road step has been smoothed since the offset, so the roadside fence actually shows the step better. It is down to the west; in other words, the latest movement(s) have been raising the Southern Alps, as we would expect. There is no indication at this location of sideways displacement.

Box 19.3. The Alpine Fault.

The Alpine Fault forms the sliding boundary between the Australian and Pacific tectonic plates.



Numbers 1 to 4 are places where the trace of recent movements on the fault can be seen close to a major road.

1. SH6 at Haast
2. Close to SH73 (Arthurs Pass road) at Inchbonnie
3. SH7 Lewis Pass to Reefton, near Maruia Springs
4. SH63, Kawatiri Junction to Blenheim, near Branch River and close to the power house of the Argyle Power Scheme

What is the Alpine Fault?
It is one of the world's major plate boundary faults with a dominant sideways component of fault displacement (the San Andreas Fault of California and the Anatolian Fault of Turkey are others). It separates the Australian and Pacific Plates, and its character derives from the fact that the motion of the Pacific Plate is towards the southwest, close to being parallel with the actual fault. This motion determines that the Alpine Fault is right-handed or dextral in character. The fact that the motion is not absolutely parallel to the fault, but is slightly towards the fault, gives a component of compression at the fault. This causes the uplift of the Southern Alps (Box 19.1 A), because the two plates here are both carrying a cap of continental crust, and continental crustal rocks (the rocks we see at the surface all over New Zealand, plus what is beneath them to a depth of 25 km or so) are too buoyant to be pushed down a subduction zone into the dense rocks of the earth's interior. Hence, they have no way to go except up.

The best place to see the recent fault trace, and its cumulative effect as recorded in displaced river terraces, requires a short walk from the Lewis Pass road (SH7) where it crosses the Alpine Fault just west of Marble Hill. Look for the Marble Hill Camping and Picnic Area, on the north side of the road about 6 km east of Springs Junction; this is the start of the Lake Daniells walking track. About 50 m beyond the campsite name and rules board, the track drops a couple of metres over a step – this is the recent scarp on the Alpine Fault. Look across the paddock to the right (northeast) from the step, and you will see a low concrete wall about 30 m long at right angles to the fault step. The wall was built in the 1970s by the late Professor Frank Evison across the Alpine Fault, so that next time the fault moves in an earthquake the break in the wall will tell us exactly what the movement was. At the time of writing (2009), there has not been an earthquake here. Geological wits usually cannot resist observing that Professor Evison clearly gummed up the fault with his wall and stopped any more movement!

Fig. 19.3 shows the view back from the wall to the track. There are several terraces and steps in the ground here, and it is important to sort them out. The step caused by the fault runs straight across the paddock from the point in the track described above to the middle of the wall, and beyond it to the northeast. It is crossed by several other steps that are the edges of river terraces at different heights; these steps have been progressively offset by movements on the fault. Clamber around for a while and compare terraces across the fault (starting with the lowest), and you will see that the displacement across the Alpine Fault is oblique, i.e. a combination of right-hand sideways movements and up and down movements (up to the southeast, of course, because the Southern Alps are rising to the southeast of here). Horizontal movement is generally three or four times greater than vertical movement – e.g. 9 m horizontal against 1.6 m vertical for one terrace edge.

The few metres of vertical and lateral motion we see recorded here are, of course, only the most recent in a long line of similar movements spread over 25 myr. This is also a suitable place to stand at the bottom of the fault scarp with one foot on the riser (the Pacific Plate) and the other on the flat ground of the Australian Plate on the other side of the fault. There aren't many places in the world where one can straddle a major plate boundary!

Fig. 19.3. The Alpine Fault trace at Marble Hill Camping and Picnic Area can be seen displacing river terraces. Photographer Bruce Hayward.

Glaciation

The Southern Alps contain extensive ice fields and valley glaciers, especially in the highest regions around Aoraki/Mt Cook. During glaciations they become much more extensive, occupying most of the mountains and with glaciers reaching some distance out onto the surrounding lowlands – especially on the wetter western side. The last glaciation peaked around 20,000 years ago, and there is a glaciation every 100,000 years. Relating this to rates of uplift discussed earlier, we see that an uplift of 10 mm/year around Aoraki/Mt Cook amounts to 1 km between glaciations. Consequently, we are not going to see landforms from earlier glaciations in our high mountains, because they are long gone to erosion. Even in the 20,000 years since the last glaciation, there has been 200 m of uplift.

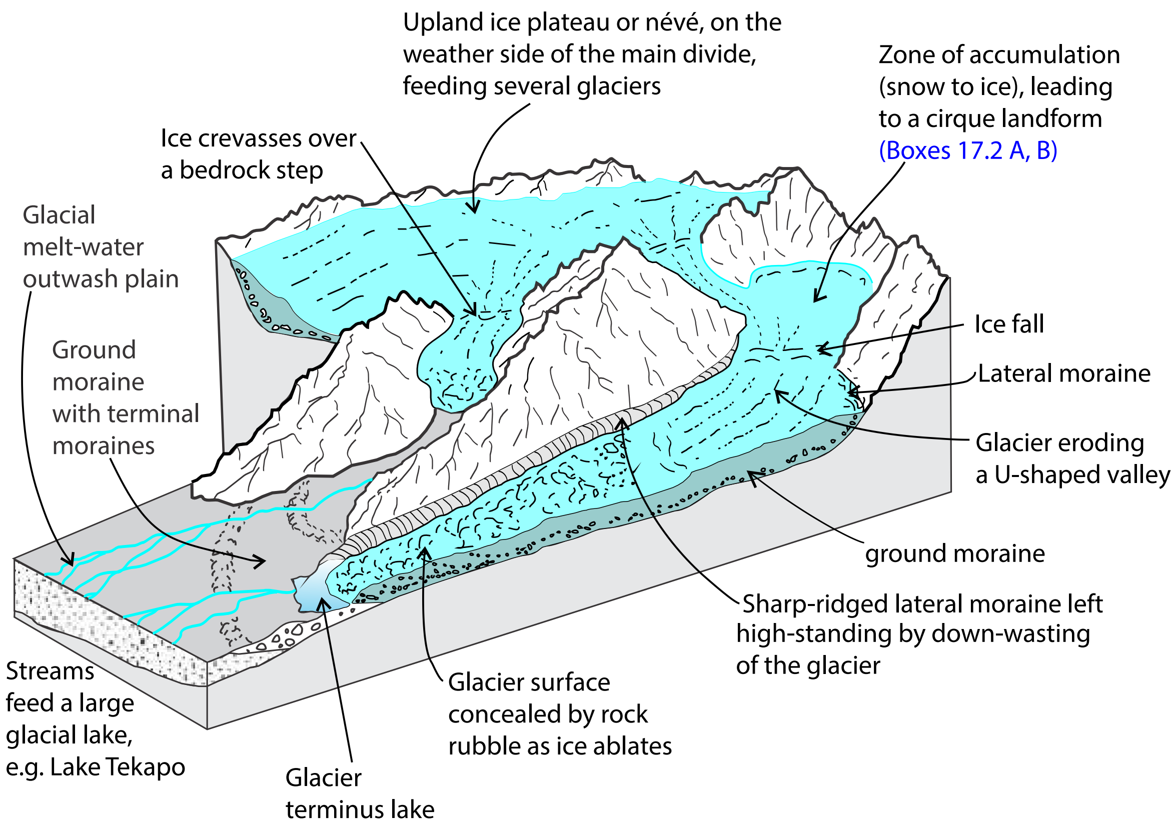
Because of rapid erosion in high mountains, records of our glacial history are found only in the surrounding lowlands, usually in glacial deposits like moraines. Even there, the record of older glaciations is very sketchy because successive glaciations tend to destroy the deposits of earlier ones. And because a glaciation involves repeated advances and retreats of glaciers, the record of older advances tends to be destroyed by later advances in the same glaciation.

New Zealand mountains display the classic features of mountain glaciation that were first recorded in the European Alps – Box 19.4 has a summary of the erosional and depositional features. Erosional features such as U-shaped valleys are found well beyond the area presently bearing glaciers (e.g. throughout Fiordland) as relics of the last glaciation. How well these landforms are preserved depends to some extent on the rock types. The solid granites and gneisses of Fiordland, for example, hold glacial shapes better than the highly fractured greywackes of the Southern Alps.

There are significant differences between the wet and dry sides of the Southern Alps as regards glaciers. Typical wet-side glaciers are Fox and Franz Josef on the West Coast (Chapter 20). They descend the steep western face of the Southern Alps from a névé (ice field) that is liberally supplied with new snow. They move fast (metres per day), advance and retreat on decadal timescales in response to climate cycles, and descend well into the rainforest.

Fig. 19.4. The highest parts of the Southern Alps around Aoraki/Mt Cook contain numerous ice fields and valley glaciers. Here New Zealand's highest mountain has the Hooker Glacier on the west side (left) and Tasman Glacier on the east (right). Photographer Lloyd Homer, GNS Science.

Box 19.4. Mountain glaciers – a general model for New Zealand.



Bear in mind that during the last ice age, 20,000 years ago, ice in the New Zealand mountains was much thicker and more extensive than it is now. Bear in mind also that there were many earlier glaciations, and that the mountains are rising between 6 and 11 mm per year. At the beginning of the third millennium glaciers worldwide are receding. See Boxes 17.2 A, B for glaciated landforms.

In contrast, dry-side glaciers have smaller supplies of ice and move much more slowly. Recent warming is having dramatic effects on them. They are thinning (e.g. 100 m in 100 years in the Tasman Glacier), and, since 1970, melting to produce terminal lakes that are getting larger quite rapidly. Professional mountain guides note that high-altitude ice fields here are also reducing in thickness and extent.

By far the greatest concentration of glaciers and ice fields is located around the highest mountains, in Aoraki/Mount Cook National Park. This is the only area east of the Main Divide where you can access glaciers and their moraines from short, low-altitude walks. When approaching the national park on SH80, remember that the four separate glaciers present today – Mueller, Hooker, Tasman and Murchison – combined during the last glaciation to form one large glacier that filled the valley now holding Lake Pukaki. The lake is held in by the terminal moraine of that glacier, and the road runs mostly on its lateral moraine, which gives an idea of the depth of the ice. Lake Pukaki also has an extensive lake-head delta (post-glacial), made of gravel and rock flour washed out from the glaciers. It is crossed by braided channels of the Tasman River, and during wind storms is the source of much dust – the raw material of loess found further west (Chapter 15).

Glaciation outside the Southern Alps

The thickness and rate of movement of ice sheets and glaciers is dependent on the rate of supply of fresh snow. It would be expected that, during a cold glacial period, the high-precipitation region (several metres a year of rainfall equivalent, equating to a much greater thickness of snow) of the southwestern corner of South Island would be home to a vigorous set of névés and glaciers. That was indeed the case, and thus we have the fjords of Fiordland – deep, U-shaped glacial valleys scoured well below present sea-level, and drowned by the post-glacial rise of sea-level.

As noted earlier, the granites, diorites and gneisses of Fiordland are tough and relatively unfractured rocks that are resistant to the mechanical erosion that predominates there. Thus, glacial erosional landforms are very well preserved. Fiordland's famous waterfalls nearly all drop from hanging valleys (Box 17.2 A).

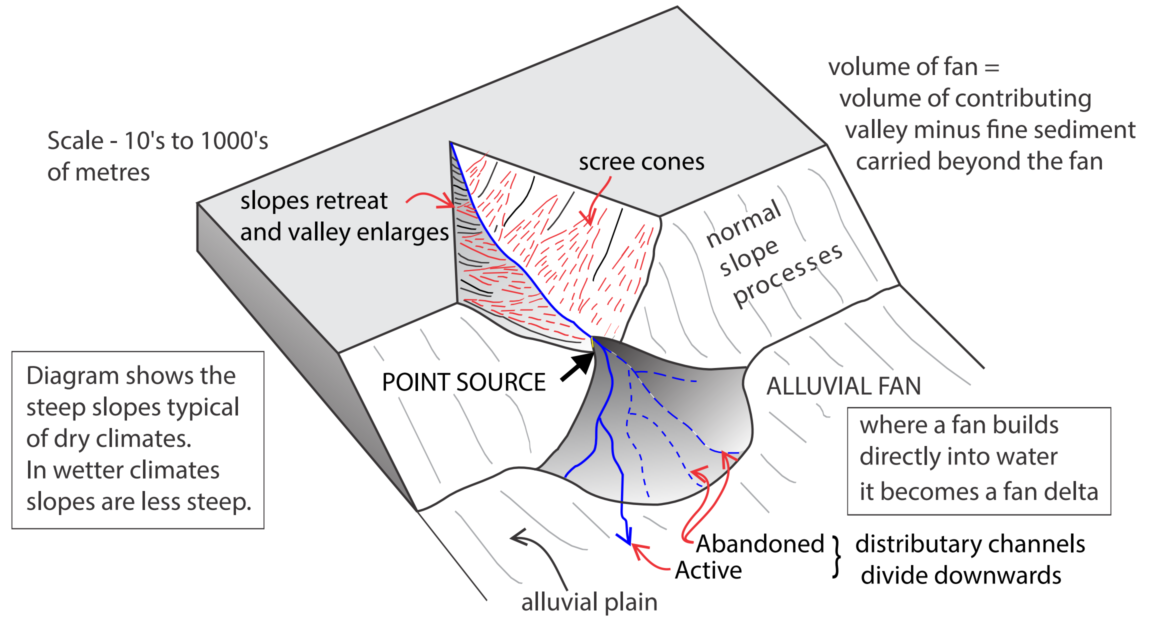
Today, there are tiny remnant glaciers in Fiordland, but elsewhere along New Zealand's mountain chain we generally infer past glaciation purely from the landforms and moraine deposits. The ranges of Central and North Otago and South Canterbury supported glaciers, as did the ranges of east and northwest Nelson. The northernmost Southern Alps in Nelson Lakes National Park spawned voluminous glaciers that scoured Lakes Rotoiti and Rotoroa, and occupied the valley of the upper Buller River (Fig. 22.6). The Inland and Seaward Kaikoura ranges are the highest mountains outside the Southern Alps, but are on the dry side of the country; they were also glaciated and now hold snow for most of the year. The Rimutaka Range and, possibly, the Tararua Range of southern North Island were also glaciated, as was Mt Ruapehu in the centre of the North Island, which still has a number of small remnant glaciers.

Scree, Alluvial Fans and Debris Flows

The presence of heavily jointed and fractured greywacke rocks at high altitude leads to extensive mechanical breakage from frost and ice. Water trapped in cracks is frozen from the outside inwards, expanding as it does so and thus levering open the cracks. The result is huge numbers of loose blocks of greywacke, which form screes. In some parts of the dry eastern side of the Southern Alps, screes are the dominant feature, whereas on the wet western side they are less prevalent because vegetation binds the blocks. All screes post-date the retreat of glaciers.

The alluvial fan (Box 19.5) is a related feature, formed when loose debris generated in a gully is carried to the mouth of the gully, and there builds a cone or fan at the point where the angle of the slope lessens. Gullies enlarge themselves by what is called headwards sapping. In mountains, physical/mechanical erosional processes dominate over chemical weathering. Slopes retreat continuously and generate rubble continuously, so that as the gully grows, so does the alluvial fan.

Box 19.5. Alluvial fans and debris flows.



An alluvial fan forms where a stream loses momentum at a change of slope, and dumps its coarse sediment load. The fan then builds outwards and upwards.

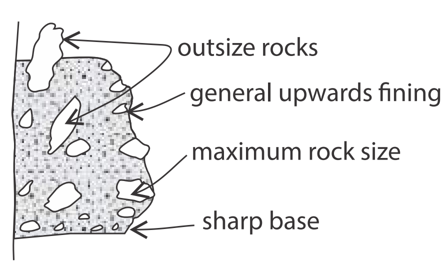
Processes on the fan

Scree cones, which commonly feed gravel onto fans, are fed by individual rocks falling from above. Gravel and sand are carried to the apex of the fan by stream action. They are then dispersed across the fan by a combination of stream action and periodic debris flows.

A **debris flow** is a type of sediment gravity flow, i.e. a flow of sediment caused by gravity acting directly on the grains of sediment (see Boxes 6.5 A, B) for another kind of sediment gravity flow). A debris flow is a mixture of mud, sand, gravel and water which flows smoothly, with little internal turbulence. Flowing wet concrete is a debris flow.

The proportion of debris flow deposition to normal stream deposition on the fan varies according to climate. In dry climates stream action is minimal and occasional rainstorms initiate debris flows. In wetter climates like New Zealand, streams and debris flows are approximately equal in importance.

A typical debris flow deposit is from 0.3 to 3 metres thick, and is structured internally so that the largest rocks are located from 10 to 30 cm above the base.



The outsize rocks float in the general background sediment.

In New Zealand alluvial fans are best developed, along with scree cones, on the dry, eastern, rain-shadow side of the Southern Alps.

Fig. 19.5. This alluvial fan at the mouth of a gully can be seen from the Arthurs Pass highway (73) near Cass in the Waimakariri Valley. Photographer Bruce Hayward.

The way in which gravel is spread around the alluvial fan is interesting. The stream issuing from the gully flows across the fan surface, of course, and does carry some material. However, most of the work is done by periodic debris flows during heavy rain events. The flood generated in the gully bulks up by entraining sand and gravel, and transforms into a type of sediment gravity flow called a debris flow. The difference between a sediment-charged river flow and a gravity flow is that, in the river, gravity pulls the water and the energy thus generated is transferred to the sediment. In the gravity flow, however, gravity pulls the sediment and energy is transferred the other way, to the water. A river in flood can carry up to about 30% sediment by volume, while a debris flow typically contains around 70% . There is also a halfway stage, called a hyper-concentrated flood flow, with around 50% sediment by volume; this develops in very sandy sediment.

Debris flows are very important in building the ring plains around large volcanic cones (where they tend to be called lahars; chapters 9 and 11), and in building alluvial fans. Fans can be very large – in the big basin and range systems of the western United States such as Death Valley, they can be many kilometres across. A typical succession of events is as follows. First, a flood of clear water – from rain in the mountains, the collapse of an ash dam holding a volcanic lake (e.g. at Mt Ruapehu in 1953, resulting in the Tangiwai disaster; Box 9.2 A), or an eruption into an ice- and snow-filled crater (e.g. Mt St Helens, Washington state, USA, following the eruption of 1980) – picks up loose rubble on the steep slope and bulks up into a debris flow. Particularly large and destructive debris flows are generated from heavy rain falling on a large volcano that is covered with loose rocks and ash from an eruption (e.g. at Mt Pinatubo in the Philippines in 1991).

Second, in a valley the debris flow follows the river channel and picks up water, causing it to debulk. It is the bulk of the flow that carries big rocks, so as the flow debulks big rocks are left stranded and the sandier material remaining transforms to a hyper-concentrated flood flow. The periodic lahars from Mt Ruapehu's crater lake do this on the lower slopes as they approach the Tangiwai rail and road bridges.

Ultimately, the continuing uptake of water from the river transforms the flow back into a heavily sediment-laden river flow. On an alluvial fan, the debris flow spreads out over the surface of the fan, and so debulking rarely occurs.

There are several kinds of sediment gravity flow. Boxes 6.5 A, B describes turbidity currents, in which the sediment/water mixture is highly turbulent as it flows. Debris flows, in contrast, move in a laminar fashion, whereby layers of the flow slide over one another without turbulence, resulting in a different kind of deposit (Box 19.5). Bigger rocks are jostled to the edges and the top – consequently, it is common to see big rocks (or, as at Mt St Helens, items like bulldozers and trucks) half-protruding from the top surface after the flow has stopped.

We're not quite done with debris flows. Where an alluvial fan builds into a lake or the sea, a fan-delta is built. Debris flows entering the water from the fan undergo a different kind of transformation while taking up water, this time to a turbidity current that then flows on to the deepest water. Large debris avalanches entering the sea can retain their identity (though taking up water and being modified to varying extents), carrying house-sized rocks into the deep basin. This was the origin of the Parnell Grits layers in the Waitemata Basin of Auckland (Chapter 6).

Rockfalls and Landslides

The combination of rapid uplift, high mountains, fractured rock, large earthquakes and high rainfall makes New Zealand mountains prime candidates for frequent rockfalls, and a sizeable one occurs here every year. During an earthquake there may be many simultaneous rockfalls, as during the 1929 Murchison earthquake (Chapter 20). Commonly, rockfalls dam rivers. In the case of big rivers, the dam is overtopped and swept away in a day or two – as happened on the Buller in 1929. Small rivers blocked by large rockfalls can, however, be permanently dammed – Lake Stanley in Kahurangi National Park, for example, formed during the 1929 earthquake. The country's biggest landslide-dammed lake is Waikaremoana in the North Island (Chapter 12).

One of the most dramatic erosional/depositional events is the rockfall, which has enough run-out space to transform into a fast-moving avalanche of rocks and ice and air – technically also a debris flow. Such events do not come more dramatic than the collapse of the summit of Aoraki/Mt Cook on 14 December 1991 (Boxes 19.6 A, B).

Rock Glaciers

The rock glacier (Fig. 19.6) is an interesting phenomenon, being a halfway stage between an ice glacier and a scree. Scree at high altitude is commonly bound together by interstitial ice. Increasing weight on the deposit from scree accumulating higher up causes it to deform and flow, just like an ice glacier. The form of a rock glacier is identical to an ice glacier, but the heavy load of rocks creates great friction, and rock glaciers are slow-moving. They are thought to date back as far as the last glaciation, and may still be active.

There is a continuum between a rock glacier as described, and deeper-seated slides, slumps and flows. There are many such features in the Southern Alps. They can entrain an entire mountainside, and they are regarded as long-lived, dating back to the maximum extent of the last glaciation. A creeping landslide containing more than 100 million cu m of debris is located on the west side of the Sealy Range, above the Mueller Glacier (Fig. 19.7). The old and new Mueller Huts are situated on its upper surface.

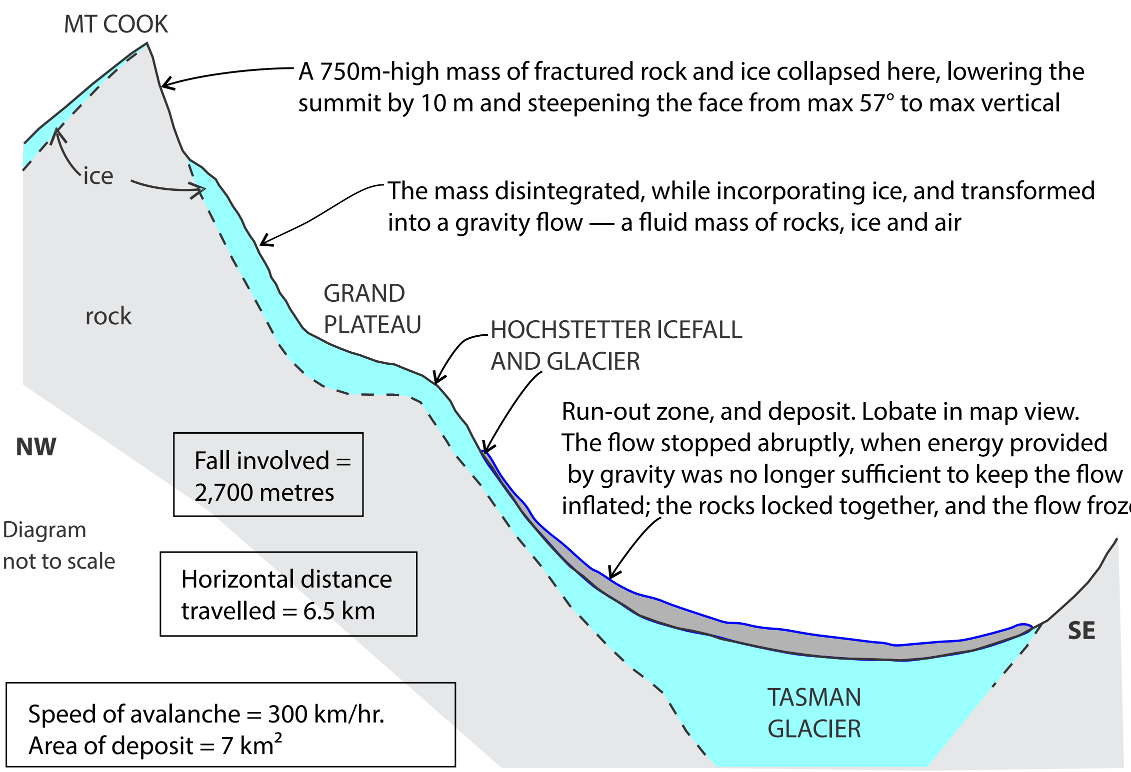
Box 19.6 A. Rock avalanches.

Every particle of soil and rock on a slope is subject to a constant downwards pull by gravity. In combination with atmospheric chemistry (rock weathering — Boxes 6.1 A, B) and the action of water and ice (Boxes 12.5 A-C, 19.4), gravity shapes the landscape. Tectonic uplift provides the necessary height.

Faster uplift = greater height = faster erosion = steeper, higher slopes = more frequent gravity-driven collapses

Boxes 12.5 A-C deal with the general spectrum of gravity effects. These boxes involve the extreme end-member of the spectrum — slope collapses leading to rock avalanches, which are more common in mountains than you may wish to know. The special case of collapses on high volcanoes is covered in Chapter 9, especially Box 9.2 B.

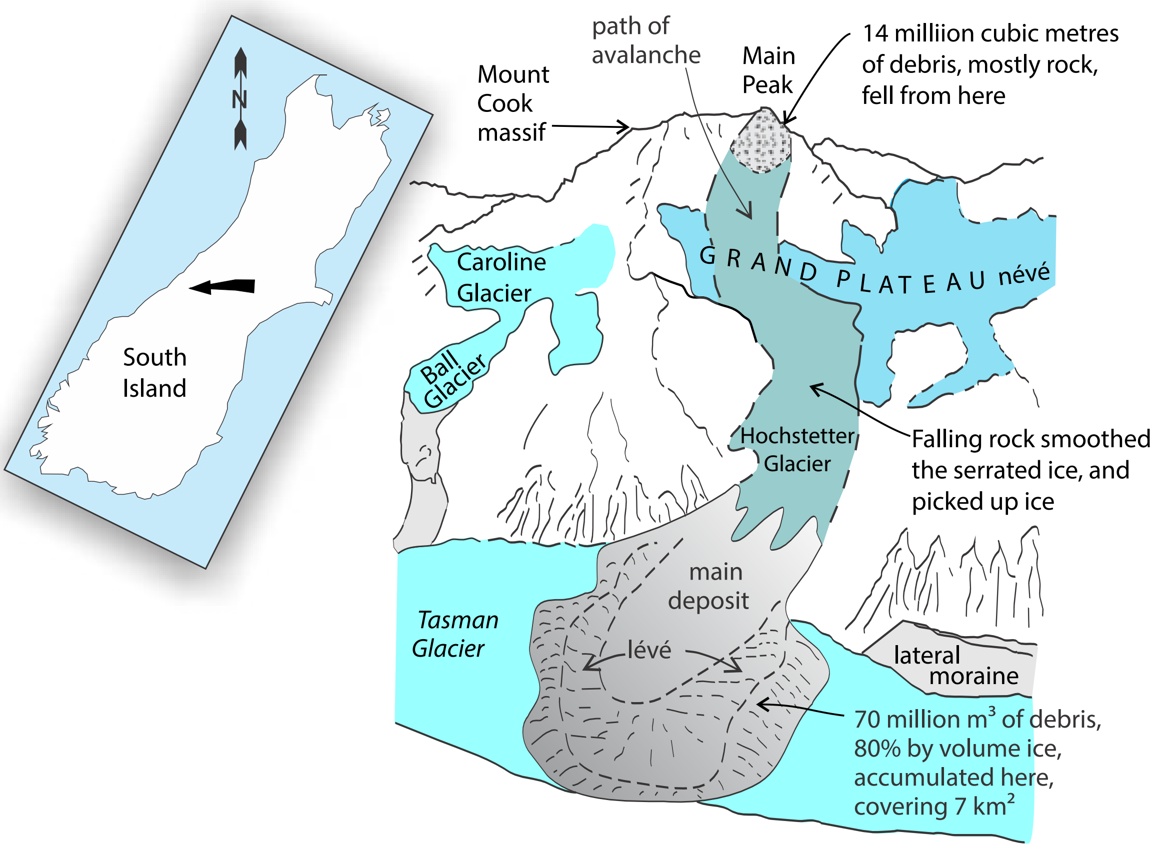
The collapse of the summit of Aoraki-Mt Cook, 14 December, 1991



The Mt Cook event occurred at night. It was witnessed by climbers in a hut at the top of the Hochstetter Icefall, who described the great noise, and the sparks, from clashing rocks. A seismic signal equivalent to a magnitude 3.9 earthquake was generated, and an air blast carrying wet mud at a speed of 300 to 400 km/hr was felt by people in a hut 5km up-glacier. previous rock avalanches from the Mt Cook massif occurred in 1974 and 1873.

(continued on next page)

Box 19.6 B. Rock avalanches *(continued from previous page).*



Oblique aerial view of the 14 December 1991 Mt. Cook rock avalanche. From a photograph.

No-one was hurt in the Mt. Cook event. Elsewhere, in the European Alps and the Andes, villages have been buried by rock avalanches, and many lives lost. A notorious event occurred at Elm in the Swiss Alps in 1881. A mountain side which had been undercut by a quarry collapsed on a Sunday morning, when many villagers were walking to church. Eye-witness accounts established that the rock mass was airborne as it ricocheted off the quarry floor. It landed halfway across the valley, ran some way up the opposite side, made a right angle turn to the left, and ran down the valley. Many people were killed, and many houses destroyed. A survivor heard the noise, walked outside and was engulfed by the flowing rocks up to his neck, just as the flow froze and stopped.

Some rock avalanches are associated with earthquakes, which is to be expected since high mountains are found in tectonically active areas. The deposits commonly impound lakes by damming rivers. Many New Zealand mountain lakes formed that way. The largest and most celebrated is Lake Waikaremoana in the Urewera Ranges, North Island. Most such lakes are temporary, being soon drained by erosion of the dam, or infilled by sediment.

Nearly 50 old rock avalanches have been logged in the Southern Alps between Mt. Cook and Arthur's Pass, with volumes between 1 million and 500 million cubic metres. The largest known rock avalanche occurred in Iran — 20 cubic kilometres. New Zealand's largest known avalanche deposit is 14 cubic kilometres; it dammed Lake Green in Westland.

Fig. 19.6. A rock glacier in the Hawkdun Range, North Otago. Photographer Lloyd Homer, GNS Science.

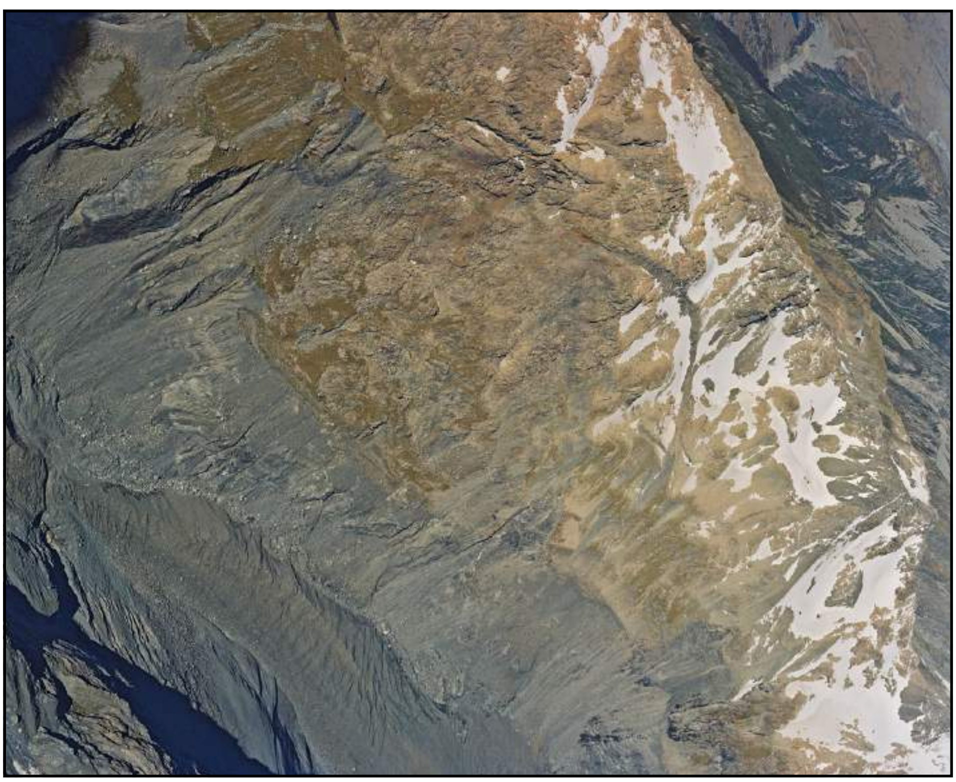
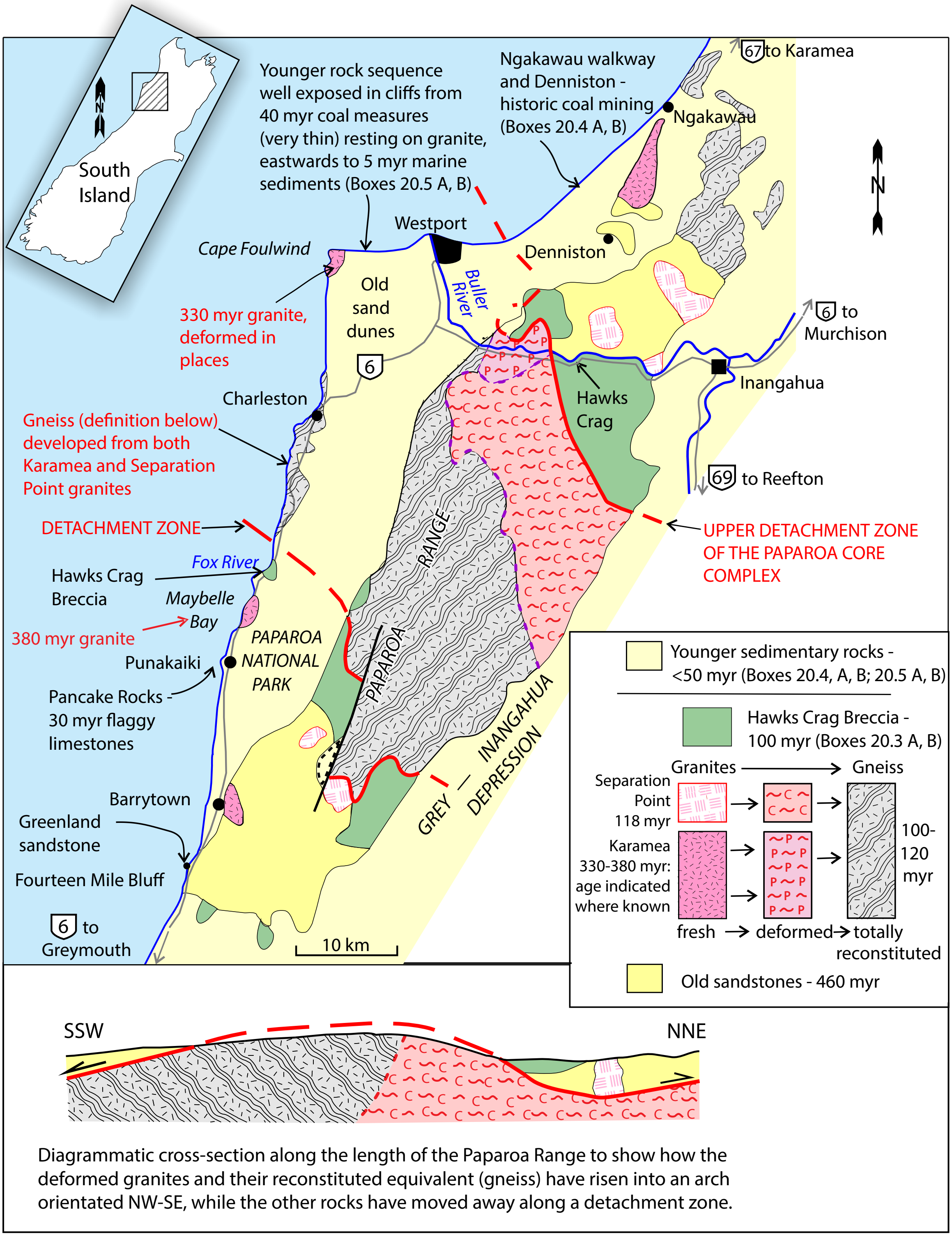


Fig. 19.7. Most of this part of the western side of the Sealy Range above the Mueller Glacier in Mt Cook National Park is a slowly creeping giant landslide. Photographer Lloyd Homer, GNS Science.

Chapter 20

The West Coast



Diagrammatic cross-section along the length of the Paparoa Range to show how the deformed granites and their reconstituted equivalent (gneiss) have risen into an arch orientated NW-SE, while the other rocks have moved away along a detachment zone.

Fig. 20.1 A. West Coast metamorphic core complexes - Paparoa Range. Other West Coast metamorphic core complexes are shown on the next page.

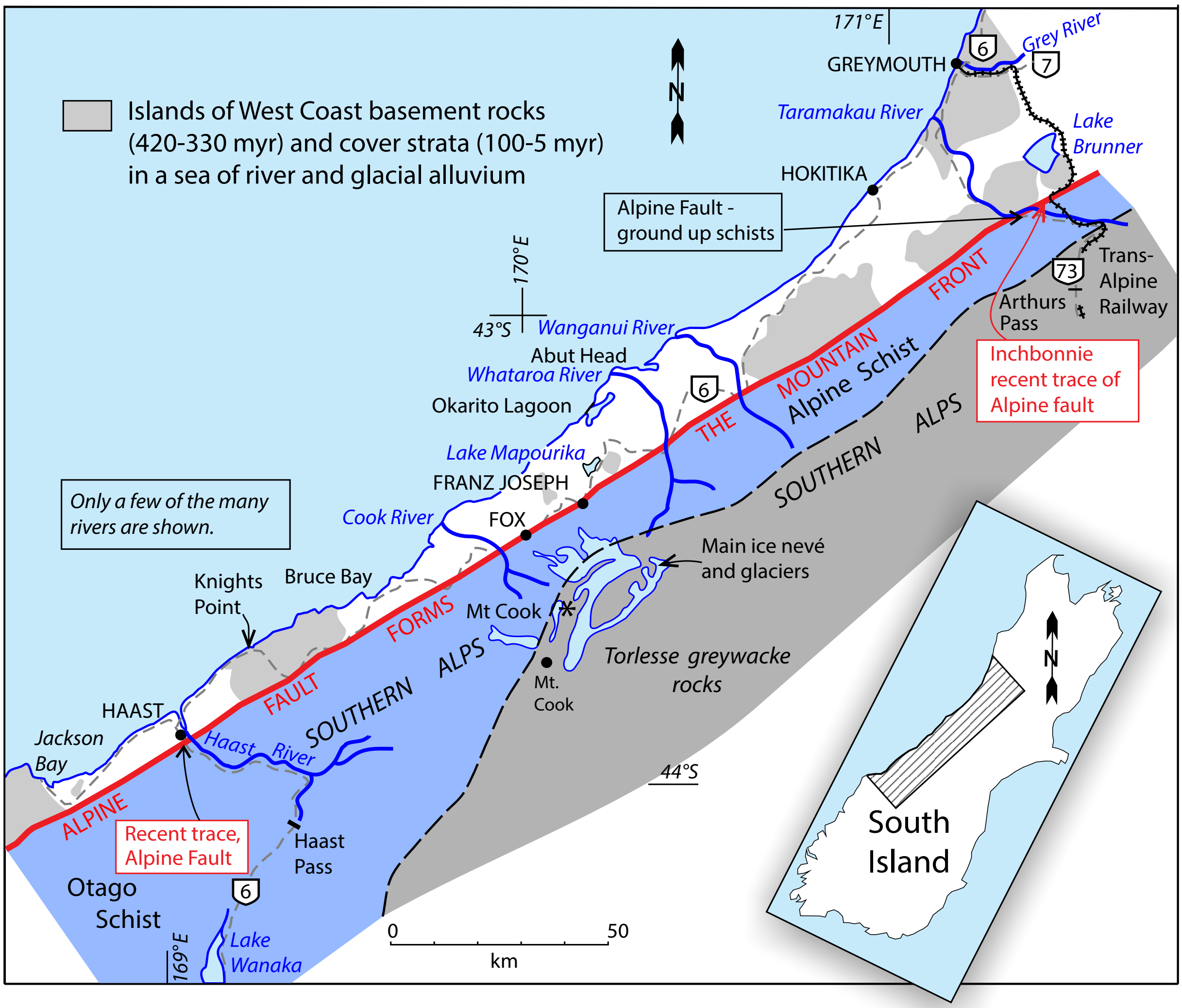


Fig. 20.1 B. West Coast metamorphic core complexes – Greymouth to Jackson Bay.

In geological terms, the West Coast comprises everything west of the Alpine Fault. The region stands out as having the most varied geology of any part of New Zealand.

If we include its geological continuations to the north (northwest Nelson) and south (Fiordland), the region contains the oldest rocks in New Zealand (dating back to 510 myr), practically all of the granite and the only high-grade metamorphic gneisses. Its location close to the Tasman Sea spreading centre around 100 myr ago resulted in the formation of metamorphic core complexes and their accompanying ancient basin and range geology. It contains a thick and varied suite of younger cover strata, including New Zealand's highest-grade and oldest workable coals. More recently, its close proximity to the Alpine Fault plate boundary has led to the formation of a wide range of structural phenomena (e.g. inverted sedimentary basins) and the most frequent occurrence of large earthquakes in New Zealand. It is traversed by huge amounts of sediment moving from the Southern Alps to the sea, and it contains alluvial gold deposits that drove early local economic development.

The Fiordland and northwest Nelson extensions of West Coast geology are treated separately in chapters 17 and 21. This chapter deals with the long, narrow strip of land that begins at the mouth of Milford Sound and extends northwards, widening as it goes, to the Buller River and Murchison. The region has a high rainfall, being on the Roaring Forties side of the Southern Alps, and, in large part, is covered in rainforest.

Old Rocks of the Coast

Fig. 20.2 shows Fiordland, the West Coast and northwest Nelson in their original relationship, before the Alpine Fault commenced its 480 km sideways offset. In order to understand the old rocks on the West Coast and wider region, we first need to delve into the tectonostratigraphic terrane story of the growth of Gondwana (Chapter 5). Of the terranes (chunks of continental crust) that make up the basement rocks of New Zealand, the first two in order of attachment are found here: the Buller and Takaka terranes. Remember that there is no correlation between the age of the rocks that make up a terrane and the time at which the terrane 'docks' with its new continent, except that the rocks must be older than the time of docking.

The Buller Terrane consists mostly of a flysch sequence of alternating sandstones and mudstones (Box 6.5 A), called the Greenland Group, which is similar to many others around the country but differs in two ways. First, the sandstones are dominated by the mineral quartz, which is unusual in this country; and second, many exposures show a well-developed axial plane cleavage – a layering in the rocks that cuts across the original stratification and records strong compression and folding. Axial plane cleavage is a halfway stage in the production of schistosity planes (Box 15.3 A). In places, compression was so strong that the rocks have been metamorphosed to a coarsely layered gneiss.

The original age of the Buller Terrane rocks is known from rare fossils called graptolites, extinct colonial organisms that look like pieces of fretsaw blade (Fig. 21.2). They indicate that the rocks were deposited (by turbidity currents on a submarine fan adjacent to somewhere on Gondwana) early in the Ordovician period, around 480 myr ago. So, if the Buller rocks were laid down 480 myr ago, when did the terrane dock with our part of Gondwana? And when did the terrane next door, the Takaka Terrane, dock with the Buller Terrane?

If we assume that the folding and metamorphism of the Greenland Group rocks accompanied the docking event (i.e. that the rocks were strongly compressed by the docking process itself), we can therefore date the docking if we can date the folding. Potassium–argon radiometric ages suggest that folding and metamorphism took place around 440 myr ago. To back this up, Buller Terrane rocks have younger cover strata, dating from around 415 myr ago (Early Devonian period), at Reefton. These shallow marine, fossiliferous strata were deposited after deep erosion of the Greenland Group, long after docking. Thus the Buller Terrane docked with Gondwana around 440 myr ago, and was an integral part of the super-continent by 415 myr ago.

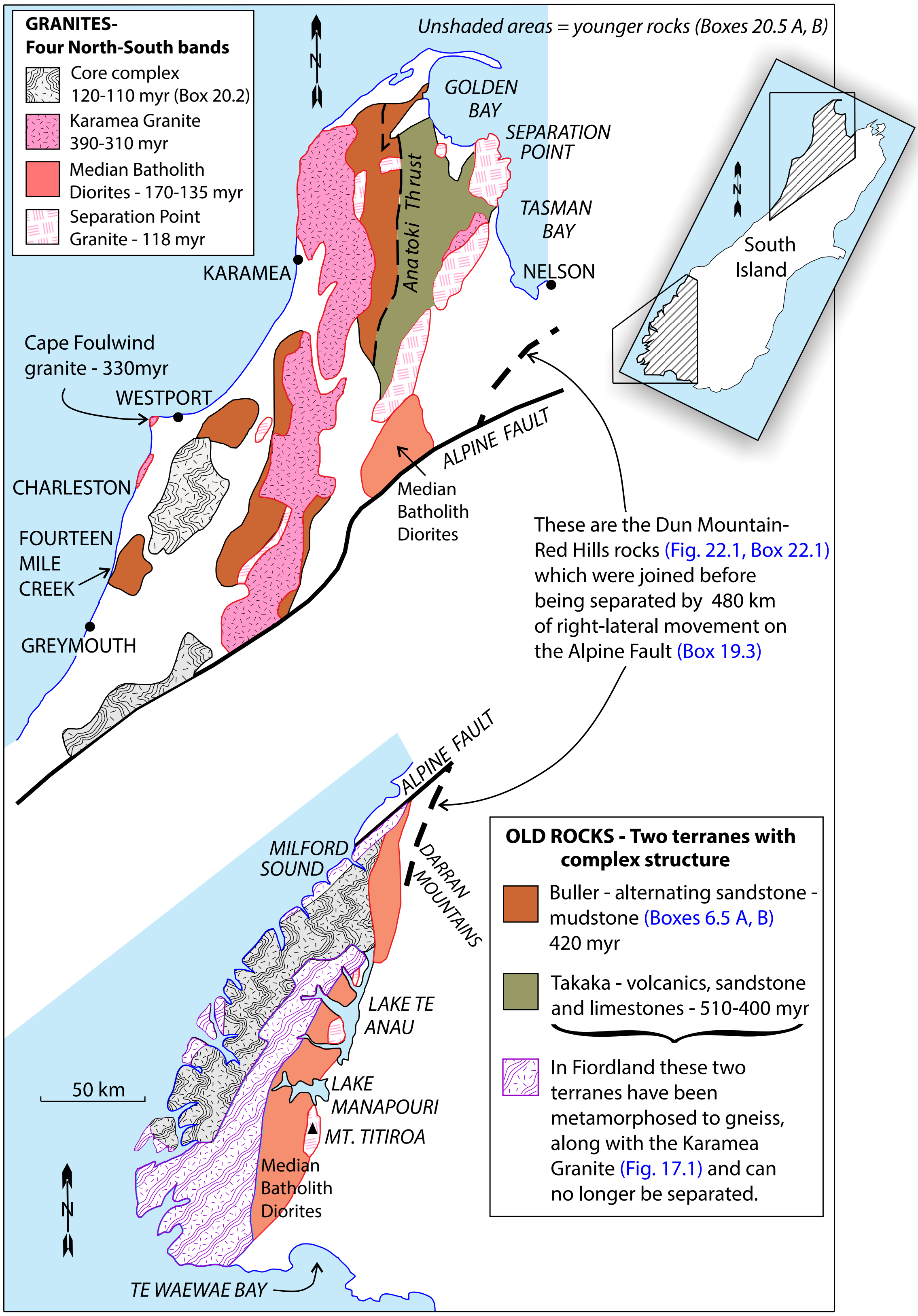


Fig. 20.2. Old rocks and granites of Northwest Nelson–Westland–Fiordland.
(Discussed in the following Box)

Box 20.1. Old rocks and granites of Northwest Nelson–Westland–Fiordland, discussion.

Northwestern South Island and Fiordland are depicted in their original relationship, before being separated by movement on the the Alpine Fault.

Old Rocks

North of the Alpine Fault the old rocks are aged between 510 and 400 million years, and are divided into major tectonic terranes (Box 5.3 explains) that have been brought together by tectonic processes at a major fault called the Anatoki Thrust. The western rocks (**Buller Terrane**) comprise alternating sandstones and mudstones of the Greenland Group (Boxes 6.5 A, B for detail of the rock type), which are steeply folded (Boxes 15.3 A, B for diagrams). They are best seen in coastal exposures on the West Coast alongside SH6, at 12 Mile, 14 Mile and 17 Mile bluffs north of Greymouth. The structure is simplest at 14 Mile, right by the road. Age — Ordovician, from very rare graptolite fossils, about 420 myr. The eastern rocks (**Takaka Terrane**) occur only in northwest Nelson. They comprise volcanic rocks, sandstones and limestones (the Arthur Marble — Fig. 21.1), with very complex structure. Best seen adjacent to SH60 on Takaka Hill (Arthur Marble — e.g. track to Harwoods Hole), and in the Cobb Valley in Kahurangi National Park (volcanics and trilobite rock). Age — middle Cambrian to Silurian, about 510 to 400 myr. Some of these rocks originated in the northern hemisphere and the tropics before being transported to the Gondwana margin by plate tectonic processes. South of the Alpine Fault, i.e. in Fiordland, deeper crustal levels are exposed at the surface, and there these rocks, along with the Karamea Granite, have been severely metamorphosed to form gneisses (Boxes 15.3 A, B). They are no longer distinguishable from each other. Most easily seen in the outer portions of Milford Sound, or the westernmost part of the middle arm of Lake Te Anau.

Granites

Granites are plutonic igneous rocks (from Pluto, god of the underworld) that have cooled and crystallised slowly, underground, from a molten state (Box 20.2 for general information). North of the Alpine Fault there are four discrete bands of granite and related rocks, all trending north—south. In age order, the **Karamea Granite** (390-310 myr) is related to the growth of the Gondwana supercontinent. Most easily seen in the Buller River valley, from SH6. The **Median Batholith** — (170-135 myr) is a subduction-related group of igneous rocks that for the most part are not granites in the strict sense (silica content of 70% plus, with crystals of quartz) but diorites (silica content of 60 to 70%, without crystals of quartz). Median Batholith rocks comprise one of the dozen or so unrelated terranes brought together by plate tectonic processes at the Gondwana margin (Box 5.3). Best seen at the Nelson Lakes, or from the Te Anau-Milford Sound road. **Separation Point Granite** (118 myr) was a late product of the subduction beneath Gondwana that resulted in the assembly of the various terranes; subduction ceased around 100 myr ago. Best seen in Abel Tasman National Park or in the southern arm of Lake Te Anau. The gneissic (i.e. strongly sheared and metamorphosed) granites of central Fiordland and the Paparoa-Charleston area were formed in features called **metamorphic core complexes** (Box 20.2 for details) between 120 and 110 myr ago. These structures were formed in the early stages of crustal stretching that preceded the opening of the Tasman Sea (80-55 myr ago). Best seen at Charleston, by SH6, on the coast south of Westport, or at Milford Sound village. South of the Alpine Fault, in Fiordland, the Karamea Granite has been severely metamorphosed, along with the sedimentary rocks of the Buller and Takaka Terranes, to form gneiss (Boxes 15.3 A, B). It is no longer distinguishable. The other three granite belts are younger than that metamorphism, and are therefore easily distinguishable.

The Takaka Terrane (for more on this, see Chapter 21) has a much wider age range than the Buller, with its youngest rocks (at Baton River in northwest Nelson) dating from around 415 myr ago (Early Devonian). Like the Reefton cover strata of the same age, these have a shallow marine origin and are fossiliferous, but they are not cover strata, and the important thing to note is that Takaka Terrane strata were still accumulating – somewhere – while the Buller Terrane had already been docked for some time.

As a note in passing, the presence of strata of a similar age, with similar fossils, at Baton River (northwest Nelson) and Reefton (north Westland), geographically quite close, led to assumptions in earlier times that the two areas were close together and presumably in their present relationship in the Devonian period. The point about the tectonostratigraphic terrane theory is that it turns such assumptions on their heads, stating that because two things are close together now, it should never be assumed that they have always been so.

The best evidence of the docking date between the Buller and Takaka terranes comes from igneous intrusions that cut across both terranes after they had docked – so-called 'stitching intrusions'. Buller Terrane rocks are intruded by the Karamea Granite, which is dated at 390–310 myr. The granite itself does not intrude into Takaka Terrane rocks, but a related pluton, the Riwaka in Nelson, dated at 370 myr, does. Thus, we deduce that the two terranes were definitely docked with each other after 415 myr (Early Devonian), definitely before 370 myr (Late Devonian), and very probably before 390 myr, the apparent beginning of the Karamea Arc. That is, the Karamea magmatic/volcanic arc (of which we see only the roots, the batholiths, none of the volcanic superstructure having survived) records subduction at the Gondwana margin taking place after the two terranes had come together, and before the arrival of the next terrane. We looked at the wider implications of this fact in Chapter 5. All this occurred long before the next terrane on the block, the Brook Street Terrane, had even begun to accumulate.

The subduction process that brings terranes to continental margins normally drives a magmatic/volcanic arc that is fed from sources around the 100 km depth contour of the subducting slab of oceanic lithosphere. As Boxes 5.1 and 5.2 explain, subducting slabs have to respond in some way to the docking of a new terrane, and the magmatic arc is our only long-term record of this response. The arrival of the new terrane usually disrupts the subduction process and temporarily switches off the arc. A terrane should never be intruded by its 'own' magmatic arc, so to speak.

The interesting thing about the relationship between the Buller and Takaka terranes and the subduction zone beneath Gondwana is that, apart from the old magmatic/volcanic arc that forms the lowest part of the Takaka Terrane, and represents a much earlier regime from somewhere distant, there are absolutely no indications of contemporary volcanic activity. If there was an arc on Gondwana accompanying the arrival of the two terranes, none of its products was reaching the Gondwana coast.

Where to see the rocks

The best places to see Greenland Group strata are north of Greymouth – they form the coast between Twelve Mile Bluff and Seventeen Mile Bluff, which is easily accessed from SH6 (but only at low tide; beware of the incoming tide and surging waves). Two notable things about the alternating layers of sandstone and mudstone are that they are almost always steeply dipping to vertical, and that they have a well-developed axial plane cleavage (see above). Exposures vary in structural complexity. The structure is simplest about 100 m south of Fourteen Mile Creek, where, at the back of the beach, beds dip steeply to the north and get younger in that direction (Box 6.5 A describes how you work this out). Mudstone beds display vertical cleavage planes, which change direction when they enter sandstone beds and dip steeply to the south (this is called cleavage refraction). The relationship between dip, younging direction and cleavage tells us that there is an anticlinal fold to the south (Fig. 20.3).

Karamea Granite can be seen at the coast at Meybille Bay, just north of Punakaiki's Pancake Rocks. It is a true granite that contains glassy feldspar and both white and black mica. This rock can also be seen around the Cape Foulwind lighthouse, where it is a younger member of the Karamea suite, dated at 330 myr. The age range of the suite is 390–310 myr; we would expect a range of ages (which are cooling ages) because the magmatic arc along the subduction zone was active in the same position for long periods between the arrivals of new terranes.

Takaka Terrane rocks in Westland are located in a narrow strip nestled into the curve of the Alpine Fault at Springs Junction, on either side of SH7 where it crosses the Alpine Fault. Coincidentally, the best place to see them here is at the Marble Hill Camping and Picnic Area, also one of the best places to see the trace of recent movements on the fault (see Chapter 19 for directions). Marble Hill, a ridge of Arthur Marble, a Takaka Terrane rock, sits directly adjacent to the fault trace site on the west side. The Anatoki Fault, the major terrane-boundary fault that separates the Buller and Takaka terranes, is mapped as passing right through Springs Junction and meeting the Alpine Fault a bit further south at Crooked Mary Creek.

Fig. 20.3. Coastal exposure of Greenland Group sedimentary rocks (Buller Terrane) at 14 Mile Creek, north of Greymouth. Width of photo 40 cm.
Photographer Bruce Hayward.

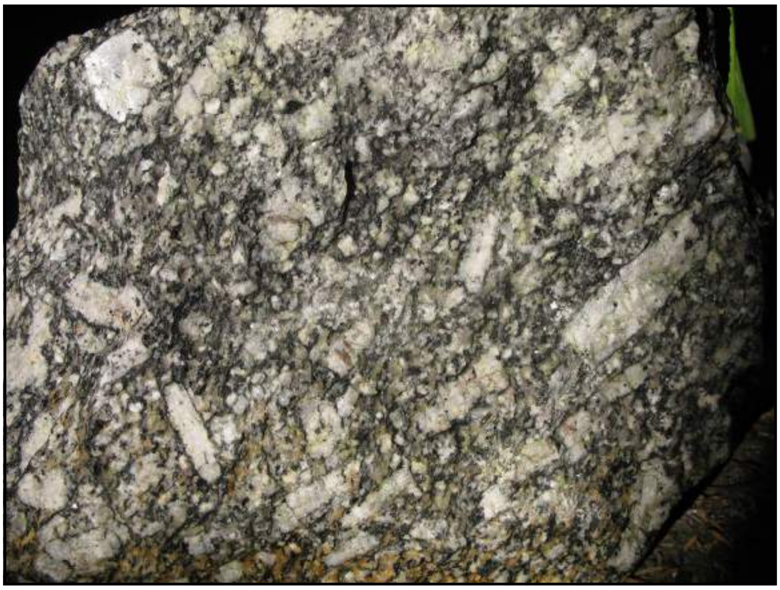


Fig. 20.4. A block of Foulwind Granite (Karamea Granite suite) from Cape Foulwind. Photo 30 cm across. Photographer Andy Tulloch.

Because the Takaka Terrane is right up against the Alpine Fault here, it follows that this is where it gets chopped off by sideways movement on the fault, to reappear on the other side, 480 km away, in Fiordland. There, the rocks are severely metamorphosed, having been brought up from a deeper structural level than the rocks we see at Springs Junction. Takaka Terrane rocks extend in a north–south belt for 300 km from Golden Bay to Springs Junction, although they are replaced by younger granite, and overlain by young cover strata, between Mt Owen in Kahurangi National Park and Springs Junction.

Younger Subduction Zone Granites

Fig. 20.2 shows two granite bodies that are distinctly younger than the Karamea suite. These are Median Batholith diorites dated at 170–135 myr, and Separation Point Granite dated at around 118 myr. The name Median Batholith is used to refer to the assemblage of 'granites' that marks the root zone of the Median magmatic/volcanic arc that post-dates the Karamea Arc (Chapter 16). It is located 20–40 km east of the Karamea Arc, which is as expected, because this is the arc that followed the accretion of the Brook Street Terrane, the basement terrane that followed the Buller and Takaka terranes. In effect, the subduction zone was forced to step backwards by the addition of the terrane.

The Median Batholith can be traced from offshore west of the North Island, through Nelson, the West Coast, Fiordland and Stewart Island/Rakiura, and out to sea onto the Campbell Plateau, at least as far as the Bounty Islands, where the granite is dated at 195 myr. That is a distance of around 2000 km. The oldest known Median Batholith rocks are found in the western Longwood Range of Southland, and are dated at 247–207 myr (Chapter 16), while the youngest member of the subduction zone granites is the Separation Point Granite at 118 myr. This rock is definitely arc-related, but it represents the 'last gasp' of the arc as subduction of the old Phoenix Plate ceased and was finally replaced by a new tectonic regime of spreading (Chapter 4).

There was a 60 myr gap in magmatic activity 310–250 myr ago between the Karamea and Median arcs. There is no known tectonic event that would have switched off the Karamea Arc around 310 myr (Late Carboniferous). The next terrane to arrive, the Brook Street, had not even begun to accumulate. There are, however, various possible reasons why subduction-related magmatism may cease for a time; these are explored further in Chapter 5.

Following the turning off of the Median Arc, the Separation Point Granite became involved in the first steps of New Zealand's next big tectonic phase: the start of a new spreading plate boundary. Along with the Karamea Granite, it played a part in the formation of the Paparoa Metamorphic Core Complex, as described below.

Direct Links with Gondwana

Although all of our older rocks are inferred, on very good grounds, to have been part and parcel of Gondwana prior to our separation around 100 myr ago, there are surprisingly few direct links with the super-continent – i.e. rocks, minerals and fossils that we share with other parts of it, particularly Australia and Antarctica. The fact that we can now date individual grains of the mineral zircon, using the decay of atoms of the uranium that occurs abundantly in zircon crystals, has been a revelation (Chapter 1). Many such grains have now been dated, showing that New Zealand rocks contain zircons dating back as far as 3 byr.

Zircon is first formed as a primary mineral in granite, i.e. it crystallises from the molten magma, complete with its cargo of uranium atoms, which promptly start to decay. Zircon is never a common mineral, but it is one of the toughest substances known to man, and from here on it may follow one of several routes. It can be released from the granite by weathering and erosion of the minerals around it (just like quartz), and begin a career as a sand grain. The sand grain can be transported thousands of kilometres while suffering little wear, be incorporated into a sandstone deposit, be buried and heated, be uncovered by a new cycle of erosion, be transported again, and so on. This cycle can be repeated many times, while the zircon grain is gradually worn from a prismatic crystal with faceted ends to a smoothly rounded, elliptical shape.

During all this time, the radiometric clock keeps ticking. The age of a zircon crystal as ascertained by uranium–lead dating is the age of its original formation – it tells us nothing of where it formed, or where it has been. We can, however, obtain some of that information by using the sister dating technique of fission tracks. Each atomic decay event in the zircon crystal discharges an active particle that causes damage to the zircon crystal lattice – this is called a fission track and it can be seen under the microscope after suitable preparation of the crystals. By comparing the number of fission tracks with the total amount of uranium in the crystal (it varies from crystal to crystal), an age is obtained that can be compared with the uranium–lead age obtained from a mass spectrometer. The two ages are completely independent. The fission tracks are small tubes, which collapse if the zircon grain is heated above 200°C. So, a zircon grain that is heated above 200°C loses its tracks, but they start forming again when it cools below this point. The fission track clock has been reset, and therefore gives the time since the crystal last cooled below 200°C, while the uranium–lead age gives the completely independent age of the original formation of the crystal.

If its host rock is melted, the zircon grain finds itself back in molten magma. The zircon grain does not melt, and as the new magma cools, new layers of zircon form around the old crystal. This can happen more than once, and each new layer contains its own radiometric clock. Thus, a single zircon crystal measuring less than 1 mm can give us several different magmatic ages. These are measured using a Sensitive High Resolution Ion Microprobe (SHRIMP) machine, which records the uranium and lead content in a tiny spot of the crystal using an ion beam.

The point of this discussion is that, while we cannot tie individual zircon grains back to a precise source rock, the fact that New Zealand rocks contain many zircon grains that are much older than their host rocks tells us that we have been around very old continental masses for all our history, until we separated from Gondwana 100 myr ago.

There are other more recent and specific ties to Gondwana. Fossils of the plant *Glossopteris* are very common in other continents that were part of Gondwana during the Permian period (300–250 myr ago). It was always a puzzle that none of these fossils was found in our Permian rocks – and then a leaf was discovered in 1970. Since then, fewer than a dozen specimens have been found, but that is because our Permian rocks are either volcanic or marine in origin, unlike the coal-measure strata in which *Glossopteris* occurs elsewhere.

Terrestrial animal fossils provide another link to Gondwana. In particular, fragments of Cretaceous-age (100 myr-old) dinosaurs have been found in recent years in northern Hawke's Bay and the Chatham Islands. We must still have been attached to Gondwana when they walked here.

Earlier, we discussed the usefulness of younger cover strata in helping to tie down the time at which the various terranes docked. The same cover strata can provide links with Gondwana. Thus, there is a tiny fault-bounded patch of rocks in the hills east of Reefton, 2 sq km in area, dating from 220–210 myr ago (Late Triassic), comprising terrestrial volcanic rocks unlike any other Triassic rocks in New Zealand. What's more, these rocks are intruded by sills of basalt and dolerite (the Kirwans Dolerite) that are around 160 myr old and link directly with the very voluminous Ferrar basalts and dolerites of Antarctica and Australia. This may be only a tiny link with Gondwana, but it's enough. The volcanic rocks are a rare remnant of the volcanic superstructure that accompanied the intrusion of the Median Batholith.

Finally, the Parapara suite of rocks in northwest Nelson provide another link with Gondwana. This is described in Chapter 21.

Paparoa Metamorphic Core Complex

The phenomenon of the metamorphic core complex is described in Chapter 17, with reference to western Fiordland. The Paparoa metamorphic core complex was created at the same time, during the stretching phase 120–90 myr ago (Chapter 4), before New Zealand separated from Gondwana. Box 20.2 summarises the formation of the Paparoa metamorphic core complex, which is a more complete example than the Fiordland one because it retains the upper crustal rocks to either side of the metamorphic core. The most accessible exposures of the core gneisses are found around Constant Bay at Charleston on SH6 near Westport. They were formerly Karamea Granite or Separation Point Granite. During metamorphism in the lower crust, the granites were strongly deformed in the eastern half of the exposed core, but were completely reconstituted into gneiss in the west, e.g. at Constant Bay.

Box 20.2. Paparoa metamorphic core complex.

What is a metamorphic core complex?

It is a place where extreme stretching of continental crust (which is typically 30 to 40 km thick) has allowed rocks of the lower half of the crust to rise in a dome or arch shaped structure, while rocks of the upper half of the crust have slid away from the dome or arch along a major surface of detachment. Another way of describing the detachment is as a low-angle extensional fault zone. In Fig. 20.1 A, the Paparoa core complex comprises everything between (on the map) or below (on the cross section) the detachment surface, the thick red line. The Fiordland core complex comprises all the high-grade metamorphic rocks of the western part of Fiordland — see the map in Fig 20.2. The detachment surface is not clearly defined in Fiordland.

What are the rocks in a core complex? Being from the lower crust, where temperatures and pressures are high, they are typically high-grade metamorphic rocks, or gneiss (pronounced nice) (Boxes 15.3 A, B).

Relationship between gneiss and granite

Boxes 15.3 A, B explain this close relationship. There is a complete transition between granite and gneiss, in both directions. i.e. granite can be deformed to form gneiss, as has happened during formation of the core complexes, or gneiss can be completely melted to form granite, as happened during formation of the Karamea (390-310 myr) and Separation Point (118 myr) granites.

When did the core complexes form?

They were broadly contemporaneous with the Separation Point Granites, around 120 myr to 100 myr. They were the first thing to happen in the brand new tectonic regime of stretching and continental extension which was set in place around 100 myr ago, following a very long period of convergence and subduction at the Gondwana margin (Boxes 5.5 A-C). The Separation Point Granites were one of the last events caused by the subduction, and clearly granite formation lingered and became tangled with the new processes of stretching and arching of the lower crust. See also Fig. 20.2, Box 20.1.

Hawks Crag Breccia (Boxes 20.3 A, B)

At the same time as stretching was allowing core complexes to form at mid-crustal levels, it was also causing the development of basin-and-range fault-block topography at the surface. In the fault-angle depressions, which were above sea-level despite the crustal stretching (because of the associated heating of the crust), there accumulated alluvial fan and debris-flow gravels (Boxes 19.5, 19.6 A, B). Many of the cobbles and boulders were granite. This is the Hawks Crag Breccia, most easily seen at Hawks Crag, an obligatory stop on SH6 in the Lower Buller Gorge.

Also use Boxes 20.3 A, B for outcrops of the core complex rocks between Greymouth, Westport and Inangahua.

Old folds versus young folds

Note in Fig. 20.1 A how the Paparoa core complex is oriented NW to SE, and the stretching direction as indicated by stretching structures in the gneisses and deformed granites was from NE to SW. However, note also that the Paparoa Range and its extensions in old rocks to the north and south form a whale-back arch surrounded by depressions preserving younger sedimentary rocks, all trending from NE to SW, at right angles to the core of the core complex. This fold direction is a much later feature, the folds recording compression from NW to SE, in the last 20 myr, in association with the development of the Alpine Fault (Box 19.3).

The associated stretching in the brittle upper crustal rocks above the rising dome causes them to break up into a basin and range association similar to that seen in Central Otago today (Chapter 18), but with one important difference – the whole regime is extensional, and the range-bounding faults are normal, fairly low-angle, extensional faults (Box 20.3 A). What we have preserved around the Paparoa metamorphic core are the deposits that filled the basins, collectively known as the Hawks Crag Breccia. The angular breccias are old alluvial-fan and debris-flow gravels, many of the stones being granite and Greenland Group quartz sandstones. There were lakes in some of the basins, and old lake deposits are associated with the fan deposits. The formation is named for Hawks Crag on SH6 in the lower Buller Gorge. The breccia is well exposed by the road in and on either side of the crag; use Boxes 19.5 and 19.6 A to interpret what you see. The debris-flow layers are 1–3 m thick and dip southwestward at an angle of 20°.

Extensional basin and range structures were developed around much of what is now New Zealand at this time, because much of the crust was being stretched. Many of these structures are now preserved offshore, deeply buried under younger sediments (e.g. in the Taranaki Basin; Chapter 11), where they play an important role in the hydrocarbon story. Northwest Nelson contains the deposits of one of the basins, the Pakawau Basin, turned inside out by later tectonic activity (Chapter 21), and we shall shortly encounter the same story here on the West Coast.

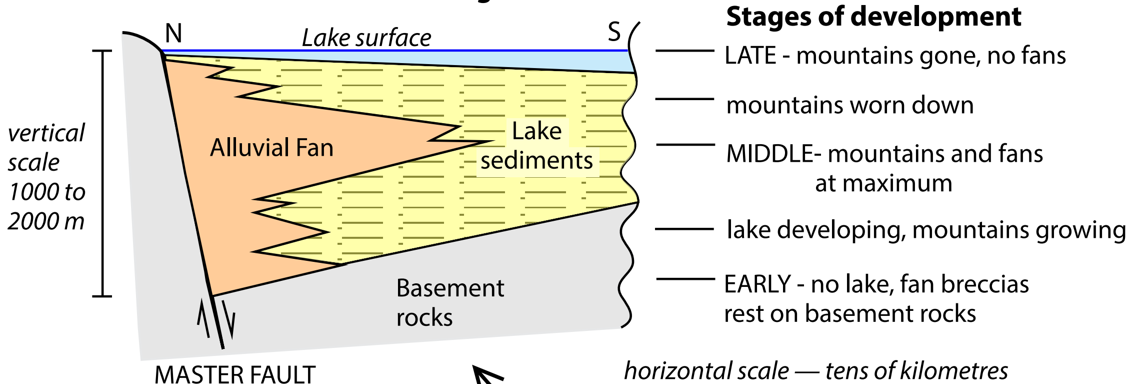
Fig. 20.5. Exposure of 100 myr-old Hawks Crag Breccia in the Fox River, Westland.
Photographer Mo Turnbull, GNS Science.

Box 20.3 A. Hawks Crag Breccia – the stretching of New Zealand.

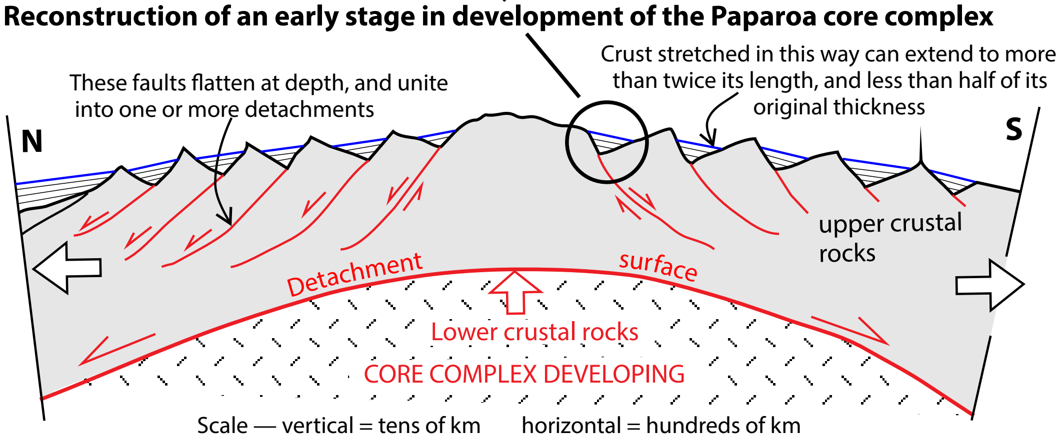
Hawks Crag Breccia (HCB) is found around the Paparoa metamorphic core complex (Box 20.2). It is the surface manifestation of the same crustal stretching process that formed the core complex at mid-crustal levels. The breccia and associated rocks are the product of a basin-and-range fault-block province, similar in scale to the present-day Otago basin-and-range province (Box 18.1) or the Death Valley province in western USA. It was situated above the developing core complex, 100 million years ago. Breccia is an old (fossilized) gravel deposit. It is made of angular fragments of rock bigger than half a centimetre. Maximum size of fragments can be kilometres. Fragments can be touching, or separated in a matrix of sand and mud. Conglomerate is similar, but with rounded fragments. Breccio-conglomerate is half-and-half.

There is an important difference between the HCB scenario and present-day examples such as Otago and western USA — climate. The HCB province developed at high latitude in a wet climate, and the fault-angle basins were occupied by extensive lakes. For this reason, the alluvial fans (Box 19.5) cascading off the growing mountains usually had wet feet — the debris flows carrying the gravels commonly ran into the lake, forming an alluvial fan delta. Debris flows that enter standing water tend to carry on flowing, to drop their gravel load, and to be transformed into another type of sediment flow, the turbidity current. The new turbulent flow carries the sand load further into the lake, eventually depositing a size-graded sand bed, the turbidite — Boxes 6.5 A, B describe turbidites. Beyond, and between, the turbidite sand layers, lake muds were deposited.

Cross-section of a HCB fault-angle basin



Reconstruction of an early stage in development of the Paparoa core complex



(continued in following box)

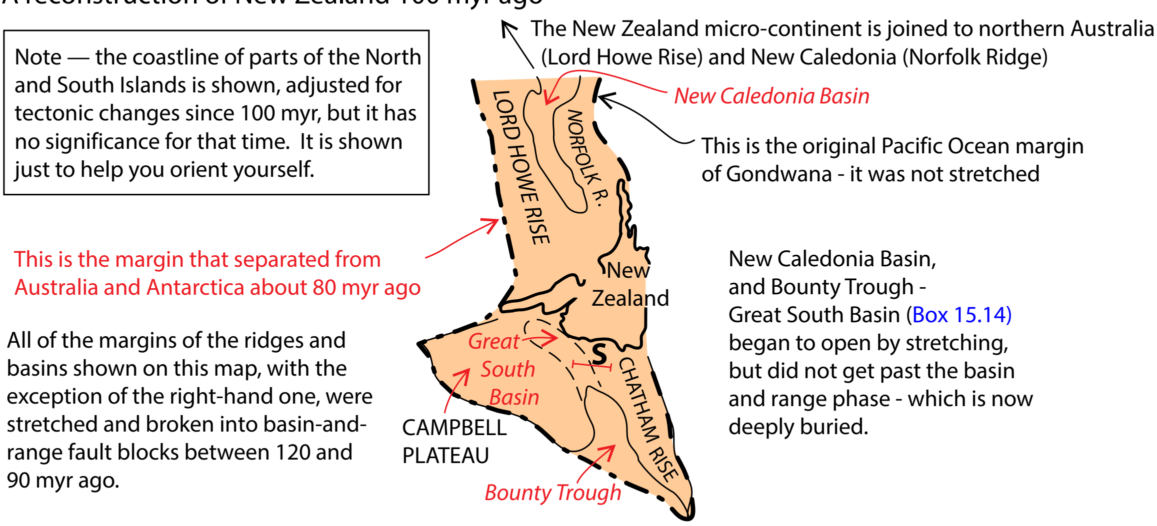
Box 20.3 B. Hawks Crag Breccia – the stretching of New Zealand *(continued).*

Why are there only very limited coal seams in the HCB, when the lake basins were surrounded by forests, and the lake muds contain large quantities of carbon from fossilized plant fragments? The answer lies in the lakes — they were too big and deep for extensive bogs and mires to develop. See Boxes 7.3 A; 20.4 A, B for more on coal.

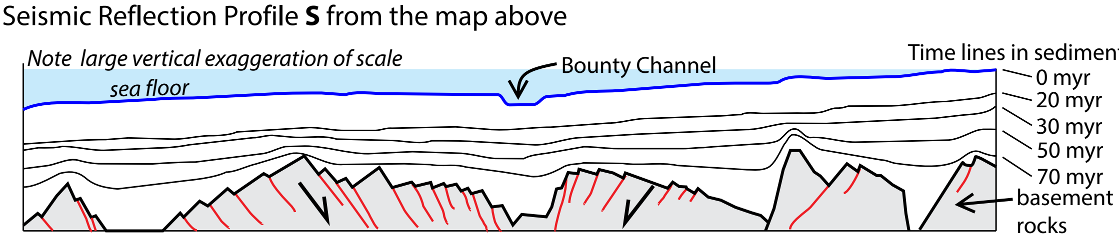
New Zealand at this time — 100 million years ago — was still part of Gondwana. The core complex and the basin-and-range faulting were the first expressions of crustal stretching that lead ultimately (80 to 55 myr ago) to the opening of the Tasman Sea and the separation of the New Zealand micro-continent (Box 5.5 B).

How extensive was the HCB association? On land it is only preserved in a few places — on the West Coast (Fig. 20.1 A) and in a few locations in Otago (Kyeburn and Henley) and southern Fiordland (Fiordland is another core complex — Fig. 20.2, Box 20.1). The reason for this is the recent growth of mountains, and the general uplift of the land area, leading to widespread loss of cover strata to erosion. Offshore, however, it is a different story. The regional subsidence which followed New Zealand's separation from Gondwana resulted in the drowning of most of the microcontinent by the sea by 30 myr ago — in fact, despite the more recent growth of the land area, New Zealand can still claim the dubious distinction of being the world's largest submerged continent. As a result, most of the length of our continental margin — thousands of kilometres — is beneath the sea. Here, seismic reflection profiles show us literally thousands of deeply buried HCB-equivalent fault-angle depressions — see the map and the tracing from such a profile, below.

A reconstruction of New Zealand 100 myr ago



Seismic Reflection Profile **S** from the map above



This profile is an interpretation of the basin-and-range block-faulting of the basement rocks beneath the Bounty Trough—Great South Basin. Note the inwards-facing symmetry of the fault blocks — the centre of the trough lies beneath the present-day Bounty Channel. Clearly, this trough aborted before either ocean crust or a core complex formed.

Younger Cover Strata

In general terms, following the stretching and consequent formation of metamorphic core complexes and basin and range structures over a wide area around the Tasman Sea 120–90 myr ago, spreading finally became concentrated on the Tasman Sea spreading centre, and new oceanic lithosphere began to form there 80 myr ago. Basin and range formation ceased, and the crustal hot rocks began to cool – very slowly, because rock is a very poor conductor of heat. The stage was set for the gradual accumulation of cover strata, as the mini-continent of Zealandia moved steadily away from the Australia/Antarctica Gondwanan remnant. Zealandia was eroded, it cooled and it slowly subsided, as its thinner-than-average continental crust floated lower in the mantle. This scenario applied all over Zealandia, but here on the west side of New Zealand the remnant stretching basins were waiting to be filled up. That is where the cover strata story begins.

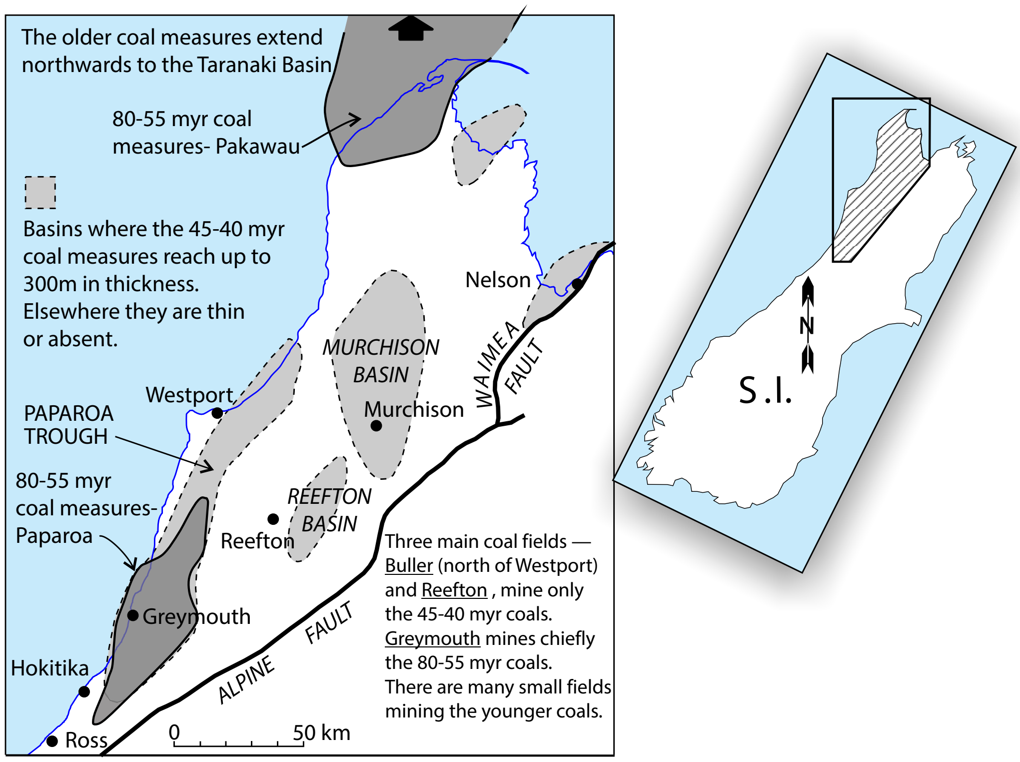
Filling of the stretching basins began in earnest around 80 myr ago, i.e. at the same time as the Tasman Sea started to open up. The climate at this time was an important factor in determining what filled the basins. There were no ice caps. Zealandia was at quite a low latitude (70–80°S), and its climate was correspondingly cool-temperate and moist. Forests flourished, and, with active rivers, swamps and lakes present in the basins, peat accumulated, to be turned into coal following deep burial. Thus, between 80 myr and 55 myr ago – the entire time that the Tasman Sea was opening – the West Coast's older group of coal strata, the Paparoa Measures, were accumulating. As Boxes 20.4 A, B and 20.5 A, B show, there were two stretching basins in the northern West Coast and northwest Nelson region: one around Greymouth; and one in the far north, around the Whanganui Inlet and Cape Farewell (Chapter 21). The latter was actually the southern end of a much larger basin that is now known as the Taranaki Basin (Chapter 11).

The Paparoa Trough and Structural Inversion

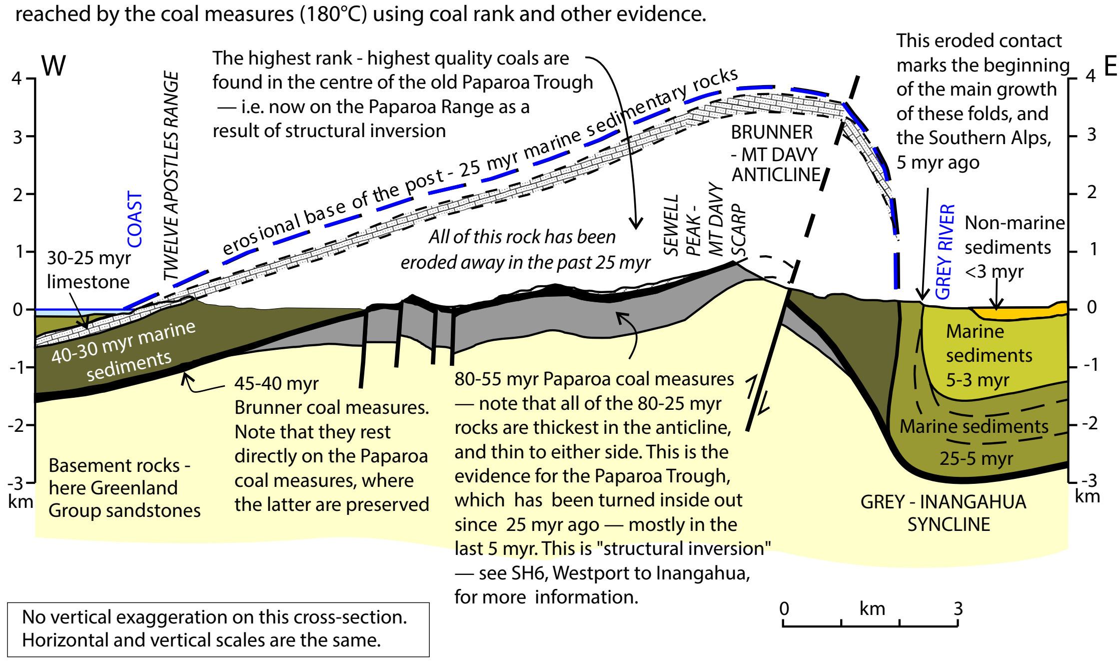
Reconstructing the Paparoa strata around Greymouth from geological mapping has shown that the coal-measure rocks here are folded into a large anticlinal (arch) fold, *and* that the strata are thickest in the central region of the anticline. To understand this, imagine watching a video of the formation rewinding 55 myr: the crest of the anticline sinks before you and turns into a syncline or trough – the Paparoa Trough, or basin, in which the coal-measure strata were deposited, with greatest thickness in the middle.

The fact that the original Paparoa Trough now forms the Brunner–Mt Davy–Paparoa Range Anticline means that it has been turned inside out, or inverted. This is a neat example of the phenomenon of structural inversion, and the cross section in Box 20.4 A shows the huge scale of it – the original Paparoa Trough, containing a thickness of nearly 4 km of strata aged 80–25 myr, has been turned into an anticline-syncline pair with at least 7 km (and possibly as much as 9 km) of vertical relief.

Box 20.4 A. West Coast coal measures.



Note the 7 km of "structural relief". It may be as high as 9 km, according to calculations of maximum temperature reached by the coal measures (180°C) using coal rank and other evidence.



West-to-east cross section located just north of Greymouth, to show the relationship of the two sets of coal measures to each other and to the old Paparoa Trough, and the structural inversion of the trough to form the Brunner - Mt Davy anticline/Paparoa Range.

Box 20.4 B. West Coast coal measures discussion.

General aspects of coal-bearing rocks ("coal measures") are covered in Box 7.3 A. Coal begins life as peat, formed in swamps and mires, and is converted to coal by deep burial underneath younger sediments. Coal is always associated with gravels, sands and muds that represent the deposits of beaches, dunes, rivers, streams, bogs or lakes. In some cases the strata are organised into regular cycles of upwards fining, formed by the systematic meandering of river channels back and forth across river flood plains.

Two sets of coals

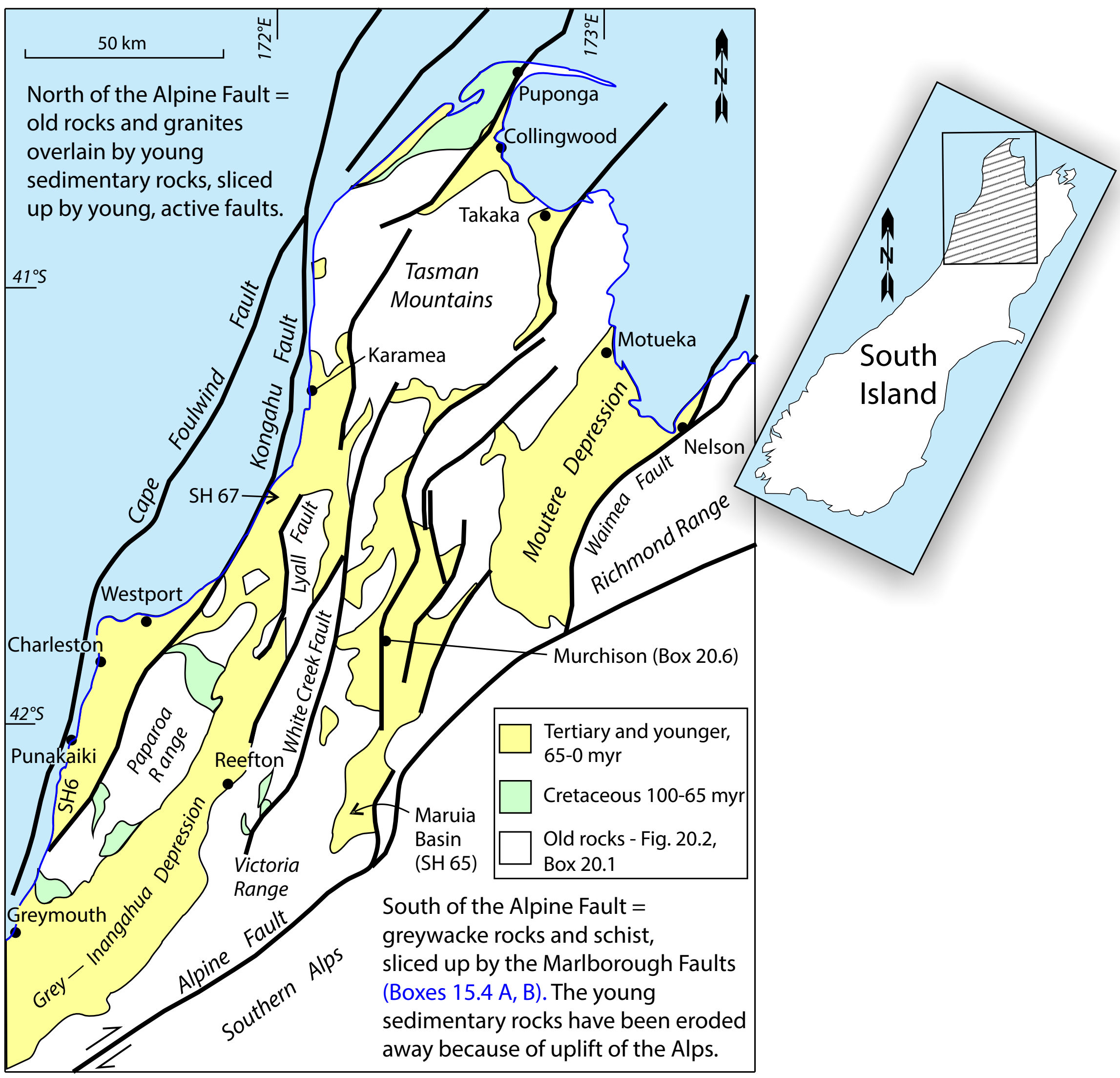
The West Coast differs from the rest of the country by having two quite distinct sets of coal measures. The younger set (45-40 myrs old) is shared with much of the rest of the country, but the older set (80-55 myr) is unique to the west side of both islands. In the South Island older coals are present in two areas, Greymouth and the northern tip of the island. The latter set of coal measures extends northward beneath the Taranaki Basin and alongside the North island at least as far as Auckland, but beneath the sea and deeply buried. They are important there because of their role as source rocks for natural gas in the Taranaki Basin. Greymouth coalfield mines chiefly the older set of coals, and there are a number of different seams mined. They vary a great deal from place to place (reflecting the complex geography of the swamp and lake systems).

In the Greymouth coalfield (see the cross section in Box 20.4 A) the younger coals rest directly on top of the older ones, and were mined in the Dobson and Blackball mines. The other two principal coalfields, Buller (north of Westport) and Reefton, mine only the younger coals, and there are numerous minor fields, past and present, mining the younger coal.

Why are the 80-55 myr coals confined to the west of the country?

The answer lies in the break-up of Gondwana and the opening of the Tasman Sea between 80-55 myr ago. During the crustal stretching process that preceded the appearance of oceanic crust in the Tasman, between 100 and 80 million years ago, (seafloor spreading — Box 5.5 A) several phenomena resulted (Boxes 20.2; 20.3 A, B; 20.5 A, B), including the production of rift valley systems akin to the present-day East African Rift valleys. It was one of these that allowed the formation of the 80-55 myr coal measures, and then further allowed their preservation because that particular rift aborted and did not develop into an ocean basin. The same situation also occurs on the opposite side of the Alpine Fault in Southland, at the Ohai and Nightcaps coalfield (Chapter 16). On the reconstruction of New Zealand for 80 myr ago, this area was well removed from the West Coast (Boxes 5.5 B, C), and therefore may be indicating yet another failed rift valley.

Box 20.5 A. Younger rocks of Westland – Nelson.



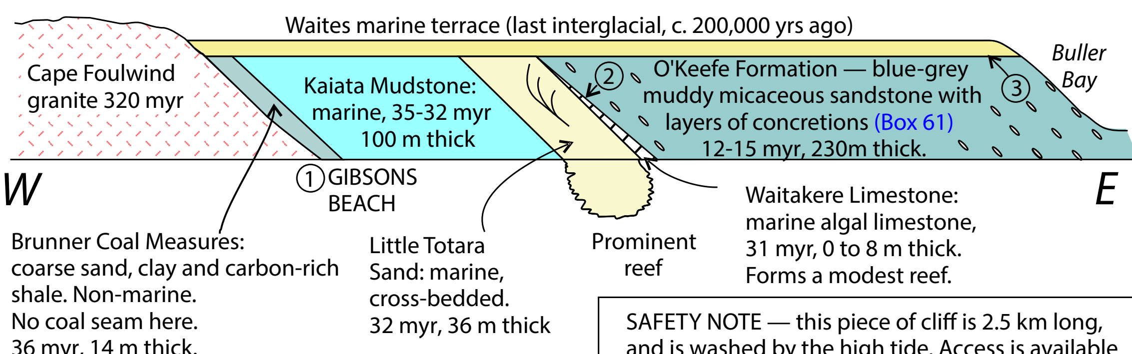
Young sedimentary rocks (cover strata less than 100 million years old) were laid down over the whole region. Their general nature is a perfect mirror-image of the same-aged rocks in eastern South Island — see Box 15.9 A. They lack only the basaltic volcanic rocks. The rocks have been preserved from erosion, since the uplift of western South Island above sea level about 5 million years ago, only where faults and folds have left them in depressions — either between faults or in the fault-angles adjacent to individual faults. The faults on the map are part of the Alpine Fault system — note how they form an angle of about 30° anticlockwise from the Alpine Fault itself. That is a typical pattern of subsidiary faults in relation to a master fault (the Alpine Fault) in a right-handed (dextral) sideways-moving fault system. Some of these faults were active 20 million years ago, some are still active. See Boxes 20.4 A, B for the West Coast coal-bearing rocks, Boxes 20.3 A, B for the Hawks Crag Breccia, Box 20.6 for the Murchison and Maruia Basins, and Figs 20.1 A, B; Box 20.1 for the older rocks of Westland-NW Nelson.

Box 20.5 B. Well-exposed cliff section of younger strata west of Westport, Cape Foulwind to Buller Bay.

In this cliff exposure there are three unconformities — contacts representing time gaps when erosion occurred (Box 4.1)

1. Coal measures rest on an eroded and weathered surface of granite (time gap 320 to 36 myr).
2. O'Keefe Formation on Waitakere Limestone — note the slight difference in angle of dip and the absence of the limestone at the top of the cliff. This is the same limestone that forms Punakaiki Pancake Rocks 40 km to the south — there the erosion forming this unconformity did not extend so deeply into the older sequence. This unconformity records the New Zealand-wide change in tectonic environment at 25 myr ago — Box 5.5 C. Here the time gap in the rock sequence is 31 to 15 myr.
3. Tilting of the strata probably began 5 myr ago, when compression began to raise the Southern Alps. Erosion of the Waites terrace and deposition of 3 m of coastal sand took place at sea-level around 200,000 yr ago. The terrace has been uplifted by 18 to 38 m.

Note the vertical exaggeration in this cliff diagram.
The actual dip of the strata is 10° to the SE.



SAFETY NOTE — this piece of cliff is 2.5 km long, and is washed by the high tide. Access is available only at the two ends. Do not get trapped by the incoming tide.

FOSSILS —

1. Shells are sparse but widely distributed in the Kaiata and O'Keefe Formations.
2. Marine microfossils (mostly Foraminifera) are abundant in the same two formations (strong lens or microscope needed)
3. Trace fossils — burrows in the sedimentary rocks — are abundant (Box 6.5 B for examples)

The trough is now the anticline, and the adjacent structural high, on which hardly any sediment accumulated between 80 myr and 25 myr, now forms the Grey–Inangahua Syncline, with its base 3 km below sea-level and a 3 km-thick accumulation of sediment younger than 25 myr. If you compare the thickness of sediment in the Grey–Inangahua Syncline aged 25–5 myr with the strata accumulated over the last 5 myr, you will note that the rate of accumulation of sediment – and by extension the rate of formation of the syncline – increased markedly at 5 myr. As we have seen elsewhere, this was the time at which the tectonic tempo of the whole South Island stepped up by several notches. Most of the growth of the older anticline-syncline pair has occurred in the past 5 myr. Note also that the syncline was an arm of the sea until 3 myr ago, when the inpouring of sediment eroded from the rising Victoria Range and Southern Alps to the east filled it up.

Structural inversion comes about when extensional fault-bounded basins are subjected at a later date to a completely new tectonic regime – compression. The original normal faults are reactivated as reverse faults, retaining the direction of dip of the original, but moving rocks in the opposite direction. Consequently, what had been thick sediment in a trough becomes an arch at a higher level.

The Paparoa Trough is far from being the only example of structural inversion in New Zealand. Cross sections of the Taranaki Basin in Chapter 11 (Boxes 11.1 A, B) show buried examples that have been reconstructed in three dimensions during oil exploration. The Cape Farewell Basin has also been inverted. Closer to hand, the Victoria Range to the east of the Paparoa Range was at one time the site of a basin in which Paparoa Coal Measures were laid down, remnants of which (not mined) are preserved around Reefton today. The basin was inverted to form the Victoria Range sometime after 25 myr ago. All of this followed the change from extensional to compressional tectonics as predictably as night follows day.

Inversion of relief

The Grey–Inangahua Syncline is synonymous with the Greymouth–Reefton–Inangahua depression, which contains the Grey, Mawheraiti and Inangahua rivers, and is followed by SH69 and SH7. From the road, the dominant feature is the prevalence of river terraces, as well as river gravels exposed in cliffs up to 100 m high. Continuing compression from the growth of the Southern Alps is causing some uplift of the syncline. The ranges to either side of the Greymouth–Inangahua Syncline depression – Victoria to the east and Paparoa to the west – are anticlinal in origin.

It is appropriate here to comment on another form of inversion, namely inversion of relief. The similarity of the term to structural inversion raises the possibility of confusion. It is commonly the case that long-term erosion of folded strata results in rivers following anticlines, and synclines forming ridges. This is called inversion of relief, i.e. the relief does not follow the geological folds. It comes about because strata are stretched around the crest of anticlines, causing defects like joint cracks to open up and let in the agents of weathering and erosion. In synclines, by contrast, defects are squeezed shut, and the rocks are more resistant to weathering and erosion.

In the case of the Grey–Inangahua Syncline, the structures are still actively forming and the relief follows the folds. Inversion of relief has not had time to take effect here, and tectonic movements and sediment deposition are outpacing erosion.

Economic consequences of structural inversion

The rank of coal (it ranges from peat through lignite and bituminous coal to anthracite) is determined by the cooking that accompanies burial. The deeper the burial, normally, the higher the rank and the more valuable the coal (although note that this is a bald summary of what is actually a complex topic). Referring to Box 20.4 A, we can see that the Paparoa Coal Measures in the Greymouth Coalfield were buried to at least 3 km, and as a result they are high-rank bituminous coals that are in demand for industrial use.

If the Greymouth Coalfield rocks had stayed buried 3 km down, mining them would have been next to impossible. So, structural inversion has placed them within reach of miners, although it has also created its own problems. Mining on the summits of mountains is difficult, to say the least. Attempts were made to mine at Mt Davy on the crest of the Paparoa Range near Greymouth, but had to be abandoned. *Editors' note: This inversion caused additional structural complexity that made mining more difficult, as witnessed by the Pike River Mine disaster (2010), which was attempting to mine a sliver of Paparoa Coal Measures in a fault-angle depression on top of the Paparoa Range*.

The best place to see Paparoa Coal Measures from the road is along SH6 north of Greymouth. Here, a coarse, conglomeratic phase of river and alluvial fan deposits, without coal seams, make up Nine Mile Buff. Impressive exposures of these rocks continue until Twelve Mile Bluff. Look out for the sedimentary structure cross bedding. Layers of sandstone 1–3 m thick show thin internal layers dipping at around 30° to the main stratification. Cross bedding results from the migration of dune and delta structures.

Younger Coal Measures and Associated Strata

The strata that overlie the Paparoa Coal Measures directly, or spread beyond them (i.e. beyond the original stretching basin) to rest on basement rocks, are coal measures that always rest on a deeply weathered surface of older rocks (Box 20.4 A). These are the Brunner Coal Measures, which are around 40 myr old. There was a 10 myr period of weathering and erosion separating the deposition of the two coal-measure formations. As elsewhere in the country (Chapter 7), the Brunner Coal Measures record the formation of extensive coastal swamps that preceded invasion by the sea as low-lying Zealandia slowly subsided during phase three of New Zealand's development. As things have worked out, the mines around Greymouth target mostly the older Paparoa coals, while the mines of Reefton and the Westport district (where a visit to the old mining town of Denniston is a must) target the younger Brunner coals. Like coal measures everywhere, there is variation in the development of coal seams from place to place, depending on the distribution of peat swamps and rivers at the time of deposition.

The Brunner Coal Measures coastal swamps retreated inland as the sea encroached, and shallow marine sands and muds accumulated on top of the coal. Finally, as the land was eroded progressively lower, the supply of terrestrial sand dwindled and, instead, gravel and sand made of shells and shell fragments covered the broad, shallow continental shelf. The 30 myr-old limestone that formed as a result of the burial and cementation of these deposits is well developed around the Brunner-Mt Davy-Paparoa Range Anticline, and gives rise to the spectacular landscape that inspired the establishment of the Paparoa National Park.

Fig. 20.6. Flaggy limestone forms Pancake Rocks at Punakaiki, Westland. Photographer Bruce Hayward.

The limestone is well displayed at the coast in the Pancake Rocks at Punakaiki. The formation dips gently, and has a marked 'flagginess' (the 'pancakes'), created by the division of the rock into alternate hard and soft layers a few centimetres thick that are picked out by wave and spray erosion (Fig. 20.6). The rocks also have prominent near-vertical joints, and erosion along these has formed pillars and 'castles' of spectacularly sculpted rock, as well as impressive blowholes.

The limestone rises inland, over the Punakaiki Anticline, underneath the Barrytown Syncline and finally onto the anticline of the Paparoa Range (Box 20.4 A). It has been eroded off the crest of the Paparoa Range, uncovering the gneisses of the Paparoa metamorphic core complex (Fig. 20.1 A).

Landforms on the Quartzose Coal Measures

The Brunner Coal Measures are characterised by sandstones that are rich in quartz. The quartz was the residue from tens of millions of years of erosion and chemical weathering of older rocks, during which most of the common rock-forming minerals (e.g. feldspar and mica) were changed to clay (Box 6.1 B). As a result of their quartz content, the rocks have been called the Quartzose Coal Measures. Quartz accumulates during weathering and erosion because, unlike most other minerals, it is very tough, is unaffected by chemical weathering and is virtually insoluble in water. Those same properties mean that soils forming on quartz-rich rocks have virtually no elements to make use of, and so are thin and infertile. The combination of infertility and resistance to erosion gives rise to distinctive plateau landforms where the strata are dipping at low angles. The high, treeless plateau at Denniston is a good example – another reason for visiting the extraordinary old mining town.

Cover Strata of the Last 25 Million Years

The tectonic turn-around at 25 myr caused initial uplift in some places along the West Coast, with the result that the 30 myr limestone layer was partly or wholly removed. The marine deposit most typical of the past 25 myr is a blue-grey muddy micaceous sandstone containing layers of calcareous concretions (Boxes 15.11 A, B). It occurs widely around the West Coast region, and was known as the 'Blue Bottom' to the alluvial gold-miners. In places it contains developments of flysch, alternating sandstone–mudstone sequences (Box 6.5 A), e.g. as found in the Murchison Basin (see below).

A well-exposed cliff section showing nearly all the cover strata can be seen to the west of Westport. It runs from the Cape Foulwind Granite eastwards to Buller Bay, and is shown in Box 20.5 B. A notable feature of the section is that the 30 myr-old limestone layer is thin (less than 10 m), owing to the localised erosion discussed previously that preceded deposition of the post-25 myr sediments; this represents a hiccup accompanying the tectonic turn-around at 25 myr. However, just 2 km to the southwest, the limestone was thick enough to quarry for feedstock for the nearby cement works, which is now closed.

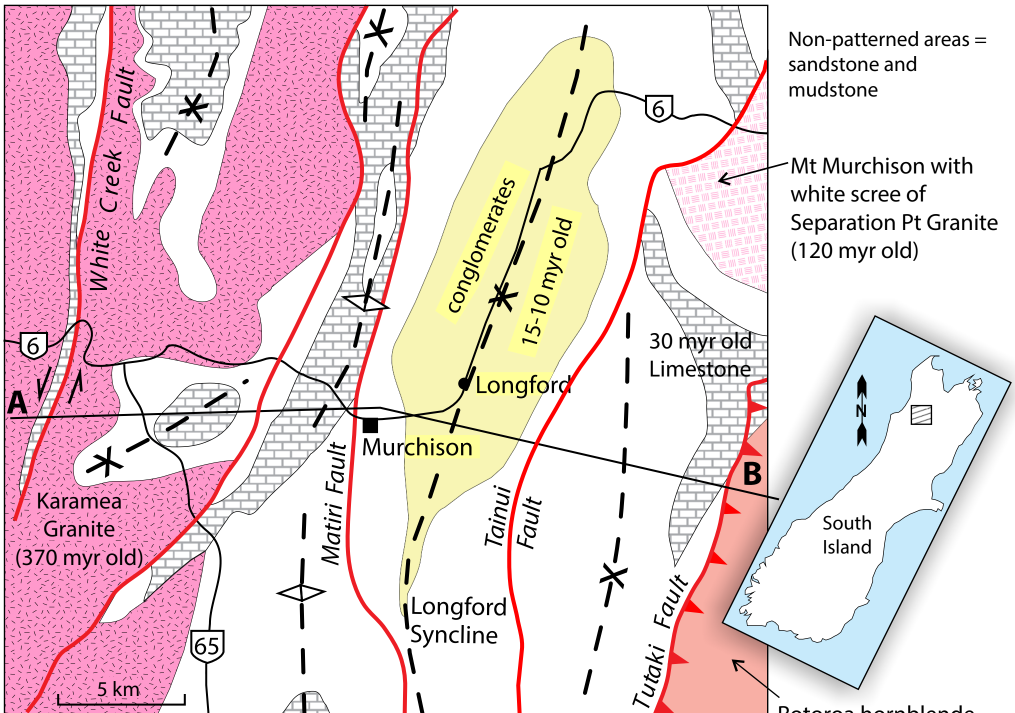
Effects of the Accelerated Uplift on the Alpine Fault

As with the South Island east of the Alpine Fault, so the West Coast has been strongly affected by compression across the fault in the last 5 myr. Most of the folding and faulting shown in Boxes 20.4 A, B and 20.5 A, B has taken place during that time. However, while compression is pushing up the Southern Alps (Chapter 19), the West Coast is acting as the backstop. The southern, narrow section of the West Coast is part of a southwards-tapering sliver of continental crust that effectively ends at Milford Sound. The northern, wider part, however, is backed up by the Challenger Plateau (Fig. 1.3), part of a long strip of continental crust, the Lord Howe Rise, that extends to Queensland.

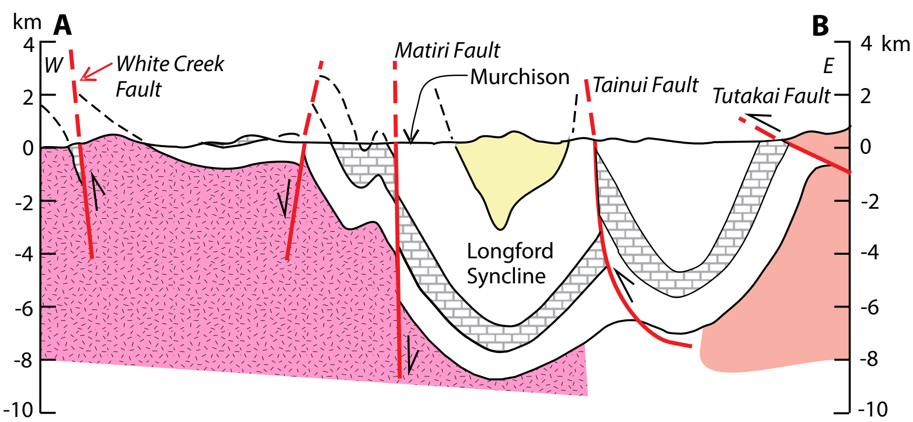
The Murchison and Maruia basins

We've dealt with the remarkable Brunner–Mt Davy–Paparoa Range Anticline and Grey–Inangahua Syncline pair, which has more than 7 km of vertical relief, most of it formed within the past 5 myr. There is an equally remarkable pair of synclinal basins at Murchison and Maruia (Box 20.6), which began life as subsiding sedimentary basins following the 25 myr tectonic turn-around. The earlier sequence of cover strata, relating to the subsidence of Zealandia, is preserved across the area, with limestone deposited in shallow seas about 30 myr ago. After 25 myr, however, rapid subsidence of the Murchison Basin began, centred just east of Murchison town. Initial deposits were mudstones and sandstones, but, as the tectonic tempo picked up, conglomerates were deposited between 15 myr and 10 myr ago. The huge volume of sediment held in the basin tells us that there was comparable uplift and erosion in the surrounding areas, and that the relief of these areas increased in order to provide the later gravels.

Box 20.6. Murchison Basin.



To Springs Junction, through the Maruia Basin - see Box 20.5 A for its full extent. The geology is similar to the Murchison basin, but access is difficult.



West-to-east cross section through the Murchison Basin. There is no vertical exaggeration on this cross section. All the dip angles and depths are genuine. This is a remarkably deep structure. Note that the older sedimentary layers (more than 25 myr old) are much thinner than the younger layers.

The great thickness of the latter was determined by the fact that the Basin was deepening simultaneously with the deposition of the sediments; and since sediment has to have a source it also tells us that surrounding areas were being uplifted at the same time to provide the sediment. The unlabelled areas of rock on the map and cross section are sandstones and mudstones. Most of the faults (not all are shown — this is a simplified map and cross section) have a component of east-west shortening in their movement, i.e. they are reverse and thrust faults (Box 6.6C). Being associated with the Alpine Fault (it is just to the east of this map) they also have a component of sideways motion, mostly right-handed but left-handed in the case of the movement on the White Creek Fault during the 1929 Murchison Earthquake, when the eastern side of the fault rose 5 metres where it crosses the road. That is, the faults are truly oblique-slip faults — they have all possible components of motion. (See Box 20.5 A for the bigger picture of West Coast faults).

The Maruia Basin lies to the south. Highway 65, between SH 6 and Springs Junction, passes through the western side of it. It is very similar in style and history to the Murchison Basin, but access is not easy. It is also closer to the Alpine Fault, and its southeastern portion has been displaced by the Alpine Fault and lost.

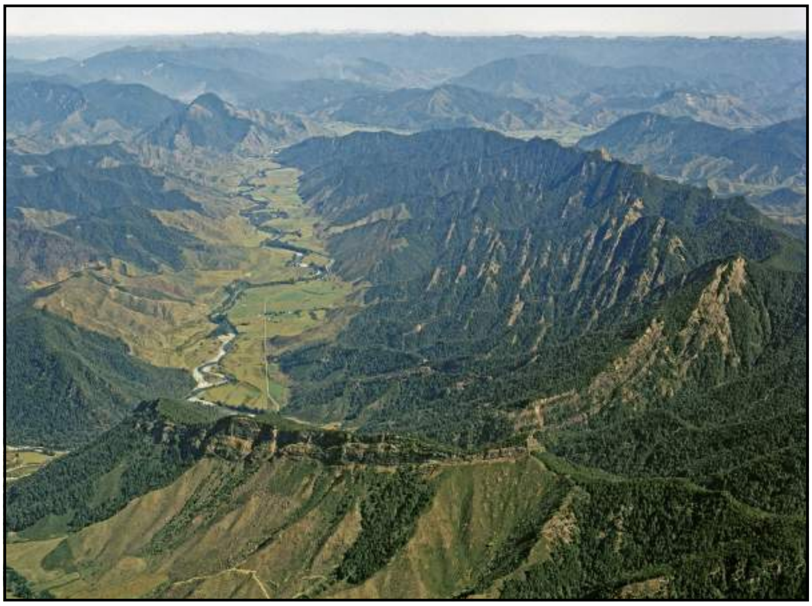


Fig. 20.7. Part of the Buller River runs down the eroded exis of the Longford Anticline near Murchison. The syncline was formed by strong tectonic compression during the last 5 myrs. Photographer Lloyd Homer, GNS Science.

The original Murchison Basin was considerable wider than the present-day syncline (known as the Longford Syncline), because the area has been strongly compressed during the past 5 myr. The Longford Syncline today is followed by SH6 and the Buller River for nearly 20 km north of Longford. The two limbs of the syncline dip inwards at very steep angles, and the conglomerate beds form conspicuous ridges that can clearly be seen around Murchison. This is a remarkably deep syncline. Its bottom is at least 8 km below the present land surface (Box 20.6), and it could be even deeper – there is no drill-hole information, and there is considerable uncertainty in projecting dipping strata to these sorts of depths. A companion syncline, nearly as deep, lies just to the east, although the intervening anticline has been sheared out by reverse movement on the Tainui Fault.

The Maruia sedimentary basin was located to the south of the Murchison Basin. It is very similar in style and history to the Murchison, but is younger, largely inaccessible, and its southeastern portion has been taken away by the Alpine Fault, uplifted and lost. However, the source of many of the stones deposited in the basin has been pinpointed far to the southwest on the opposite side of the fault, and the basin is an important piece of evidence pointing to a young age for a considerable part of the sideways displacement on the Alpine Fault.

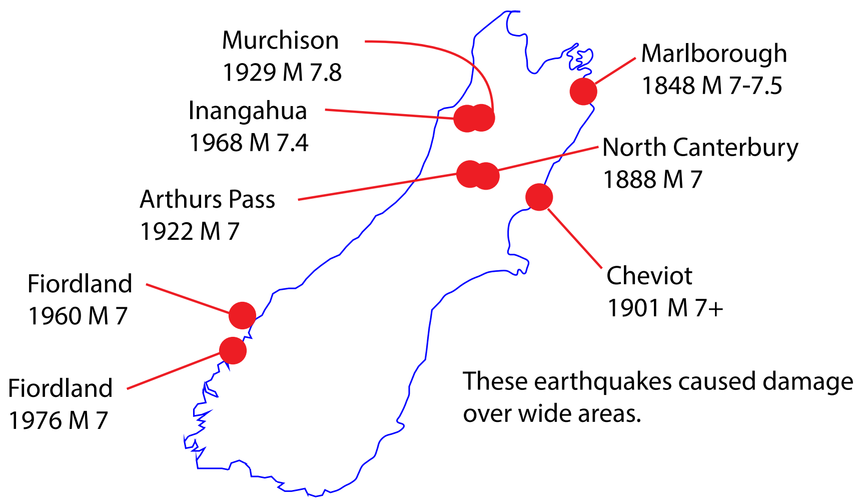
The Murchison Basin has attracted considerable interest for oil exploration. There are oil and gas seeps throughout the area, including at least one permanently burning gas seep. Several exploration wells have been drilled, but without success – in deep and steeply dipping strata like these, it is very difficult to pinpoint and drill an oil and gas reservoir.

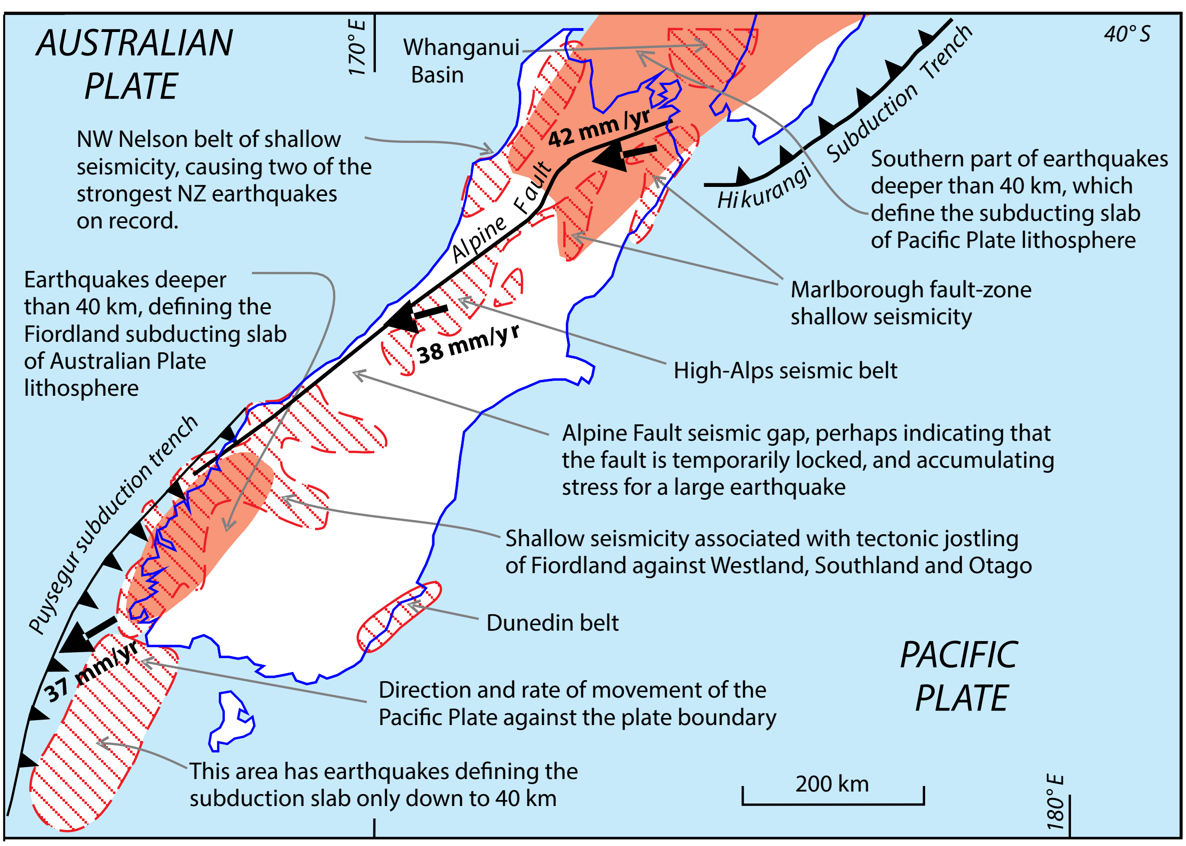
Earthquakes

Editors' note: Peter Ballance wrote this book before the recent spate of devastating earthquakes and fault ruptures (Darfield 2010, Christchurch 2011, Seddon 2013, Kaikoura 2016) that struck the northeastern South Island. If he was still with us we are sure he would have updated this section and added descriptions of these earthquakes and their implications.

Box 20.7. South Island earthquakes (pre-2000).

Destructive earthquakes, magnitude 7 or greater, 1840 - 1999





Map of seismic events shallower than 15 km for the period 1990-1993. There are many isolated small events outside the marked areas.

While there hasn't been an earthquake on the Alpine Fault itself in recorded time (though it's fair to say that a big one is due), other faults shown in Box 20.5 A have moved in association with large earthquakes. The 1929 Murchison, or Buller, earthquake had a magnitude of 7.8 (Box 20.7), the third largest on record in New Zealand, the two larger being the 1931 Napier earthquake (magnitude 7.9; Chapter 12) and the 1855 Wairarapa earthquake (magnitude 8-plus; Chapter 13). During the Murchison earthquake the White Creek Fault ruptured. The road (now SH6) was displaced vertically by about 5 m up to the east, and sideways sinistrally (to the left) by 2.5 m. The fault movement was thus an oblique reverse movement – i.e. part sideways and part compressional. There was no visible fault escarpment prior to the earthquake, indicating that the fault had not moved for a long time – at least 18,000 years. However, the fault was not new – it has a deep fault-angle depression that preserves steeply dipping cover strata, as shown in the cross section in Box 20.6 (this explains why there is a lime quarry a short distance west of the fault rupture).

The effects of the highly damaging earthquake were spread over a large area (the tower of Nelson College collapsed, for example). Explanatory signs of the effects of the earthquake can be seen on SH6 at White Creek. The Maruia Falls, developed on the 30 myr limestone and now seen by SH65, had been buried by a landslide at some earlier date. However, renewed movement on the landslide in 1929 pushed the river channel to one side and re-exposed the falls.

A more recent earthquake, in 1968 at nearby Inangahua, had a magnitude of 7.1. It was shallow and was also felt over a wide area, although it caused less damage than the Murchison earthquake.

The sedimentary record of the past 5 million years

With the Southern Alps rocketing upwards a short distance to the east, it is to be expected that we will find a record of that uplift in gravels on the West Coast. And given the variety of rocks on the other side of the fault, those gravels might contain a record of how and when these rocks moved past given points on the West Coast. The gravels found are known as the Old Man Group, and they contain mostly greywacke and schist pebbles from east of the fault. The oldest gravels contain only greywacke pebbles, indicating that uplift of the Southern Alps had not yet brought schist to the surface. Younger gravels contain schist pebbles that include high-grade varieties, indicating that the present-day pattern, in which the deepest, highest-grade schists are pushed up the actual fault plane, was established part of the way through the history of the Southern Alps.

Global climate moved progressively into glacial mode following the start of mountain-building in the Southern Alps 5 myr ago, and hence there was a corresponding development of glaciation in the Alps (see below). Glacially influenced sediments thus play an increasingly important role on the West Coast at this time.

Glaciation on the West Coast

The general features of glaciation are covered in Chapter 19, but there is a fuller record of New Zealand's glacial history on the West Coast than anywhere else. This is because the region is unusually responsive to small climate changes during glaciation, by virtue of its extreme wetness.

The two existing glaciers, Fox and Franz Josef, fluctuate on a timescale of decades, in response mainly to short climatic cycles that vary the quantity of snow accumulating on the high névé. During colder ice ages, many more glaciers descend the steep western face of the Southern Alps and extend across the full width of the coastal plain, especially in the south where the plain is narrow and precipitation greater. In some cases, they extended beyond the present coastline, because of course sea-level is lower during glaciations. Any fjords eroded by these glaciers were small and have since been filled by the voluminous sediment output from the mountains.

Thus the coastal strip from the Jackson Bay/Haast area northwards to Greymouth is a patchwork of curvilinear glacial moraines and associated moraine-dammed lakes. Most valleys are flanked by two ridges of lateral moraine, which in many cases are perched high and dry on bedrock, as elongate mounds of rubble, parallel to the rivers. Both the north and south heads of Bruce Bay, for example, are moraines. As glaciers retreated into the mountains 20,000–10,000 years ago, pauses and slight readvances created horseshoe-shaped terminal moraines, convex to the sea. These moraine ridges are now vegetated, like all of the West Coast, and consequently are harder to spot than the bare moraines seen around the glaciers in Aoraki/Mount Cook National Park. Look for them in all the valleys crossed by SH6, or take to the air, as they are easier to see from above (Fig. 20.8).

As with the eastern side of the Southern Alps, rapid uplift (as much as 1 km/glacial cycle), combined with the propensity of glaciers to bulldoze away all previous glacial deposits, means that records of earlier glaciations (they occur every 100,000 years) are hard to find. Such records are best preserved on the wider coastal area behind Greymouth, where deposits of the last four glaciations have been identified, dating from around 30,000, 130,000, 250,000 and 360,000 years ago.

No trip to the West Coast is complete without a visit to either the Fox Glacier or the Franz Josef Glacier, two of New Zealand's most spectacular natural features. As noted in Chapter 19, these two glaciers are well fed, steep, fast moving (several metres per day) and highly responsive to decade-long fluctuations in snowfall on the névé that feeds them (Fig. 20.9). The rapid flow carries the ice down well below the bush-line before it melts.

Fig. 20.8. The Waiho Loop, near SH6, is the terminal moraine of a major advance of the Franz Josef Glacier about 12,000 yrs ago. Photographer Lloyd Homer, GNS Science.

The two glaciers fluctuated markedly in length through the late nineteenth and twentieth centuries, and since 1900 both have retreated by more than 4 km. This overall retreat has been punctuated by periodic advances – e.g. the Franz Josef advanced between 1915 and 1930 (a small distance), between 1943 and 1950 (200 m), between 1965 and 1989 (400 m), and between 1983 and 1990 (600 m).

Bearing in mind that this is a dynamic and dangerous environment in which unwary tourists have been killed, things to look for when visiting the glaciers include:

- The outflow tunnel from the glacier, and the frequent collapse of ice blocks into the meltwater river emerging from the tunnel (Fig. 20.10).
- The bouldery nature of sediment produced by the glacier.
- The rapid development of an incipient terminal moraine at each successive ice terminus.
- Blocks of stagnant ice left behind in moraine accumulations, which leave 'kettle holes' full of water when they melt.
- Ice-smoothed and ice-plucked rock hills, called roches moutonnées, meaning 'rock sheep'.
- Scratchings and horizontal gouges in the rock sides of the valley, recording ice erosion at higher ice levels (Fig. 20.11).

Note also that rapid changes around the glacier snouts mean that access roads and paths are continually being modified.

Both glacial valleys have ice-polished surfaces that are good places to see alpine schist, this being east of the Alpine Fault. The schistosity layering (Box 15.3 A) is very steep to vertical, because of the way the Southern Alps are growing by near-vertical upthrusting of the schists against the Alpine Fault (Chapter 19). Vertical schistosity banding bends downslope as it nears the hill surface – this is called downhill rock-creep.

On the access roads to and from the glaciers you cross the Alpine Fault. In this dynamic environment, evidence of the fault is hard to find, because it moves only every few hundred years and hence fault traces are soon eroded or covered.

Fig. 20.9. The ice field névé feeds the Fox Glacier on left. Aoraki/Mt Cook is on right. Photographer Lloyd Homer, GNS Science.

Fig. 20.10. Meltwater from within Franz Josef Glacier disgorges from the outflow tunnel in the glacier terminus in 2014. Photographer Bruce Hayward.

Fig. 20.11. These horizontal gouges on the 300-m-high walls of Cone Rock on the side of Fox River were created by rocks being dragged along by the Fox Glacier when it was much bigger than it is today. Photographer Bruce Hayward.

West Coast pakihi plains

Pakihi plains (pakihi is the Maori word for open grassland or barren land) and terraces are typical of much of the West Coast. They consist of poorly drained glacial and river sediments in which a sub-surface iron-pan deposit has formed. The iron pan is impervious, so the soils are wet and infertile. A lot of work is required to break up the iron pan and make the soils usable for agriculture – it has been done on a large, farm-size scale, by turning over of the top 2 m of the ground and bringing the underlying sand to the surface.

West Coast gold

All the basement terrane rocks of the West Coast and northwest Nelson contain large numbers of gold-bearing quartz veins, on account of their history of folding, metamorphism and intrusion by several generations of granite. Most of these veins are found in the Greenland Group of sedimentary rocks, and there have been many attempts to mine them, e.g. around Ross, around the Paparoa Range, Buller-Mokihinui, Lyell and, especially, Reefton. At the time of writing, a new bulk-rock mining operation was underway from a large open-cast pit at Reefton. This is a similar operation to the active gold mine at Macraes Flat in North Otago (Chapter 18), recovering the gold contained in numerous small quartz veins, and the same company runs both operations.

Alluvial gold is widespread throughout the West Coast, and like all South Island alluvial gold it has been eroded from the quartz veins of basement rocks. Recent geological history of the West Coast has favoured the accumulation of heavy gold grains in two locations: one in old beach sands related to earlier interglacial high sea-levels, now uplifted; and one in riffle-like situations in gravel stream beds just downstream of the fronts – and hence terminal moraines – of glaciers.

Nineteenth-century gold-miners recognised these two situations, which they called leads, and mined them accordingly. Old beaches are long and string-like, and were worked by digging a long line of shafts and gullies. River flats, on the other hand, lent themselves to large-scale working by floating bucket dredges. Gravel tailings dumped from these dredges can be seen today as striking man-made landforms, which from the air look like the folds of a concertina (Fig. 20.12).

Fig. 20.12. Gold dredge tailings on the West Coast are mostly obscured by rapidly growing vegetation. Here at Kumara their original shape can still be detected alongside the Taramakau River (bottom left).
Photo courtesy of Google Earth.

The 'Blue Bottom' referred to in many historic accounts of alluvial gold-mining is a blue-grey mudstone formation that is a widespread member of the younger cover strata on the West Coast, aged around 10 myr (see below). It underlies much of the young gravel and sand, and so miners knew that when they reached the Blue Bottom they had come to the bottom of the gold-bearing gravels. They also knew that the richest gold-bearing gravels tended to be at the bottom of the pile, i.e. just above the Blue Bottom, so it acted as an important reference point.

Because sea-level falls by 100 m or more during each glaciation, and because rivers cut down their valleys to follow suit, gold-bearing gravels extend well below present sea-level. However, it is impossible to mine such sands and gravels in underground workings, because of the huge water inflows. Gold, of course, is still being shed from basement rocks, and every stream on the West Coast has gold in its sand.

The Dynamic River Environment

As noted earlier, the steep, high western face of the Southern Alps is the cause of high orographic rainfall. The combination of height, steepness and wetness creates a highly dynamic river system. The rivers are closely spaced, fast moving and supplied with large amounts of sediment, especially gravel – a common engineering problem on West Coast roads is the build-up of gravel at bridges.

The sediment tends to be supplied to the river in spurts. For example, 11 km upstream from the main road bridge over the Poerua River is the 2130 m Mt Adams. One night in December 1999, a large slab of rock fell from the summit of Mt Adams, broke up, and turned into a rock avalanche with an estimated volume of about 12 million cu m. The avalanche created an air blast that knocked down trees, and it fell 1650 m. It dammed the Poerua River, which then formed a lake measuring 1 km long and 300–400 m wide, containing an estimated 6 million cu m of water. Six days later, during heavy rain, the dam collapsed, sending a large flood downstream. The river relocated part of its course across farmland, and spread large quantities of gravel there. The excess gravel from the avalanche and dam will continue to be spread downstream during floods for years to come, causing problems at the road bridge and other points along the river's flow. Similar rock falls happen in the Southern Alps nearly every year.

All West Coast rivers are subject to dramatic course changes, on account of the large volumes of sediment they carry and their propensity to flood. South of Harihari, for example, the Waitangitaona River used to flow north from the mountain front to join the Whataroa River. Following a build-up of sediment in its bed, it switched course during a flood in 1986, just at the point where it is crossed by SH6, and flowed instead into Lake Wahapo and thence into the Okarito Lagoon. The effect of this change in course was to turn Lake Wahapo into a turbid body of water, to kill a stand of kahikatea trees, and to endanger SH6 where it brushes the outflow from the lake. The lake outflow, of course, became much stronger after the course change.

West Coast rivers and the Main Divide

As we saw in Chapter 19 (for example, Box 19.3), the geology of the Alpine Fault determines that the crest of the mountains is being pushed continually northwestwards. The same mountains, however, create the ultra-wet weather system that is trying to destroy them, so there is an ongoing conflict between the forces of tectonics and erosion. The result is that the Main Divide, the division between west-draining and east-draining rivers, does not coincide with the line of maximum rock uplift, but is pushed back towards the east.

Just how far the Main Divide is pushed at any one point depends on the vagaries of river erosion, and more particularly the processes of headwards sapping and river capture. Rivers enlarge their valleys by removing the debris that slope-forming processes deliver to them, thus encouraging the latter to work harder. The steeper and more vigorous the river, the faster the valley enlarges and lengthens by headwards 'sapping'.

Headwards sapping eventually lowers a ridge separating two rivers, and the lower and more vigorous river 'captures' the headwaters of the less vigorous. The capture is often recorded as a pronounced bend in the lengthened course of the vigorous river, marked by a wind gap (a saddle in the ridge indicating where the weaker river used to go). The beheaded river becomes an underfit (or misfit) river that, having lost its headwaters, is 'too small' for the valley through which it flows. An example of river capture pushing the Main Divide eastwards occurred when the Haast River in southern Westland captured the Landsborough River, which had formerly flowed south into Lake Hawea and thence to the east coast. The upper Landsborough thus switched from east of the Main Divide to west of it, while its lower reaches remain as the Hunter River feeding Lake Hawea. This tendency is occurring all along the western side of the Southern Alps. Every low point on the Main Divide marks a place where two rivers are cutting back into the ridge, with the potential for one to capture the other.

The most voluminous river on the West Coast, the Buller, crosses the region at its widest point, flowing 100 km from Lake Rotoiti, just below the Main Divide in Nelson Lakes National Park, to the Tasman Sea at Westport. Like most large rivers it must have a long history, but there is little record of this. However, the fact that it cuts across all major geological structures, apart from a 10 km stretch north of Murchison, where it takes the easy route and follows the centre of the Longford Syncline (see above), indicates that it is superimposed on the younger geology. It has managed to maintain its course by downcutting as fast as the various structures have risen. Since the development of those structures has taken place mostly in the past 5 myr, the Buller River is probably at least that old.

The Dynamic Beach Environment

'Dynamic' is a word that can easily be overused with reference to the West Coast, where all geological processes are in top gear. However, it is the only suitable word to use when talking about the region's beaches, which are generally steep-faced as a result of the combination of coarse sand grains and strong swells. As summarised in Boxes 14.3 A, B, persistent swells from the southwest, driven by the prevailing winds of the Roaring Forties, push a longshore drift of beach sand and gravel towards the north along this coast. The drift is recognisable from the features it forms, e.g. the pronounced asymmetry of Heretaniwha Point, the south head of Bruce Bay. Here, longshore movement of sand has built up the beach on the west side of the point almost as far out as the end of the point. The beach east of the point, however, is recessed by nearly 2 km, because sand is being moved northeastwards away from it all the time. The point is, in effect, a natural groyne.

Among the pebbles and cobbles on the beaches, some are obtained from the greywacke and schist rocks of the Southern Alps, i.e. from east of the Alpine Fault, because the Main Divide is several kilometres east of the fault. Of these, fewer are schist pebbles, because schist is quite a soft rock and so is readily broken down by wave action on a beach. There are also pebbles of much older basement rocks (granites and hard sandstones) of the West Coast. White quartz pebbles come from both sources (quartz is by far the most resistant of the common minerals), and white quartz veins can be seen in pebbles of the other rock types.

There is a great variety of minerals among the sand grains, although you will need a 10× hand lens to see them well. Quartz is always common (generally clear and glassy), along with feldspar (white and opaque). In addition, there will be grains of garnet (red), ironsand (two types, magnetite and ilmenite, both purple-black), and augite and hornblende (both green to brown), as well as the much rarer zircon, monazite and gold. Ilmenite is a titanium iron oxide, and is one of the ores of titanium – West Coast beaches have been investigated as a possible mining source.

Chapter 21

Northwest Nelson

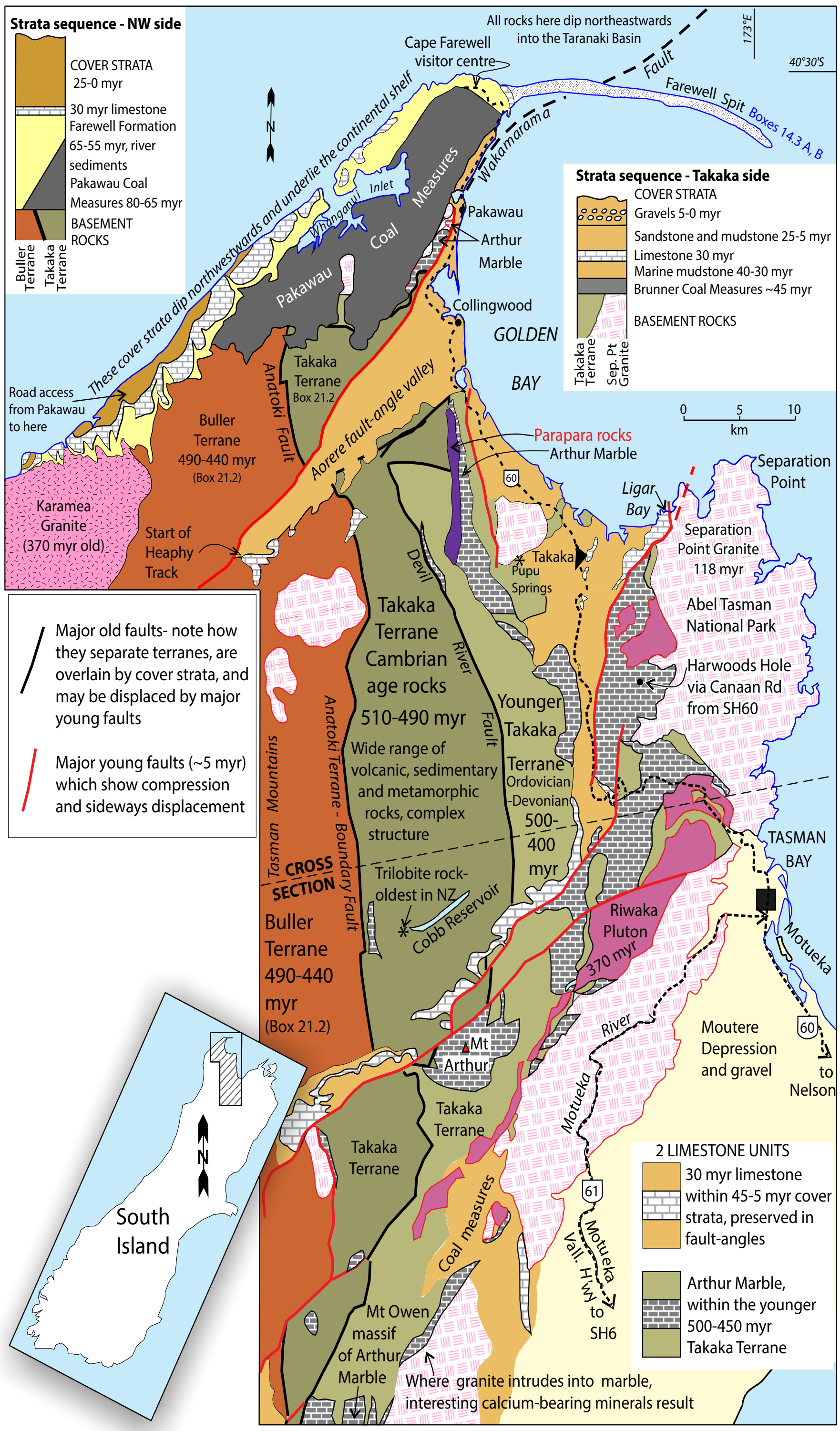


Fig. 21.1. Geology of Northwest Nelson. Cross section is shown in Box 21.2.

Northwest Nelson contains the oldest rocks in New Zealand. These are best seen in one of our largest national parks, Kahurangi, where they are in their least altered – and therefore most recognisable – form. The same rocks occur further down the West Coast and in Fiordland, but in many cases are either inaccessible or have been changed by metamorphism into something completely different. In terrane terms, the rocks belong to the Buller and Takaka terranes (Fig. 20.2, Box 20.1).

Geological History

As discussed in Chapter 20, the western Buller Terrane was deposited adjacent to a continent (presumed to be Gondwana, but where exactly we don't know) around 480 myr ago. No volcanic activity is recorded in the terrane. It docked with Gondwana around 400 myr ago.

The Takaka Terrane, on the other hand, contains older rocks, but came from further afield and was attached to the Buller Terrane (and thereby to Gondwana) later, some time after deposition of its youngest sediments (415 myr) and before intrusion of the Riwaka Complex rocks at 370 myr. As discussed in Chapter 20, the Karamea Granite (390–295 myr), which occupies a large area of Kahurangi National Park, marks the line of a major continental magmatic/volcanic arc that accompanied subduction of the old Phoenix Plate beneath Gondwana. In view of the antithetical relationship between arc magmatism and terrane docking discussed below, the Takaka Terrane was most probably docked before the Karamea Arc got started 390 myr ago.

A long period of erosion followed the addition of the Buller and Takaka terranes to Gondwana. It was some time before the next terrane, the Brook Street Terrane, docked, and it was during that interval that the Karamea Arc was active. As noted in Chapter 20, the Karamea Arc on Gondwana was switched off at about the same time as the Brook Street Terrane (an old oceanic magmatic/volcanic arc, and the next terrane to dock) began to accumulate. There is no known connection between the two facts. We don't see the Brook Street Terrane in northwest Nelson, but it is present in east Nelson (Chapter 22).

As discussed in Chapter 5, it was generally the case that the magmatic arc on Gondwana (the Karamea Batholith followed by the Median Batholith), which was driven by subduction of the old Phoenix Plate, was active between the times of arrival of terranes, i.e. while subduction was proceeding 'normally', but was switched off during the docking of new terranes. This was presumably the result of the adjustments to subduction parameters necessitated by terrane docking events (Boxes 5.1, 5.2, 5.3). We do not have an explanation of why the arc was inactive for 50 myr between 295 myr and 245 myr ago (Box 5.4). Subduction must have proceeded as usual, because it brought the Brook Street Terrane to the margin for a docking event dated at 255 myr. The arc-free interval between 255 myr and 245 myr may therefore have been the result of adjustments to subduction required by the docking of the Brook Street Terrane, which would explain the relocation of the new Median Arc to a position 40 km east of the Karamea Arc.

We see the Median Arc now as a belt of granites and related deep-seated igneous rocks, emplaced between 245 myr and 105 myr ago. We call it the Median Batholith, and it marks continued subduction of the old Phoenix Plate. The volcanic superstructure of that arc, which would have been highly voluminous, has all been eroded away apart from a tiny remnant on the West Coast (Chapter 20) and some remnants in Fiordland and Stewart Island/Rakiura (Chapter 16).

Following our separation from the Antarctica-Australia sector of Gondwana and subsequent subsidence during phase three of New Zealand's formation, younger cover strata were laid down over the whole northwest Nelson region, as they were over most of the rest of the country (Chapter 4). The strata have been eroded away from most areas of the region, because of the tectonic activity of the last 5 myr, but remain where they have been tucked down into fault-angle depressions – see the typical linear form of younger rock outcrops, adjacent to a major fault, on the map in Fig. 22.1.

The cover strata here are similar to those on the West Coast in that they include two sets of coal measures at the base. The older set is located in the far northwest, beyond Collingwood, and is described in Fig. 21.1. It comprises the fill of an extensional rift basin, the Pakawau Basin, which was one of many formed between 120 myr and 90 myr ago (Chapter 4). The basin has been inverted by later compression, as discussed in Chapter 20 and Box 21.1 A, B. As on the West Coast, there was a period of uplift and erosion, lasting in this case from 55 myr to 40 myr ago. 'Normal service' resumed at 40 myr, with the standard pattern of slow subsidence of Zealandia accompanied by gradual deposition of a transgressive sequence, beginning with basal coal measures. The non-marine coal measures were overlain by shallow marine sandstones and mudstones, before the dwindling supply of sediment led to the usual prominent 30 myr limestone beds. Interestingly, the 30 myr limestone is not present around Nelson city, for reasons that are discussed in Chapter 22.

The tectonic pulse 25 myr ago saw limestone deposition terminated as more sediment became available, and blue-grey marine mudstones were laid down on the limestone. Following the energising of the plate boundary 5 myr ago, northwest Nelson was compressed from the southeast, folded, faulted, uplifted and shortened. Erosion became the order of the day.

Rocks of the Buller and Takaka Terranes

We encountered the Buller Terrane on the West Coast in the form of the Greenland Group (Chapter 20). It consists of a series of developments of flysch strata (alternating sandstones and mudstones; Box 6.5 A, separated by thicker units of black mudstone. The mudstones have been strongly compacted by folding, and are generally called shales. They contain graptolite fossils locally (Fig. 21.2), which date the rocks at around 480 myr (Early Ordovician period). The sandstones are typical size-graded turbidity current deposits. Their distinctive property is the dominance of quartz grains, which is unusual in New Zealand sandstones. In northwest Nelson, the Buller Terrane is separated from the Takaka Terrane by the Anatoki Fault, which is therefore a terrane boundary fault, or suture.

Takaka Terrane rocks are more varied. The most prominent rock unit is the Arthur Marble, named after Mt Arthur in Kahurangi National Park.

Box 21.1 A. Pakawau Coal Measures – northernmost South Island.

As noted in Boxes 20.4 A, B (West Coast Coal Measures) these are older coal measures than those seen around most of New Zealand, dating from between 80 and 65 million years ago compared with 45 to 40 myr. Pakawau Coal Measures are linked to other known coal measures of the same age, deeply buried in the Taranaki Basin to the north, by the fact that they were deposited in the same rift valley. The rift was one of many that formed as part of the continental crust stretching process (120 to 90 myr ago) that led eventually to the opening of the Tasman Sea by seafloor spreading (80 to 55 myr) and the separation of the New Zealand continent from Australia and Antarctica (the final break-up of Gondwana). Box 2.1 explains how seafloor spreading works, and Box 5.5 C puts the whole separation process in historical context. Boxes 20.3 A, B (Hawks Crag Breccia) describes the rift-valley stage of the process.

Why didn't this rift valley become the Tasman Sea?
The answer to that question lies in the way that thin skins on spherical objects split when they are stretched from within, for example a tomato skin. The split commonly takes a three-rayed form, each ray or arm being a rift valley oriented at 120 degrees to the other two. As stretching proceeds, and begins to move the pieces of skin (the lithosphere) apart, one of the three arms invariably fails to develop, becoming a "failed arm". The Pakawau—Taranaki Basin—western continental shelf North Island belt of old coal measures were deposited in just such a failed rift, which, having stopped developing, then filled up with sediment. A different rift became the Tasman Sea.

The Pakawau rocks are about one kilometre thick and rest on old basement rocks of the Buller and Takaka terranes. The lower part of the stack of strata consists of river channel and floodplain sediments, and contains many thin (up to 50 cm) coal seams. The upper part consists of shallow-marine and coastal plain sandstones and mudstones; coal seams are only developed around the southern part of Whanganui Inlet.

The only place where Pakawau rocks are well exposed is around the inner coast of Whanganui Inlet, and there it is mostly the upper unit that is seen.

Farewell Formation
This is another set of river-deposited sedimentary rocks that overlies the Pakawau Coal Measures, and in the south rests directly on basement rocks, as can be seen on the map. Despite being river-deposited there are no coal seams here, though the equivalent rocks further north in the Taranaki Basin, the deeply buried Kapuni Formation, do contain coals (important coals, too, because they are the source of much of the natural gas that is recovered from the Taranaki Basin). These rocks are aged between 65 and 55 million years, and like the Pakawau Coal Measures they are confined to the Pakawau-Taranaki Basin. They are widely exposed around the coastline, both the open coast and the Whanganui Inlet, where they present striking exposures of conglomerates and cross-bedded sandstones.

The sequence around Takaka
As shown on the map, the sequence of cover strata is different around Takaka, which was never in the failed rift valley. Pakawau Coal Measures and Farewell Formation were never deposited there (see "structural inversion", below). The Takaka area was an eroding landmass at the time (80 to 45 million years ago). In common with much of New Zealand, the Takaka area began to subside and to accumulate sediment around 45 myr ago. First, river floodplains allowed coal measures — the Brunner Coal Measures (compare Box 20.4 A) — to accumulate, then the sea encroached and marine muds accumulated on top of the coal measures.

Box 21.1 B. Pakawau geological history.

The 30-million-year limestone

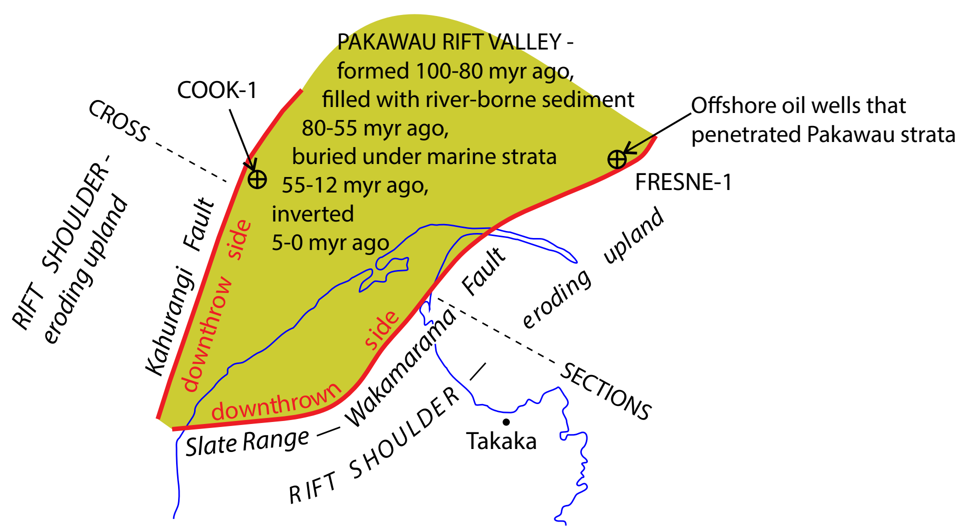
Again like most other places, a widespread limestone accumulated around 30 myr ago, in response to the low sediment yield from the now deeply eroded landscape. By this time, all of northwest Nelson was behaving in a similar fashion, and the limestone is a useful indicator formation, easy to recognise and forming prominent bluff landforms.

After 25 myr, as elsewhere, the tectonic tempo began to hot up, and limestone was replaced by a renewed influx of sand and mud from a now-rising landmass. Northwest Nelson rose above sea level around 12 myr ago, and mountains and widespread gravels appeared about 5 myr ago.

Structural inversion

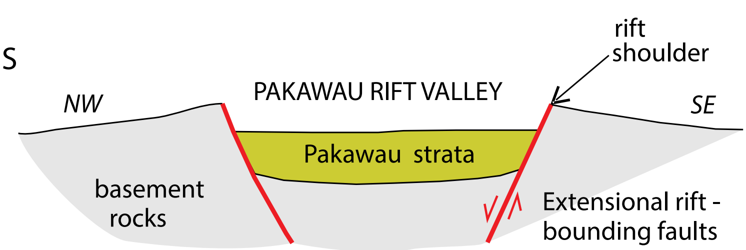
In Chapter 20, West Coast, an example of structural inversion is described, whereby an older downfolded sedimentary basin was turned inside out to become a mountain range, following a change in the country-wide tectonic regime 25 million years ago. A similar inversion occurred here in the far northwest of the island, when the Pakawau Rift was turned inside out, as set out in the following diagrams:

GEOGRAPHY OF NORTHERNMOST SOUTH ISLAND, 70 MILLION YEARS AGO



SEQUENCE OF CROSS SECTIONS

A. 70 myr ago — the Pakawau rift structure is shown here on modern geography, and was originally much wider than shown. Narrowing is a result of the regional compression and structural inversion of the last 5 myr (final cross section).

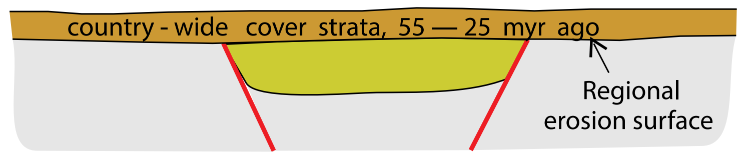


The general area stood high at this time because the rocks had been heated by the developing Tasman seafloor spreading centre underneath, 100-80 myr ago.

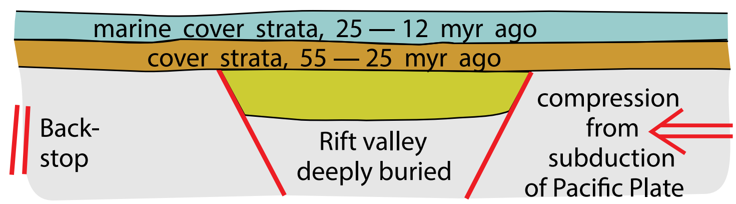
(continued on next page)

Box 21.1 B. Pakawau geological history *(continued).*

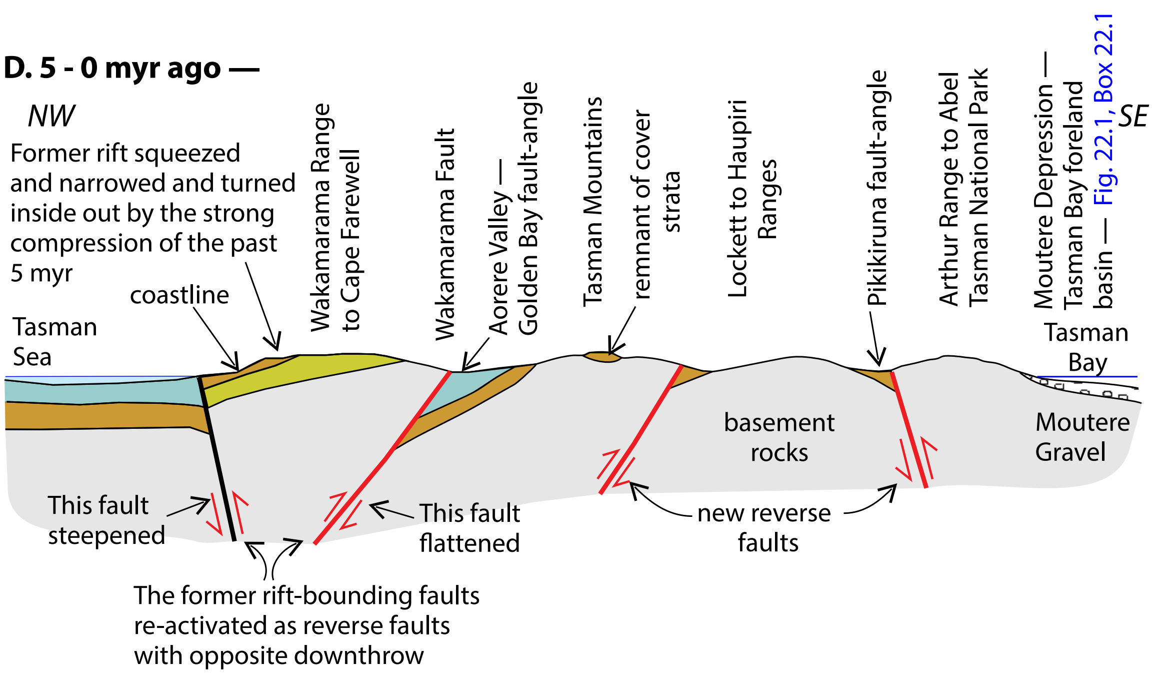
B. 30 myr ago —
the area was eroded to a near-plain, it subsided because the rocks cooled, and it was covered by a typical sequence of cover strata (Box 15.9 A, B). The rift-bounding faults were inactive from 80-5 myr



C. 20 myr ago —
vitalisation of the Australian-Pacific plate boundary 25 myr ago (Box 5.5 C) put the area under mild compression, but nothing spectacular occurred. Uplift further east, nearer the plate boundary, caused an influx of sand and mud, changing the nature of marine sediment being deposited from limestone to sandstone and mudstone. Marine strata continued to accumulate until compression raised NW Nelson above sea level around 12 myr ago.



D. 5 - 0 myr ago —



Schematic section across NW Nelson at the present day, with a large vertical exaggeration, showing structural inversion of the Pakawau rift, and other effects of 5 myr of strong compression

Fig. 21.2. Ordovician (480 myrs old) graptolite fossils in black shale from Cobb Valley, northwest Nelson. Photographer Bruce Hayward.

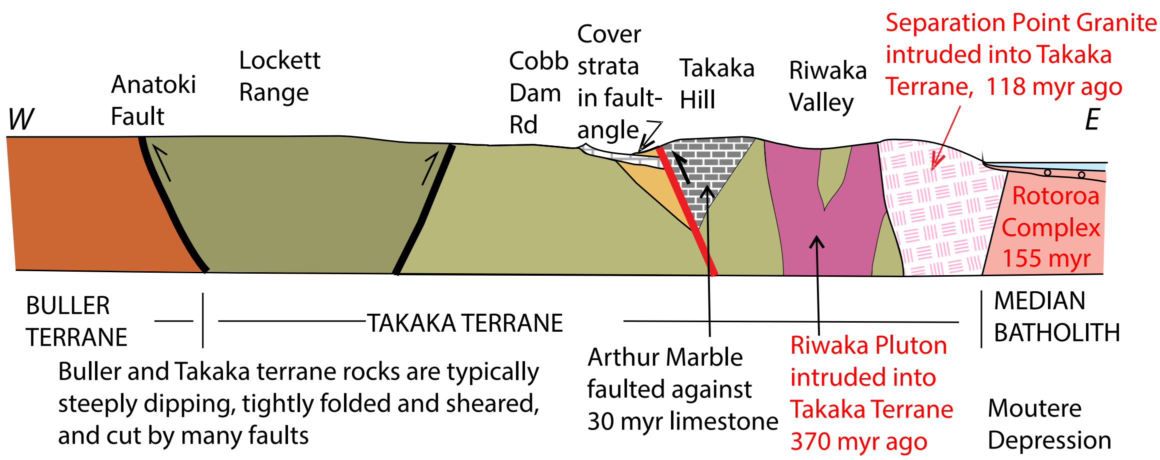
Arthur Marble

Marble is limestone that has been metamorphosed by heat and pressure so that the various particles of calcium carbonate (shell fragments and so on) have recrystallised. That is, the rock is still made of crystals of lime, but these are new crystals; none of the original components is recognisable any more. Being made of lime, moreover, marble responds to weathering processes in exactly the same way as limestone, dissolving in slightly acidic rainwater to form a karst topography characterised by fluted rock outcrops, sinkholes (tomos) and caves. The Arthur Marble contains the country's (and one of the world's) most extensive single-cave systems, including Nettlebed and Ellis Basin, on Mt Arthur itself. Being made of even-sized crystals, and often coloured by impurities (pure varieties are white), the marble is also a popular sculpting stone that is easily cut and sawn (lime crystals are softer than steel).

Arthur Marble outcrops extensively within a north–south strip running from Takaka to Mt Owen (Figs 21.1, 22.6) and there are a few smaller areas further north and south. Mt Owen is a superb example of karst topography. To recognise Arthur Marble in the field, look for areas with many natural grey rocky outcrops, commonly with vertical flutings (sharp ridges separated by rounded channels). Joints may be preferentially dissolved, forming channels filled with vegetation, and there is little or no surface running water. Sinkholes – funnel-shaped hollows in the ground surface down which rainwater enters the cave systems – are numerous and range in size from a few to several hundreds of metres across. Many such features are visible from SH60 across the top of Takaka Hill. The impressive vertical shaft called Harwoods Hole is accessible via Canaan Road.

Layering may be visible in the marble (again, seen along SH60 at the top of Takaka Hill), and it may show folding. Where the original limestone contained layers of quartz sand, you may see residual patches of harder sandstone forming upstanding blobs that represent the hinges of tight folds. The limbs of the folds have been sheared out, and the process is exactly the same as happens during the formation of schistosity in schists – the lime crystals have re-formed and the new layering is, in fact, schistosity. One place where this relationship may be seen is alongside the track from Flora Saddle to Mt. Arthur, on the first ridge-top above the bush-line, just past the hut.

Box 21.2. Cross section simplified.
Position of cross section is shown in Fig. 21.1.



Note that almost every structure in northwest Nelson indicates compression, and that many contacts between rock units are faults.

Fig. 21.3. Fluted Arthur Marble karst adjacent to SH60, near the top of Takaka Hill. Photographer Bruce Hayward.

In places, metamorphism of the Arthur Marble was incomplete, and fossils in the original limestone survive. These include rare corals and sea-lily (crinoid) stem ossicles, dating to 450 myr (Late Ordovician period). Arthur Marble also contains the only ooliths in New Zealand rocks; these are tiny spheres of lime formed in shallow tropical seas, and look like fish roe. They tell us that these rocks originally formed at, or north of, the Equator, before being transported to the margin of Gondwana as an exotic terrane.

Arthur Marble contains prominent white veins of calcite crystals, formed during the folding and metamorphism processes. These veins themselves may have been folded by continuing compression. The best place to see structures in Arthur Marble is in Christ Church Cathedral in Nelson, which is built from the stone. Look in particular at the polished cylindrical pillars inside, as these show the complicated folding of the original sedimentary layers.

Older rocks associated with the Arthur Marble

Arthur Marble is simply one component in a complex suite of old rocks (510–400 myr old) that underlie northwest Nelson (Fig. 21.1). These rocks have a wide range of volcanic and sedimentary origins and a highly complex structure, and, in some cases, have been metamorphosed to schist (Box 15.3 A).

Fig. 21.4. Christ Church Cathedral, Nelson, is made of Arthur Marble (building stone name Kairuru Marble) extracted from near the top of Takaka Hill. Photographer Bruce Hayward.

The Takaka Terrane originated as a volcanic island arc, which today forms the older part of the terrane (510–490 myr), located between the Anatoki and Devil River faults (Fig. 21.1). Rocks associated with this part of the terrane are most easily seen around Cobb Reservoir. They include Trilobite Rock, a small limestone block containing fossil trilobites dating back 510 myr (Middle Cambrian period) that is presently believed to be the oldest rock in New Zealand (Fig. 5.1). The rock is near Trilobite Hut, and is accessible by car; it is a designated scientific reserve within the Kahurangi National Park, so taking rock and fossil samples is not permitted. Most of the trilobites are small and difficult to see. The volcanic breccias and conglomerates of the arc are, on the other hand, clearly visible from the walking track running from the Cobb Dam westward up the northern slopes above the reservoir.

The younger part of the Takaka Terrane (490–415 myr), is preserved east of the Devil River Fault, and comprises sedimentary rocks that apparently accumulated on top of the volcanic arc after volcanic activity ceased. Because the older and younger parts of the terrane are now separated by a major fault, there is a degree of supposition in this statement. It is possible that the two groups of rocks should be regarded as separate terranes, because they have different compositions – there are no volcanic components in the younger rocks, for example, while the older rocks are entirely volcanic, apart from the Trilobite Rock and a few other limestone bodies, which accumulated in shallow lagoons around the marine volcanoes. A mélange (large-scale tectonic breccia), the Balloon Mélange, is associated with the boundary between the two parts, and this is normally an indication of a tectonic contact. So, if the two parts are indeed separate terranes, it would appear that they came together before docking with the Buller Terrane.

Rocks of the younger part include limestones (now mostly changed to Arthur Marble), quartz-rich sandstones (found at and around Hailes Knob) and mudstones that have mostly been metamorphosed to schist (e.g. on the western side of the Takaka Hill road, between the marble and the 30 myr limestone). The youngest rocks in this group are sandstones and mudstones exposed around Baton River in the Motueka Valley; these contain fossil brachiopods (lamp shells) dating back 415 myr (Early Devonian period). The whole assemblage is typical of a passive continental margin, i.e. where no subduction took place.

The Parapara rocks – a link with Gondwana

In a single north–south strip located northwest of Takaka, and forming Parapara Peak and the ridge leading up to it, there are sedimentary and metamorphic rocks that are around 500 m thick, and that date back 270–230 myr (Early Permian to Middle Triassic). These rocks differ from all other old rocks in the region because they were deposited where we see them now, on top of Takaka Terrane rocks, whereas all other rocks of that age in New Zealand are found in terranes that were brought here (to the Gondwana margin, that is) from locations elsewhere. In addition, the Parapara rocks contain glacial dropstones and fossils that also occur in Tasmania and eastern Australia. All of this adds up to the fact that these rocks were deposited as cover strata on the Gondwana basement, during the Permian glaciation of the Australian sector, and before the remaining New Zealand terranes were accreted to Gondwana. They are closely related to the Permian and Triassic strata of southeastern Australia, which makes them an important piece of the Gondwana story, and a valuable link between New Zealand and Australian geology.

Granites

As noted above, both Buller and Takaka terrane rocks have been extensively intruded by granites. Areas where these terranes are found – northwest Nelson, the West Coast, Fiordland and Stewart Island/Rakiura – have long been known to geologists as the Western Province of basement rocks, to distinguish them from the rest of the country's basement rocks, the Eastern Province, which are an assortment of terranes that docked with Gondwana after the bulk of the granites had been intruded, and which therefore lack old granites. The magmatic arc and terrane assemblage story is explored further in Chapter 5.

We have already discussed the Karamea Granite (390–300 myr), which comprises the roots of a major magmatic/volcanic arc recording subduction of the old Phoenix Plate beneath Gondwana (see above). That arc died, and for 65 myr there was no arc that we know of. Reinstatement of an arc occurred about 245 myr ago, and its roots today are situated 20–40 km east of the Karamea Granite, although outlying portions of it actually intrude into Karamea Granite. The older granites and related plutonic rocks of this feature, called the Median Batholith, form a large area around the Nelson Lakes (Fig. 22.6) and underlie the Moutere Depression, where they are dated at around 155 myr. Finally, the granites of the Paparoa metamorphic core complex of the West Coast (Chapter 20) and the Separation Point Batholith that forms Abel Tasman National Park in northwest Nelson were intruded between 120 myr and 110 myr ago. Separation Point Granite can be examined close up at Christ Church Cathedral in Nelson, whose steps were built from the rock. The Median magmatic arc finished around 105 myr ago, possibly for reasons associated with the docking of a number of terranes in quick succession (Chapter 5). The rocks produced by that arc are believed to extend the full length of western New Zealand.

Cover Strata

Northwest Nelson contains a varied assemblage of cover strata resting on the old basement rocks, the broad features of which are shared with the rest of the country. The oldest cover strata occur in the far northwest, around Whanganui Inlet. They comprise the sedimentary filling by rivers of one of the rift valleys – the Pakawau Basin – that formed around 100 myr ago, as a spreading regime was established under Gondwana during phase two of New Zealand's formation (Chapter 4). Box 21.1 B describes these rocks, and the subsequent inversion of the rift. Evidence of these old Gondwanan rift valleys is rare on land, although many more are preserved beneath the seabed. The Pakawau Basin is unique in that it is partly on land and partly under the sea (where it is a southern extension of the Taranaki Basin; Chapter 11).

Following the filling of the rift, there was widespread erosion between 55 myr and 40 myr. Eventually, gradual subsidence of the new continent Zealandia, now fully separated from Australia/Antarctica, led to the characteristic countrywide sequence of cover strata. Coal-measure swamp and river deposits were followed by shallow marine sediments, leading to the deposition of the very widespread 30 myr limestone. The time at which the basal coal-measure sediments were laid down varies from place to place, but in northwest Nelson it happened around 40 myr ago.

Coal seams here are sufficiently developed to have supported small mining operations in various places between Nelson city and Mangarakau, south of Whanganui Inlet. The 30 myr limestone, known locally as the Nile Group, is preserved in down-faulted areas over much of the region, particularly in the west and southwest. It is divided into two variants, or facies: the typical shelf limestone facies, forming bluffs and caves, which is widespread in the west and around Takaka; and the deeper-water basin facies, a calcareous mudstone, which forms much of the Matiri Range north of Murchison, including the hill called the Haystack.

Fig. 21.5. Mt Misery plateau (foreground) and The Haystack (beyond it, left of centre) in the Matiri Range, northwest Nelson, is formed from horizontally-bedded, 30 myr-old limestone of the Nile Group. Photographer Lloyd Homer, GNS Science.

A tale of two limestones

Northwest Nelson thus contains two prominent limestones, the old Arthur Marble and the 30 myr limestone common to most of New Zealand. In places, they occur cheek by jowl because of faulting, as you can see in Fig. 21.1. So how can you tell the difference between the two? Well, it isn't always easy, as they both weather to the same grey colour and have vertical flutings. If the dip of the limestone is low, it may form bluffs and scarp and dip-slope profiles (Box 12.6), which the marble does not, but if the dip is steep these landforms may not be present. In both cases, the typical flagginess of the 30 myr limestone may show (roughly parallel layers up to 15 cm thick, as at Pancake Rocks in Punakaiki), and fossil shells may be visible. However, schistose layering in the marble may also have a low dip and imitate the limestone.

Distinguishing between the marble and limestone is particularly difficult along SH60 on Takaka Hill, where the crest of the hill is in marble, while the lower, south-pointing Eureka Bend hairpin on the western, Takaka side is in the 30 myr limestone. In the road cuttings here, the limestone is more massive and chunky than the marble, but the fossils are hard to see (small white specks) and the fluting is coarser than on the marble. Why would the younger limestone be at a lower elevation than the marble, when it should be above it? The answer lies in a fault, the major Pikikiruna–Pisagh Fault, which is responsible for one of the elongate fault-angle wedges of preserved younger rocks referred to earlier. The fault passes along the hillside just above the Eureka Bend and has downthrown all rocks on the western (downhill) side (Box 21.1 B).

The easiest place to see the 30 myr limestone in northwest Nelson is at Tarakohe Port and the Takaka end of Ligar Bay, east of the township. A large quarry and cement works was in operation here until the late 1980s, and the cliffs and road cuttings around it display the flaggy limestone superbly. Flagginess in joint-bounded blocks of limestone up to 25 m high shows varying angles of dip, because the blocks themselves have tipped. Blocks of limestone used for the breakwater around the port show fossil scallop and fan shells on bedding planes. They also show the nature of the flagginess – note how the thin planes that separate the flags of limestone undulate, separate and rejoin. These planes are not original sedimentary layering, but were formed by chemical solution of some of the lime while the rock was buried, leading to a concentration of mineral sand grains.

Between Ligar Bay and Takaka, the 30 myr limestone is prominent, especially at the Grove Scenic Reserve in Clifton and at Labyrinth Rocks Park closer to the township. The younger strata (mudstones) that overlie the limestone can be seen around the old quarry at Tarakohe, and here all the younger strata are down-faulted against the Separation Point Granite, which can be seen at the northern end of Ligar Bay. This fault is the northern continuation of the Pikikiruna Fault mentioned above.

Fig. 21.6. Extinct fossil fan shells in 30 myr limestone near Eureka Bend on the west side of Takaka Hill summit road. Photo 15 cm across. Photographer Bruce Hayward.

Cover strata younger than 30 million years

Throughout New Zealand, the kick-starting of tectonic activity 25 myr ago, resulting from activation of the convergent plate boundary, had an immediate effect on the type of sediment being supplied to the shallow seas around the country. Increased relief saw an influx of mud, shell production was swamped, and the blue-grey formation known as Blue Bottom (Chapter 20) was deposited on top of the limestone. This now compacted mudstone is commonly called 'papa'. In some places, such as around Murchison, it contains lenses of flysch, or alternating mudstones and turbidite sandstones.

The continued tectonic compression caused uplift of the area, which is quite close to the Alpine Fault. Deeper-water mudstones gave way to shallow marine sandstones, and from around 10 myr ago the region was mostly above sea-level. The spurt in compression at the Alpine Fault, and consequent growth of the Southern Alps, 5 myr ago, was reflected in increased compression across northwest Nelson, and development of the major young faults that determine the layout of the rocks we see today.

Fig. 21.1 shows clearly how the major old north–south-trending faults (shown in black) are cut and offset in a right-handed fashion by young northeast–southwest-trending faults (shown in red). Cover strata are preserved in elongate wedges on the downthrown sides of the red faults. The faults are high-angle reverse faults with a component of dextral sideways offset, reflecting the compression of the region and its associated stress field.

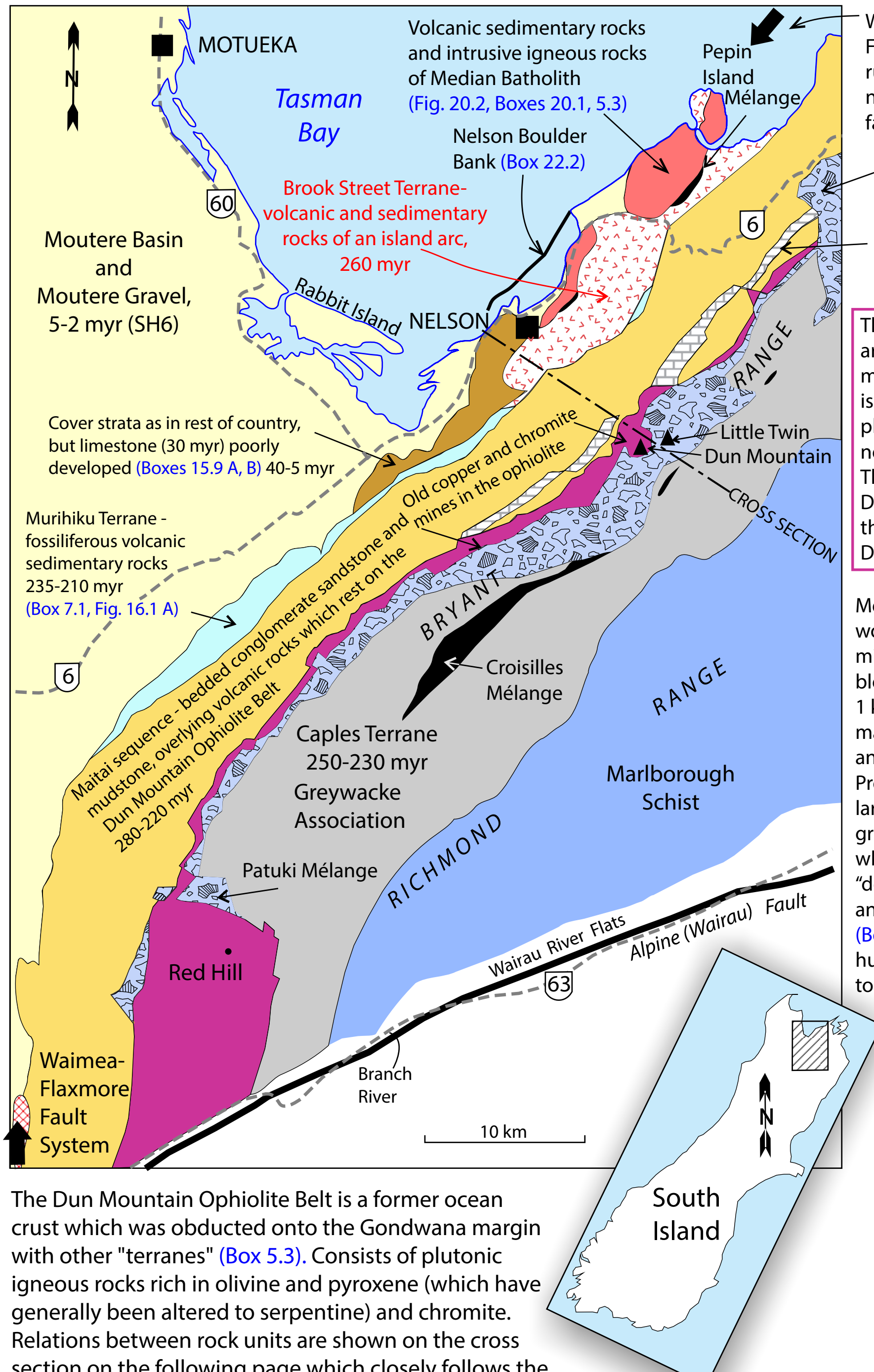
How much compression?

By measuring offsets across faults on geological cross sections from a variety of publications, it is possible to estimate just how much tectonic compression has shortened the region in a northwest–southeast direction. Between a point offshore from Kawhia Harbour (on central North Island's west coast) in the north, across Cook Strait to Hector at the north end of the South Island's West Coast, shortenings of between 3.75 km and 11 km are obtained. These numbers are not precise, but they give a good ball-park indication. They are also useful because they have a bearing on the question of when some of our major basement terranes were squeezed and narrowed through central New Zealand. This matter was expanded on in Chapter 5, but suffice it to say here that the shortenings of the past 5 myr are much less than the shortening of the terranes. The cause of the northwest-directed compression across the Nelson region is discussed in chapters 15 and 22.

We can also go back further, using the same cross sections and the same faults (because they are nearly all old faults that have been reactivated in the opposite direction), and reconstruct the amount of crustal stretching that took place during phase two of New Zealand's formation (Chapter 4). Estimates here, for the same region, range between 5 km and 20 km. Spread over the full length of the cross sections, these are lengthenings of between 3.4% and 21%. Again, they are ball-park figures, but they indicate overall that the crustal extension that took place 100 myr ago has not been fully restored by the more recent shortening.

Chapter 22

East Nelson



The ophiolite rocks are rich in magnesium, which is toxic to much plant life. There is no bush on them. The olivine rock Dunite occurs on the south face of Dun Mountain.

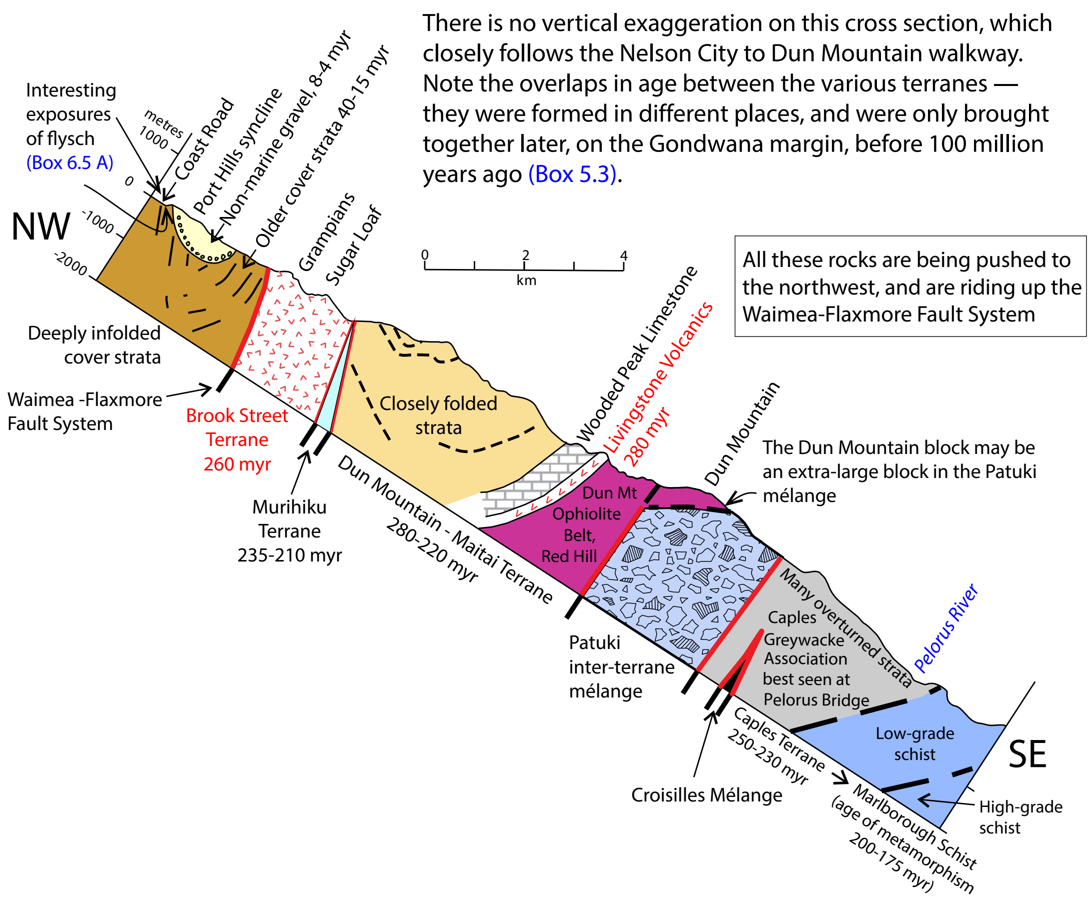
Mélange (a French word meaning mixture) consists of blocks of rock up to 1 km across in a matrix of sheared and ground up rock. Produced when large rock masses grind together, as when a terrane “docks” against another terrane. (Box 5.3). Makes hummocky topography.

The Dun Mountain Ophiolite Belt is a former ocean crust which was obducted onto the Gondwana margin with other "terranes" (Box 5.3). Consists of plutonic igneous rocks rich in olivine and pyroxene (which have generally been altered to serpentine) and chromite. Relations between rock units are shown on the cross section on the following page which closely follows the Nelson City -Dun Mountain walkway.

Fig. 22.1. Dun Mountain-Red Hills and related rocks.

Box 22.1. Dun Mountain and Nelson terranes cross section.

There is no vertical exaggeration on this cross section, which closely follows the Nelson City to Dun Mountain walkway. Note the overlaps in age between the various terranes — they were formed in different places, and were only brought together later, on the Gondwana margin, before 100 million years ago (Box 5.3).



The steepness of bedding dips and terrane sutures, and the narrowness of terranes, have been caused by tectonic compression occurring at various times. At the present time the whole of the Dun Mountain block is being driven towards the northwest, as a consequence of developments in Marlborough (Boxes 15.4 A, B; 15.5). Movement on the various Marlborough faults over the past five or more million years has been moving rock into Marlborough and making Marlborough wider. To make room, Marlborough has been elbowing eastern Nelson out of the way, causing the big bend and displacement of the Alpine Fault (Box 19.3), and causing the deep syncline in the cover strata under Nelson city.

The leading edge of the Dun Mountain block is riding over the Moutere Depression and its offshore extension, Tasman Bay, along the active Waimea-Flaxmore Fault System. These faults outcrop along the face of the ranges through the eastern suburbs of Nelson and Richmond. In fact, Moutere Depression (and Tasman Bay, its flooded northern extension) is directly caused by the over-riding Dun Mountain block — the weight of Dun Mountain rocks is pushing down the crust ahead of the moving block. The geological name for such a sedimentary basin is "foreland basin".

The arrangement of rocks in the deep syncline under Nelson city gives us some idea of the timing of these movements. Older cover strata, aged between 45 and 5 million years, are folded more deeply than younger Port Hills Gravel. This tells us that an original pulse of compression occurred around 5 myr ago, which is about the time that the Southern Alps began to grow, the Marlborough faults began to move rock into Marlborough, and the Moutere Gravel began to accumulate. It also tells us that compression is continuing to the present day.

Tectonically, the 8-4 myr old, steeply dipping Port Hills Gravel of Nelson is a precursor to the extensive Moutere Gravel. It was close enough to the advancing Dun Mountain front to get involved with a second wave of compression and folding around 2 myr ago.

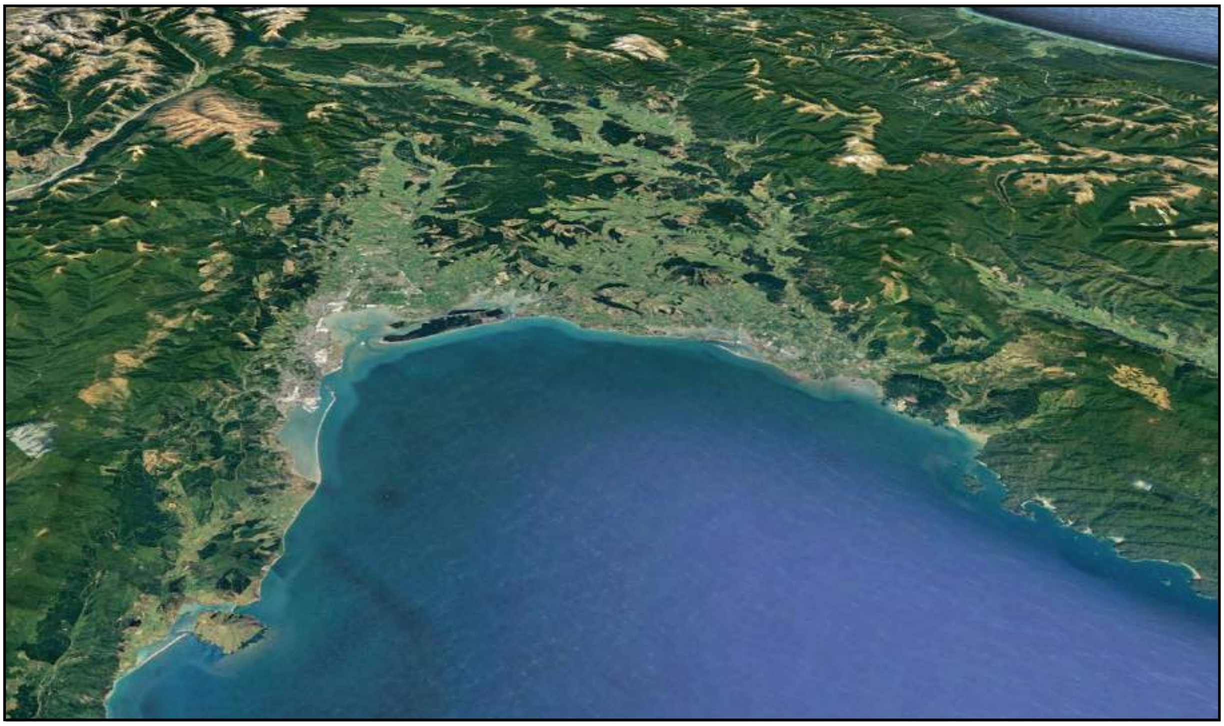


Fig. 22.2. Oblique view south across the head of Tasman Bay into the Moutere Depression with the Richmond Range, Nelson Haven, Nelson City and Waimea Inlet on the east side (left). The Arthur Range of Northwest Nelson is on the west side (top right). Photograph courtesy of Google Earth.

Nelson city and its neighbour, Richmond, lie at the centre of a district that contains a rich geological heritage. Its story is made up of many parts: the docking of the old tectonic terranes to the margin of Gondwana before 100 myr ago; the separation of Zealandia from Gondwana and its subsidence beneath the sea; a rare glimpse into the early evolution of the Australian–Pacific plate boundary; the early growth of the Southern Alps; and the more recent growth of the east Nelson ranges, and their role in the development of Tasman Bay and its continuation to the south in the form of the Moutere Depression.

The story of the old tectonic terranes is summarised in Chapter 5 and in Boxes 5.1, 5.2, 5.3 and Fig. 22.1, Box 22.1. The eastern ranges of Nelson contain portions of four major terranes (see below). In addition, there are extensive areas of rock produced by the collision and grinding together of terranes as they joined – these are the Patuki and Croisilles mélanges and the Marlborough Schist (see below).

The four terranes are narrower, or even locally absent, in Nelson than in most other places, due to faulting. This is connected in some way with their closeness to the Alpine Fault, and the fact that all of New Zealand's terranes are bent into a countrywide lazy-Z shape where the Alpine Fault cuts through the centre of the country (Boxes 5.1, 5.2, 5.3). The Murihiku Terrane, which occupies large areas of southwest Auckland (Chapter 7) and Southland (Chapter 16), has in fact been dismembered into separate slivers as it passes through central New Zealand. The four terranes have all been squeezed, as well as pulled apart in a right-handed fashion across the Alpine Fault, to form the middle limb of the Z. However, whether the displacement across the Alpine Fault actually caused the squeezing, or whether the squeezing was already there and formed a zone of weakness that was exploited by the Alpine Fault, is debated. Chapter 19 has further discussion of the matter, including why the greywacke/schist terranes further east were not affected by the squeezing.

The Median Batholith in Nelson

In Nelson city there are volcanic, sedimentary and granitic rocks of the Median Batholith. This was an eastern portion of the magmatic/volcanic arc that formed above the subducting slab of the old Phoenix Plate between 245 myr and 105 myr ago. Most of that arc is present as a long band of granites further west, as discussed in chapters 20 and 21. Nelson also contains a rare remnant of the volcanic superstructure of the arc.

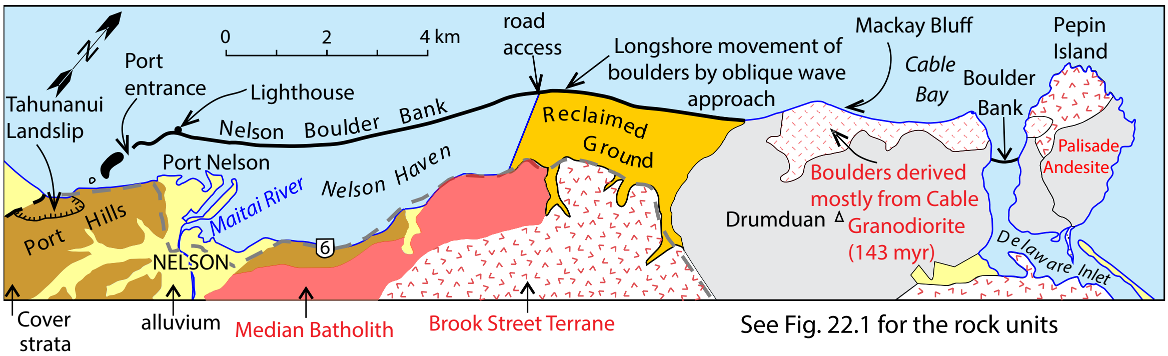
As Fig. 22.1 and Box 22. 2 show, Median Batholith rocks outcrop immediately north of the city centre. Cable Granodiorite forms the prominent hill Drumduan and the coastal cliffs around Mackay Bluff, the latter the source of most of the boulders in the Nelson Boulder Bank. In Box 22.2, the Median Batholith rocks are divided into the Cable Granodiorite (intrusive, 143 myr old) and the Palisade Andesite (extrusive lava, around 150 myr old). Everything else labelled Median Batholith comprises volcanic tuffs and associated sedimentary rocks. The latter contain plant fossils in places, which are important because there are few places in New Zealand where plant fossils of that age (broadly Jurassic – about 170 myr old) occur. In particular, all the other occurrences are in 'exotic' terranes that came to New Zealand from somewhere else. The Nelson plants are possibly the only ones of that age in New Zealand that lived where they are now found.

Median Batholith rocks are best seen around Pepin Island, on the shores of Cable Bay and Delaware Inlet, and on the coast north of the Boulder Bank. Their detailed geology is complicated, and includes a strip of tectonised rock that marks the western boundary of the Brook Street Terrane. The Median Batholith continues to the south of Nelson, where it underlies the gravel fill of the Moutere Depression, to link with Nelson Lakes National Park, where the associated different brands of 'granite' are dated at 155 myr (Fig. 22.6).

Basement Terranes of the eastern ranges

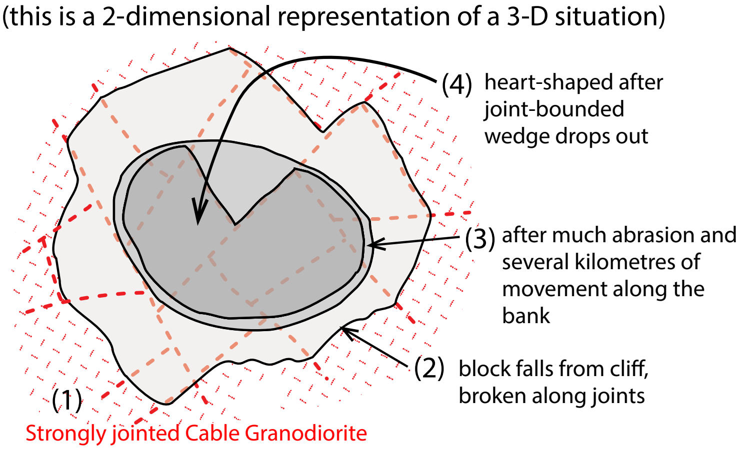
The tectonic narrowing of the four central basement terranes referred to above, along the middle arm of the lazy-Z shape, means that all four can be seen in the relatively narrow northeast–southwest-trending Bryant and Richmond ranges between Nelson/Tasman Bay and the Wairau River/Alpine Fault. Fig. 22.1 shows that the terranes form belts parallel to the ranges, while Box 22.1 shows that they all dip at steep to vertical angles. From the northwest, in order of amalgamation with each other prior to their addition to the Gondwana margin, the terranes are: Brook Street, Murihiku, Dun Mountain-Maitai and Caples.

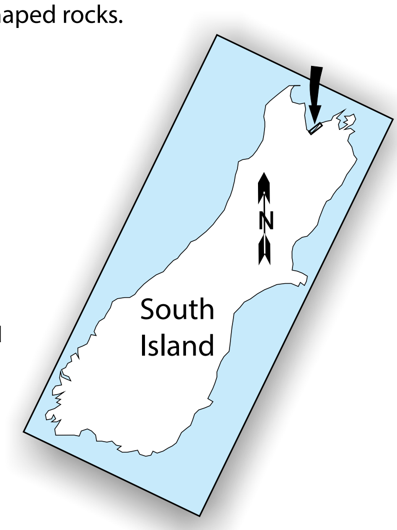
Box 22.2. Nelson Boulder Bank.



The Nelson and Cable Bay boulder banks have formed in the past 10,000 years, since the sea returned to its present level at the end of the last glaciation (Boxes 10.2 A, B). The Boulder Bank at Nelson exists because a predominant wave approach from the north, into Tasman Bay, finds a voluminous source of rocks at the foot of Mackay Bluff, in combination with a coastal embayment to the south of Mackay Bluff. Waves jostle the boulders to the south, and have been able to build the boulder bank almost across the shallow entrance to the embayment, which is now Nelson Haven. The bank is a remarkably long example of its type (13km) and is still growing — the lighthouse was built in 1862, at the southern end of the continuous bank at high tide. Now the bank is continuous for a further kilometre, to the artificial Cut which allows access to Port Nelson. Rocks falling from MacKay bluff are angular, and consist mostly of Cable Granodiorite, a reddish coloured, fine to medium grained, plutonic igneous rock closely related to granite (Box 20.2). The dark mineral is hornblende, the pale mineral feldspar and the clear glassy mineral is quartz. Boulders become progressively smaller and more rounded to the south, as a result of continuous jostling by waves. Because the cable Granodiorite rock is cut by many intersecting joint planes, it is common for joint-bounded wedges to drop out of boulders as they are bashed around by waves, leaving heart-shaped rocks.

EVOLUTION OF A NELSON BOULDER
(this is a 2-dimensional representation of a 3-D situation)





Heart-shaped boulder on the Nelson Boulder Bank. Photographer Louise Cotterall, University of Auckland.

Brook Street Terrane

This is the product of an old (280 myr) magmatic/volcanic arc formed in an intra-oceanic situation, i.e. above a subduction zone taking one plate of oceanic lithosphere underneath another plate of oceanic lithosphere, exactly as is happening now at the Tonga–Kermadec subduction zone. The location of the subduction zone, and the plates interacting there, are not known, although it is likely that the old Phoenix Plate was one of them. This terrane followed the Buller and Takaka terranes, colliding with Gondwana around 250 myr ago. Brook Street rocks form the eastern backdrop to Nelson (Brook Street is in the city), but are concealed by younger rocks to the south, until in the vicinity of Lake Rotoiti. To see them close up, head to Christ Church Cathedral grounds, where brown Brook Street volcanic rocks have been used to build drystone walls.

Murihiku Terrane

Rocks of this terrane are not present north of Nelson, because they have been tectonically excised here. They do, however, appear as a widening strip from the south of the city, where they form a very narrow strip of highly disturbed rocks (Fig. 22.1). Even here, only the Triassic portion (around 210 myr) of the Murihiku Terrane is preserved, as the Jurassic rocks have been tectonically excised (except for a sliver near Lake Rotoiti). The rocks are notable for the presence of *Monotis* fossils, which are found in Triassic Murihiku rocks throughout the country. Beds of these old scallop shells are found in the ranges behind Richmond; the species was named *Monotis richmondiana* by German geologist Karl von Zittel in 1859, but taxonomy has moved on, and it is now reclassified as *Entomonotis richmondiana*.

Dun Mountain-Maitai Terrane

This is a double banger: a slice of oceanic lithosphere and crust (Dun Mountain Ophiolite Belt, 285–275 myr old) is overlain unconformably (i.e. there was some tectonic activity between times, but the contact is a sedimentary, not a tectonic, one) by the Maitai stack of sediment layers (260–240 myr). The relationship here is that subduction was bringing oceanic lithosphere into a subduction zone beneath a continent, in the usual way. As subduction proceeded, a wedge of sediment derived both from the continent itself (quartz sand), and from the active magmatic/volcanic arc on the continent, accumulated in the subduction trench, on top of the upper layers of oceanic crust (here called the Livingstone Volcanic Group). In the Nelson area the unconformably overlying material consists of ocean-floor crust (ophiolite) material coming off topographic highs in the trench, and shells also coming off the topographic highs. At times, these shells were numerous enough to form limestone (e.g. the Wooded Peak Limestone is a persistent unit up to 100 m thick, composed of broken shell mash).

A curious distinction of the Maitai sequence, which is several kilometres thick, is that it does not display any of the features typical of an accretionary wedge (Box 6.2 C), as we would expect in that tectonic situation. Perhaps subduction had ceased, and the Maitai sequence simply filled the old trench. The sediments are typical deep-water ones (e.g. flysch), and in its upper part they contain large (hundred of metres across) blocks of limestone that had slumped down from a continental shelf. Maitai rocks can be seen in various rivers that drain west from the Bryant Range (e.g. the Maitai and Roding), and on SH6 where it crosses the Whangamoa Saddle northeast of Nelson.

Fig. 22.3. Dun Mountain, southeast of Nelson city, is composed of a large block of dunite rock (named after Dun Mountain by Hochstetter) and name bearer for the Dun Mountain Ophiolite Belt. Photographer Bruce Hayward.

The next mystery is what happened after this. At some point, Maitai rocks were compressed, folded into large and small folds, and lightly metamorphosed – to a similar grade, in fact, as the greywacke rocks in a typical accretionary wedge. As with all of our other continental margin terranes, the Maitai sediment wedge, along with, in this case, a slab of the underlying oceanic lithosphere, was somehow detached from its parent continent (which is presumed to have been somewhere on Gondwana, because there were relatively few other continents at the time) and set adrift – for 110 myr. It cannot have collided with the Murihiku Terrane to the west until the latter had ceased accumulating, and that was not until around 130 myr ago. Could that collision have caused the folding of both terranes? The wider issue of the sequence of terrane collisions was discussed in Chapter 5.

On its other, eastern side, the Dun Mountain Ophiolite Belt (a slice of ocean-floor crust) was in collision with the next terrane, the Caples. The fact that the collision was between the Caples Terrane and the ophiolites, rather than the overlying Maitai sediment stack, tells us that the Dun Mountain-Maitai package must already have been folded for the ophiolites to have been turned up – unless, of course, the colliding Caples mass somehow dug deep and brought up the ophiolites from several kilometres down. At any rate, the Caples–Dun Mountain contact is marked by a thick body of tectonic mélange, the Patuki Mélange.

Part of the geological definition of a tectonic mélange is that it contains a wide variety of rocks, including 'exotic' rocks that have been derived from a range of origins, all floating like plums in a pudding. The plums can be any size, from millimetres to kilometres across. The matrix of the pudding is usually serpentinite, an olivine-rich mantle rock that has had water added to it. Serpentinite is dark green-black in colour, soft, and typically riddled with shiny, curved shear surfaces. Some of the exotic rocks in the Patuki Mélange (i.e. not known from either the Caples or Dun Mountain terranes) are white quartz sandstones interbedded with black mudstones, and pillow lavas of a wide variety.

The entire Brook Street–Maitai–Dun Mountain sequence can be seen by walking or cycling along the old Dun Mountain Railway line between Brook Street in Nelson and Dun Mountain itself. The railway opened in 1862 to access and mine lenses of the heavy black mineral chromite in the Dun Mountain Ophiolite Belt; this chromium ore was used to make a green fabric dye. The nature of the ophiolite ensured the mining process was rather hit and miss, and as alternative sources were found for green dye and the wooden bridges on the railway began to rot, the mining venture ceased after ten years.

The ultimate source of chromite here is its occurrence as an accessory mineral (i.e. comprising 1–2%) in dunite, the rock first described and named by Hochstetter from Dun Mountain in 1859. Dunite consists almost entirely of the green mineral olivine. It is one of the principal rocks forming the mantle, making it one of the most voluminous rock types in the earth. We see dunite rarely – only where it has been obducted during plate convergence. Hydration of dunite during tectonic activity produces the serpentinite that forms the lubricating matrix, but the chromite is not affected. Its accumulation in lenses happened during the time the dunite was molten, when it was being formed at a mid-ocean spreading ridge. Heavy grains of chromite crystallised and sank through the melt, accumulating in lens-shaped bodies up to several metres in size.

Dun Mountain itself is probably a large block of dunite in the Patuki Mélange, which escaped hydration to serpentinite. Another exotic rock type that occurs in lumps of various sizes in the mélange is argillite, a very hard, fine-grained light grey rock that breaks with a smooth conchoidal fracture in the same way as glass. Early Maori living in the district used this stone for making tools.

The Patuki Mélange lies alongside the surviving oceanic crustal rocks of the Dun Mountain Ophiolite Belt (dunite, gabbro and the Livingstone Volcanics). The Belt is notable for its stunted vegetation, which is a reflection of the high content of magnesium, chromium and nickel in the soil.

North of Nelson, the Ophiolite Belt is narrow and discontinuous because, like the Murihiku Terrane, it has been tectonically excised in places. It outcrops on D'Urville Island and crosses Cook Strait, where it is responsible for the Stokes Magnetic Anomaly. This is a line of magnetic disturbance that confused early mariners using magnetic compasses, and was named after the hydrographer Captain John Stokes, who first recognised and delineated it in the mid-nineteenth century.

The Stokes Magnetic Anomaly continues northward beneath the western North Island, following the tectonic suture between the Murihiku and Waipapa (Caples equivalent) terranes, and goes out to sea at Ninety Mile Beach. The anomaly is inferred to prove the continuous existence of the Dun Mountain Ophiolite Belt (which is strongly magnetic). The inference is confirmed by a single surface outcrop of serpentinite at Piopio, near Te Kuiti, which is quarried.

South of Nelson, the Dun Mountain Ophiolite Belt widens greatly and form the Red Hills (Fig. 3.1). These rocks have been chopped off by the Alpine Fault, reappearing on the opposite side of the fault as Red Mountain and the Red Hills Range of southern Westland (Fig. 3.2). Both sites are named for the reddish colour of weathered dunite. This was the critical reference point that first indicated to geologists the great 480 km lateral offset across the Alpine Fault (Chapter 19).

Caples Terrane

The Caples is one of our greywacke association, accretionary wedge terranes (Box 6.2 C), named from the Caples River in the Ailsa Mountains on the west side of Lake Wakatipu, near Queenstown. It differs from the other, similar South Island terranes (Rakaia and Pahau) in having more volcanic minerals and rock fragments in its sandstones, commonly giving it a greenish tinge. In appearance in the field, however, it is very similar to the other greywacke terranes. As noted, Patuki Mélange forms the tectonic contact between the Caples and Dun Mountain terranes. There is another, less extensive mélange, the Croisilles, within the Caples Terrane, indicating some extensive tectonic shuffling within the terrane as well as between terranes.

As shown in Fig. 22.1, Caples rocks form a broad band trending southwest to northeast, parallel to the other terranes. The band continues beyond the map to form much of the western half of the Marlborough Sounds. The best place to see Caples rocks is at Pelorus Bridge on SH6 near Havelock. Outcrops on the riverbank by the bridge here show interbedded mudstones (black), sandstones (grey) and conglomerates that have been strongly tectonised. Look for the steep to vertical dips, and evidence of strong squeezing. That evidence includes stretched pebbles, and mudstone beds reduced to a string of lenses. Another feature to look for is incipient axial plane cleavage, the forerunner of schistosity planes, in folds of the squeezed rocks. White quartz veins in the rocks give evidence of movement of hot water through the rocks under pressure. If the weather is dry, wetting the rocks brings out the detail; compare this with the diagram of flysch sediments in Box 6.5 A, as this shows how these rocks would have looked before they were tectonised.

In the regional picture, Pelorus Bridge is located in the transition between lightly metamorphosed rocks of the Caples Terrane to the west, and the more severely metamorphosed Marlborough Schist to the east (Chapter 15). The schist is the product of the collision between the Caples and Rakaia terranes around 200 myr ago (Chapter 5).

Cover Strata

The important period of time coinciding with, and following, the separation of the mini-continent Zealandia from Gondwana and its subsequent subsidence during phases two and three of New Zealand's development, 100–25 myr ago, features prominently in this book. It gave us many economically important things like coal, oil, gas, glass sand, limestone and brick-making clays. During this time, east Nelson and west Marlborough would have been eroded and then covered by typical strata of the gently subsiding sequence like everywhere else in the country. However, because of the high degree of recent tectonic activity they have all been eroded away, apart from tiny remnants around Nelson city and just north of Picton.

Nelson's cover strata are folded into a deep syncline that underlies the city, but they are seen in only a few places because for the most part they are hidden underneath the younger Port Hills Gravel. The cross section in Box 22.1 shows the syncline. Coal measures do exist at the base of the strata, but because of the deep syncline they outcrop in only a narrow strip on the east side of the city, immediately underneath the Waimea-Flaxmore Fault System. The steep dip and sheared nature of the coal measures make mining difficult. However, seams up to 3.3 m thick exist, and three small mines were worked here in the nineteenth century.

Strata of marine origin overlie the coal measures. However, they differ from the 'normal' cover strata in at least two respects, and give us a rare glimpse into the workings of the early Australian–Pacific plate boundary.

Early Glimpses of the Australian–Pacific Plate Boundary

Processes related to the active boundary between the Australian and Pacific tectonic plates have dominated New Zealand for the past 25 myr. However, there is a puzzle connected with the boundary. Wider reconstructions of the history of the two plates, based on seafloor magnetic anomaly stripes in the Pacific, Australian and Antarctic plates, tell us that the boundary has actually existed in the vicinity of New Zealand for 45 myr. The puzzle is that there is little indication in our geology of either tectonic or volcanic effects of that boundary between 45 myr and 25 myr ago.

One place where there is a good record of the missing 20 myr is at the Moonlight Tectonic Zone in the far south (Chapter 17), where the boundary was opening, creating sedimentary basins. North of here, however, plate reconstructions suggest that the plate boundary was convergent. In other words, we should find evidence of a subduction zone. Well, we don't, but this doesn't necessarily mean that it didn't exist. If movement between two converging plates is highly oblique, or very slow, subduction may not turn on the usual cluster of tectonic and volcanic signals.

There are reasons for thinking that this was the case in New Zealand. As shown in Box 5.5 B, reconstructions for 40 myr ago indicate that the pole of rotation of the Pacific Plate – the point around which it rotates slowly anticlockwise – was located much closer to the New Zealand landmass than it is now. Places close to a pole of rotation move more slowly than places further away: think of the rotating record analogy used in Chapter 3.

This is where Nelson comes in. The 30 myr limestone bed common over much of the country is missing here, and in its place there are two formations that point to local tectonic activity. First, alternating sandstone and mudstone strata exposed alongside Nelson's waterfront Rocks Road at Magazine Point indicate the existence around 30 myr ago of a deep localised hole (not a subduction trench) into which sediment poured (Fig. 22.4). The gravity flow mechanisms transporting the sediment were a combination of turbidity currents (Box 6.5 A, B) and debris flows (Box 19.5), and the deposits hold lenses of conglomerate containing cobbles and boulders of granite. A unique feature of these rocks is the form that many of the sandstone and conglomerate layers take. They are lens-shaped mounds, with planar bottoms and convex tops, that probably originated as levées built by debris flows. The rocks dip steeply landwards, into the deep syncline that lies underneath the Port Hills (Box 22.1).

Fig. 22.4. Steeply-tilted, 30-myr-old sandstone and mudstone beds adjacent to Rocks Road, Magazine Pt, between Nelson city and Tahunanui Beach. Photographer Bruce Hayward.

The second formation is seen around Murchison, 80 km southwest of Nelson; it also lacks the 30 myr shelf limestone and received muddy, deeper-water limestone instead. This and the Nelson evidence indicate tectonic activity possibly related to limited sideways movement on the 'cryptic' plate boundary of 45–25 myr ago.

If, as postulated, the plate boundary through and around mainland New Zealand at this time was so sluggish that it left few signs of its presence, we can make a prediction: if we could follow the boundary to the north, getting further away from the pole of rotation, subduction should have sped up, and somewhere we might expect to find a magmatic/volcanic arc. There is just such evidence in the form of an arc remnant in Tonga and Fiji, and arc-related features in New Caledonia. These places are a long way apart now, but they were much closer together 40 myr ago, before a number of arc-related basins opened up.

The plate boundary 25–5 million years ago

Starting around 25 myr ago (the time of the 'revving up' of the Australian–Pacific plate boundary), the wide, sediment-starved continental shelf covering much of Nelson and the Marlborough Sounds, which had seen the accumulation of the countrywide layer of shell limestone 30 myr ago, now became the site of accumulation of voluminous mudstones and sandstones. The plate boundary no longer ran through Nelson, but had shifted eastwards to the Alpine Fault. Basins were forming (e.g. at Murchison; Chapter 20), and increased relief caused by compression at the boundary was providing much more sediment. The east Nelson region contributed some of this sediment, including stones seen today in late Miocene-age conglomerates of the Port Hills Gravel in Nelson city. Finally, around 12 myr ago, all of northwest and east Nelson was lifted above sea-level, and the record of marine sediments ceased. Local mountain-building started later, around 5 myr ago, as it did throughout the South Island.

The Nelson Port Hills Gravel and Moutere Gravel

Nelson has two records of those early mountains. First, there is the deep downfold that deforms the cover strata under Nelson city (Box 22.1). This fold formed initially before 8 myr, which we know because the Port Hills Gravel formation that is 8–4 myr old, overlies already folded strata. The Port Hills Gravel contains some pebbles and cobbles of granite that came from the northwest, but most of the stones are from the ranges to the east of Nelson. In other words, the east Nelson ranges were already growing 8 myr ago.

Much has happened tectonically in the past 8 myr, so it is no surprise to find that the Port Hills Gravel is itself folded into a downfold. Thus Box 22.1 shows that the Nelson Syncline formed in two stages. A moderate fold developed prior to 8 myr, followed by erosion of the rocks; then Port Hills Gravel was deposited on top by rivers across Nelson. This was followed by more compression that folded the gravel and tightened the older fold, and finally yet more erosion.

The second record of early mountain-building is found in the Moutere Gravel formation, aged 4–2 myr. It forms a thick fill in the Moutere Depression (Fig. 22.6), which extends southwards for 70 km from Richmond to Nelson Lakes National Park. This is a huge volume of coarse river gravel, consisting almost entirely of greywacke cobbles derived from erosion of the growing Southern Alps on the southeast side of the Alpine Fault in the vicinity of St Arnaud. The gravel was distributed by a number of braided rivers across a north-sloping piedmont plain that extended far to the east and west of the present-day Moutere Depression. It was thus similar to the Canterbury Plains, running down into western Cook Strait.

Fig. 22.5. The rounded greywacke cobbles and pebbles of the Moutere Gravel Formation were eroded off the rising Nelson Lakes region and transported north by braided rivers to form a coastal plain between 2 and 4 myrs ago. Moutere Bluff, Tasman Bay. Photographer Lloyd Homer, GNS Science.

Because the Moutere Gravel is a permeable formation and have been above sea-level ever since they were deposited, they are quite well weathered. As a result, most of the stones are now pale buff in colour instead of their original grey, and they break apart readily, and even further weathering has produced the distinctive Moutere Clay Soil. The gravel can be seen in many road cuttings along SH6, SH60 and the Motueka Valley Highway, and in cliffs at Moutere Bluff on Tasman Bay. Erosion of the homogenous formation has produced a landscape with a regular herringbone pattern of major and side ridges.

The greywacke stones in the Moutere Gravel came from early mountains on the opposite side of the Alpine Fault. Those rocks have since moved far to the southwest, sliding along the fault and being pushed up into the air. In the past 2 myr, a new regime of mountains and mountain-related features has been instituted.

Nelson Lakes National Park

Fig. 22.6 shows the region around the Nelson Lakes National Park, which is a microcosm of the geology of northern South Island. The Mt Owen massif, largely of Arthur Marble, a fragment of the old Takaka Terrane in the northwest corner of the region, is abutted to the east by Separation Point Granite and overlain to the west by the Miocene-age Murchison Basin. The Median Batholith underlies the Moutere Gravel, which is now preserved in the fault-angle basin of the Moutere Depression, and the southern end of the Dun Mountain Ophiolite Belt meets the Alpine Fault in the Red Hills near St Arnaud. The Alpine Fault curves gently across the map – this is part of the big bend in the fault described in Chapter 15. As a consequence of the bend, which causes increased compression at the fault, the St Arnaud Range is the northernmost greywacke range of the Southern Alps, and the widening area of deeper Alpine Schist seen in Fig. 22.6 marks the rapidly increasing uplift at the fault caused by the bend. Finally, the area was heavily glaciated, giving us the glaciated terrain and the two iconic lakes of the national park, the sources of the Buller River. Glacial moraines occupy a large area between the Buller and Gowan rivers, the latter one of the country's steepest and wildest rivers, flowing from Lake Rotoroa to the Buller.

The East Nelson Ranges and the Moutere Depression–Tasman Bay Depression

As shown in Fig. 22.6 and Fig. 22.7, the gravel-filled Moutere Depression stretches from Nelson Lakes National Park northwards to the coast of Tasman Bay. The first point to emphasize is that the Moutere Depression is younger than the Moutere Gravel. The latter once extended well beyond their present extent, but are now preserved only within the depression, having been eroded away elsewhere.

The second point to make is that the Moutere Depression is an asymmetrical fault-angle depression, of the type depicted in Box 6.6 G, with a reverse fault controlling the fault angle. The fault in this case is the active Waimea–Flaxmore Fault System, which is actually a family of faults rather than a single fracture. If you look southwest from Nelson city, the asymmetry of the depression is clearly evident, with the east Nelson ranges on the left, the steep slope of the fault scarp and, to the west, the general land surface of the depression sloping to the left into the fault angle.

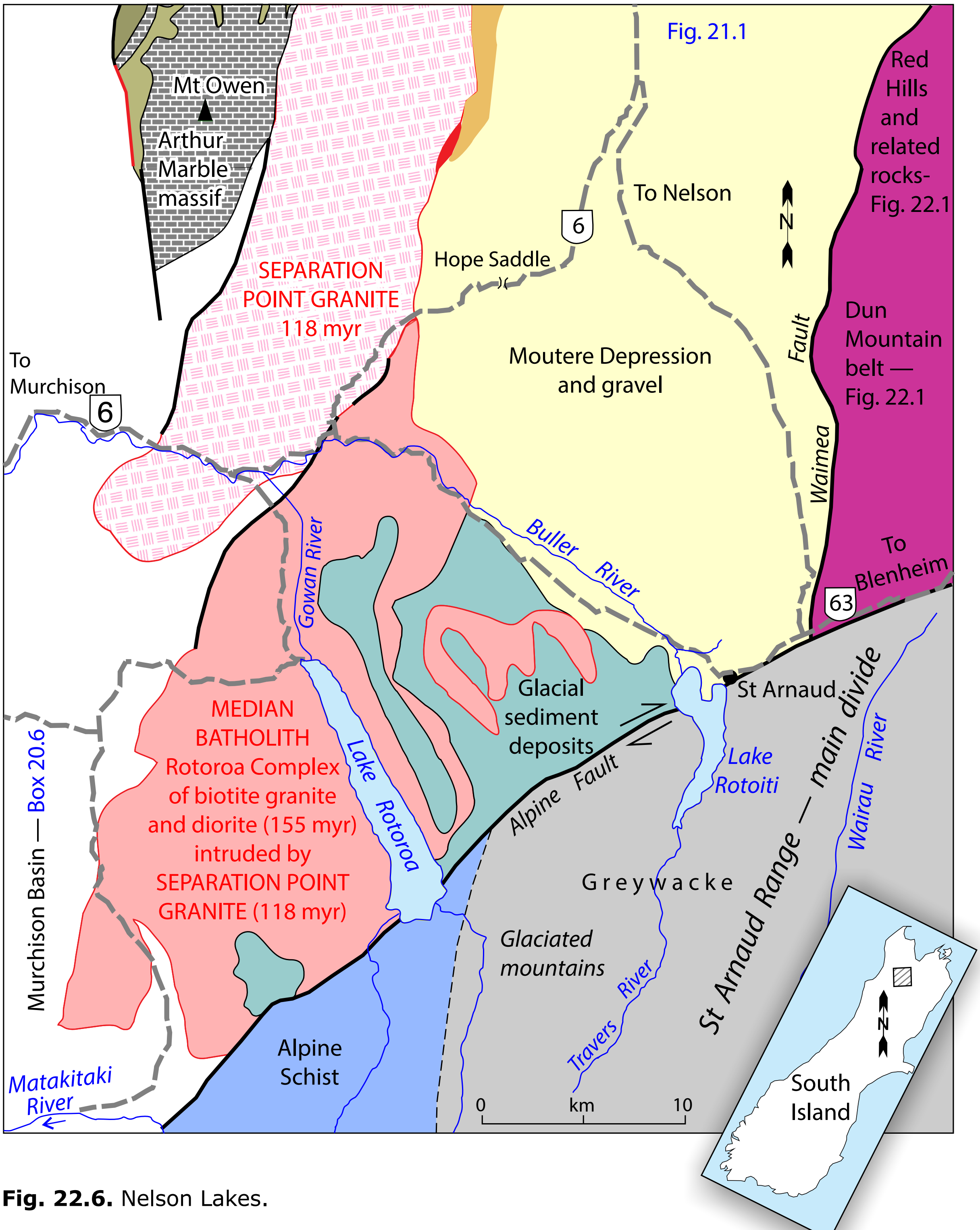


Fig. 22.6. Nelson Lakes.

Box 22.3. Nelson Lakes – discussion.

Lakes Rotoroa and Rotoiti in southern Nelson province, and the high, heavily glaciated mountains to the south of them, are the basis for the Nelson Lakes National Park.

Alpine Fault

The Alpine Fault (Box 19.3) bisects the Park, separating areas of totally different geology (because of the 480 kilometres of right-handed sideways displacement across the fault). Note also that the fault curves across the map — this is part of the big and only bend in the fault, and it is the bend that is responsible for the uplift of the high ranges (up to 2338 m) south of the fault (Box 19.1 A explains how). The accelerated uplift caused by the bend is also responsible for the wide strip of Alpine Schist south of Lake Rotoroa, brought up from great depth, while the fact that the strip tapers to nothing just northeast of Lake Rotoroa is the result of the rapid reduction in uplift as the fault straightens into its Wairau section. The northeastwards reduction in uplift on the south side of the fault also accounts for the fact that the Main Divide crosses the Alpine Fault at the saddle on SH63, a short distance east of St. Arnaud, and continues northward through Red Hill and along the Richmond Range (Fig 22.1), instead of continuing to be associated with the Alpine Fault.

Glaciation

Being high and situated on the Main Divide, the ranges attract high rainfall and have been heavily glaciated. Glaciers drained northward from the high Ella, Mahanga, Travers and St. Arnaud Ranges, which are within the Park, to the lower ground north of the Alpine Fault. This would have occurred during a succession of glaciations, though the landforms we see are largely the product of the latest glaciation which peaked 20,000 years ago. Note how the glacial sediment deposits are located on the lower, northern side of the fault. Lakes Rotoroa and Rotoiti are both glacial in origin, occupying long, straight, narrow, glacially-scoured, U-shaped valleys. Both are dammed by glacial end-moraine deposits (Boxes 17.2 A, 19.4). There are no glaciers in the mountains now, but there are many glacial cirques.

Rock types

Lake Rotoroa is located largely north of the Alpine Fault, within the plutonic igneous rocks (granite and diorite — Box 20.2) of the Rotoroa Complex. These rocks are dated at 155 myr, and are part of the Median Batholith, which separates the western and eastern geological provinces of New Zealand. (Fig 20.1, Box 5.3). Lake Rotoiti lies mostly south of the fault, within Torlesse greywacke rocks (Box 6.2 B), another basement terrane.

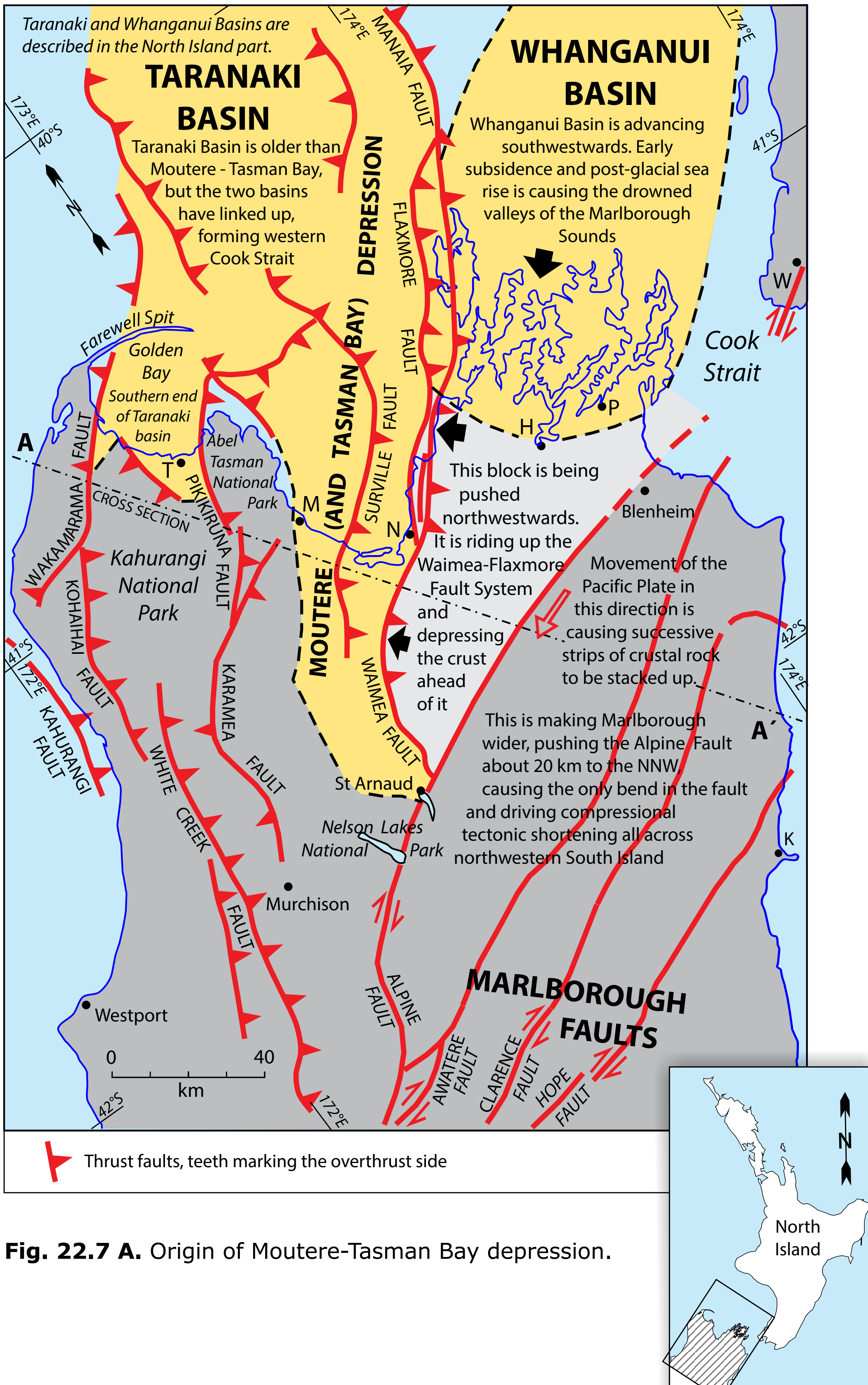
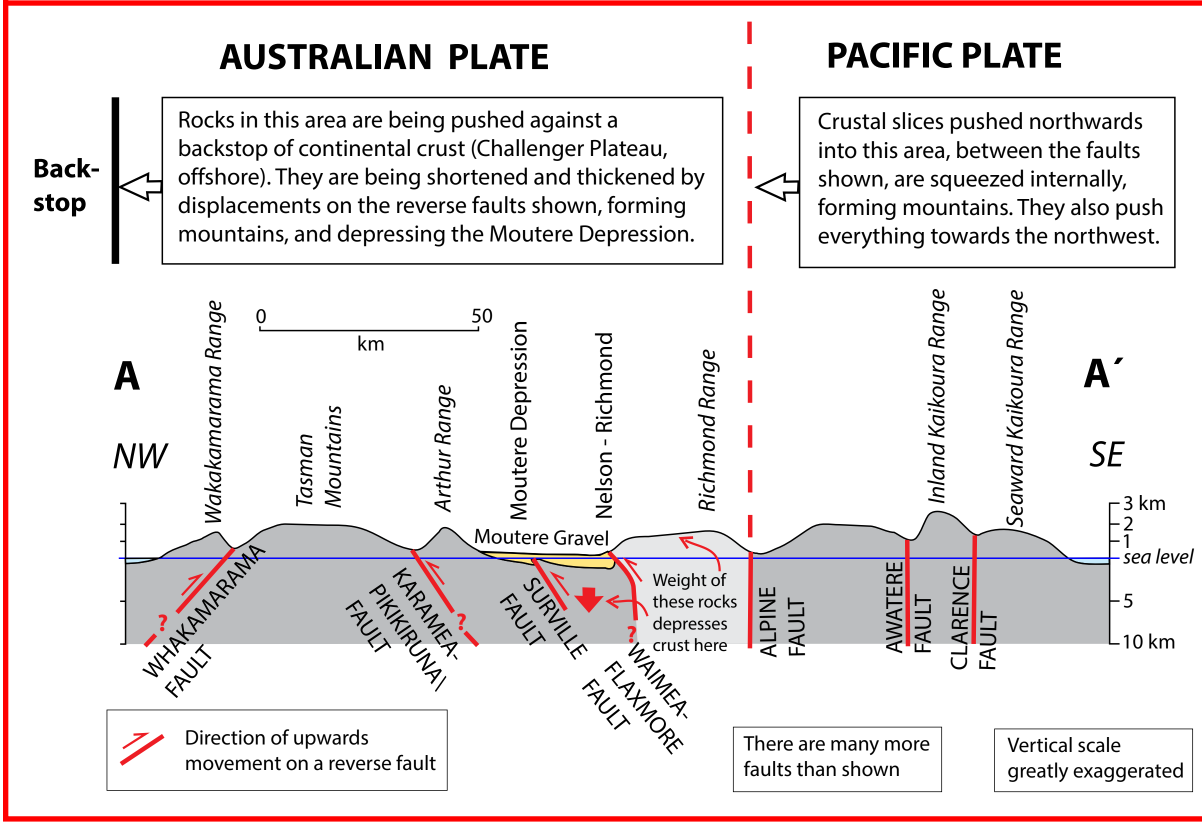


Fig. 22.7 A. Origin of Moutere-Tasman Bay depression.

Box 22.4. Origin of Moutere-Tasman Bay depression – cross section.



The Moutere Depression extends offshore as the Tasman Bay Depression, which is shown in Fig. 22.7 extending to the north past D'Urville Island. The basin system is more than 150 km long, and is New Zealand's only actively forming foreland basin. Such basins form when plate tectonic forces push a mountain range onto and over other continental crust, and the weight of the advancing mountains depresses the crust underneath. Because continental crust and the lithosphere that underlies it are elastic in nature, the depression extends ahead of the mountain front, forming a foreland basin. A good analogy would be a person standing on a plank of wood that is suspended at its two ends. The elastic plank bends under the person's weight, but the bend extends to either side of the person. The bend is deepest beneath the person, and shallows to either side. If the person now shuffles along the plank, the bend moves accordingly. That is what happens as the mountain range is pushed forward.

The foreland basin fills with sediment, as all basins do, but the sediment fill is then overridden by the mountains. This causes compressional (i.e. shortening and thickening) structures in the sediments, such as folds and reverse faults. These effects can be seen in and around Nelson city, in the folding of the cover strata and Port Hills Gravel formation immediately beneath the basal faults of the overriding east Nelson ranges, as noted above.

Foreland basins can be very big. The Rocky Mountains of North America have a foreland basin lying to the east of them (from northern Alberta, through Denver to the Gulf of Mexico), which has been forming for more than 100 myr and is an important oil and gas province. The Himalayan mountains are being pushed southwards onto Pakistan, India and Bangladesh. Their foreland basin forms the wide valleys of the Indus and Ganges and Brahmaputra rivers. Sediment carried through that foreland basin from erosion of the Himalaya is building up the two largest submarine fans in the world, which are filling the northern parts of the Arabian Sea (the Indus Fan) and the Bay of Bengal (Ganges and Brahmaputra Fan), on either side of India.

The Moutere Depression–Tasman Bay Depression is on a smaller scale. The overriding mountains are the east Nelson ranges, which are being pushed northwestwards by tectonic events in Marlborough, as explained in Fig. 22.7, Box 22.4 and Chapter 15, and are overriding the basin along the Waimea–Flaxmore Fault System rather than a single basal thrust. The faults of this system are not low-angle thrust faults, but steeper reverse faults; that is, they are still compressional faults causing shortening and thickening of the rocks. They reach the surface along the base of the ranges, some of them passing through the eastern suburbs of Nelson and Richmond. They are active faults, presenting an earthquake hazard, although earthquakes on them are much less frequent than on the Alpine Fault.

To the north, underneath Cook Strait, the Tasman Bay Depression merges with the older Taranaki Basin, itself a foreland basin formed 25–5 myr ago but now inactive. The bounding Flaxmore Fault merges northwards with the older Manaia Fault. It made mechanical sense for the faults to link up, because they happen to lie on approximately the same line, and are both steep reverse faults separating a foreland basin from its overriding rock mass.

Interestingly, the overriding rock mass of the Manaia Fault has largely disappeared. This is because it has subsided, on account of the southwards advance of the Whanganui Basin, a completely separate and quite different type of basin, which is also swallowing up the east Nelson ranges to form the Marlborough Sounds (chapters 10 and 15).

Because the Moutere Gravel is so thick, rocks underneath them are not exposed, except around Nelson city. They are known only from geophysical surveys and from two oil exploration wells (Tapawera 1 on the Motueka River, and Ruby Bay 1 on the coast of Tasman Bay). From these we know that:

- the basement rocks of the Moutere Depression–Tasman Bay Depression are 'granites' of the Median Batholith;
- cover strata are preserved in two old rift valleys that date from the period of crustal stretching coinciding with phase two of New Zealand's development 100–80 myr ago (Chapter 4);
- the once normal bounding faults of the rift valleys have been reactivated as reverse faults, as is commonly the case elsewhere;
- the main such fault is the Kawatiri or Surville, located beneath the centre of the depression;
- sediments of early parts of the tectonically active period, 25–4 myr ago, are present as river gravels underneath the Moutere Gravel; and
- the maximum depth of the basin on land is 4 km and the maximum thickness of the Moutere Gravel is 2 km.

Glacial Variations in Sea-level

Most recently in Nelson's geological history, ups and downs of sea-level associated with periodic glaciations (Box 10.2 A) have given us the extensive drowned river valley estuaries of Tasman and Golden bays. The last glaciation was at its peak 20,000 years ago, when sea-level was about 130 m below present, and there were glaciers in the surrounding hills. The many associated sand spits (e.g. Farewell Spit; Boxes 14.3 A, B), barrier islands (e.g. Rabbit Island) and boulder banks (e.g. Nelson Boulder Bank – Box 22.2 and Fig. 22.8) have formed since sea-level returned to its present level about 7500 years ago.

Fig. 22.8. 11-km-long Nelson Boulder Bank was mostly formed by southwest longshore transport of gravel from its source around Mackay Bluff when sea level was near its present height in the last 7000 yrs. Photographer Lloyd Homer, GNS Science.

That brings us to the end of our geological excursion around New Zealand. The 17 chapters describing the different geological regions illustrate the huge variety of geological processes that are working, or have been at work, in different parts of the country. New Zealand has always been recognised as a place with a high level of geological activity, but it is only since the advent of the plate tectonic theory in 1968 that geologists have been able to analyse that activity in terms of so many different tectonic and volcanic processes. New Zealand is a veritable living laboratory of geological processes, and I count myself fortunate to have been a professional geologist in New Zealand throughout that whole 40-year period. I look back to my undergraduate and graduate student years in Britain during the 1950s, and remember that geology was remarkably dull then. I hope that readers have obtained some appreciation of the remarkable mobility of planet earth and of the geological splendours of this wonderful country.

FURTHER READING

New Zealand-wide

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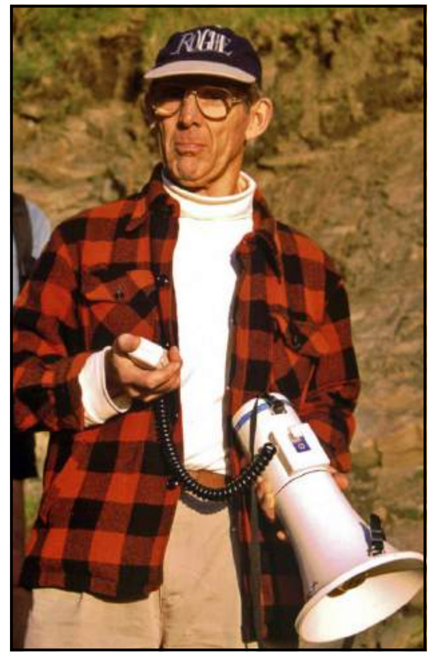
SOME WEBSITES TO VISIT

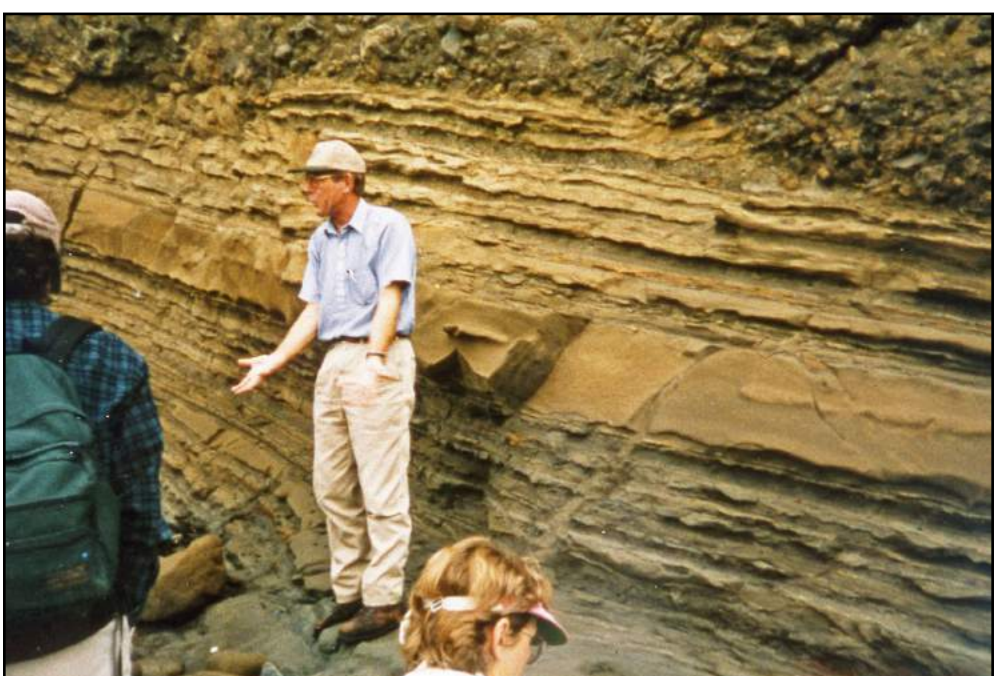
NZ geology	http://www.gsnz.org.nz/information/nz-geology-i-26.html [Click on resources]
Evolution for Teaching (includes New Zealand geological history)	http://sci.waikato.ac.nz/evolution
Geoscience Society of New Zealand	http://www.gsnz.org.nz/
GNS Science	http://www.gns.cri.nz/
National Institute of Water and Atmospheric Research	http://www.niwa.cri.nz/
Rock and Mineral Clubs of New Zealand	http://www.cmlclub.evanta.co.nz/?page_id=33
University of Otago, Department of Geology (Central Otago rail trail and other topics)	http://www.otago.ac.nz/geology/research/environmental-geology/metals-in-the-nz-environment/index.html
Geology field trips	https://www.geotrips.org.nz/
Field guide to Mt Mangere geology	http://www.gsnz.org.nz/information/nz-geology-i-26.html [Click on resources]
Geopreservation Inventory	https://services.main.net.nz/geopreservation/
Wikipedia	https://en.wikipedia.org/wiki/Geology_of_New_Zealand
Te Ara Encyclopedia of New Zealand	https://teara.govt.nz/en/geology-overview

ABOUT THE AUTHOR

Dr Peter Ballance was born and educated in England. Soon after his graduation with a PhD in geology, he took up a position as the first lecturer in Sedimentology at the University of Auckland. Although he and his wife, Queenie, planned to stay in New Zealand just a few years, they loved it so much here that they stayed on. During his time at the University of Auckland, Peter lectured to thousands of undergraduate students and supervised the post-graduate studies of many. He also gave lectures and led field trips for public groups. In his research, Peter soon showed that the Waitemata Sandstones around Auckland were not shallow-water deposits, but were typical deep-water flysch similar to that back home in Europe. In the 1970s, Peter and University colleague Bernhard Spörli joined together to propose the hypothesis of the displaced Northland Allochthon to explain the previously incomprehensible complexity of northern New Zealand geology. After the general acceptance in the 1970s of the plate tectonics theory to explain many of the complex features of world geology, Peter started applying it to explain much of the geology of New Zealand, and this aspect is a major feature of this book. After his retirement in 1998, Peter's major geological project was the preparation of this book but unfortunately, although largely completed, it had not been published by the time of his death in October 2009.



Photographer Bruce Hayward, Ballance Retirement Symposium field trip, 1998



Waiwera. Photographer Ghada Abrahim (PhD Student), 1997

ABOUT THIS eBOOK

This book represents 11 years of retirement research and writing by the late Peter Ballance, prominent New Zealand geoscientist. While on sabbatical leave in the United States, Peter had been impressed with the Roadside Geology guides for various states and decided that he would produce one in similar format with two-colour diagrams for his beloved New Zealand.

He formed a partnership with University of Auckland graphics specialist, Louise Cotterall who, over many years, drafted up the detailed and often complex diagrams that Peter conceived to explain his text. Since his initial draft was completed, the book has had a chequered history. It was accepted for publication as a book by more than one publisher with a succession of editors requesting various changes, at one time dividing it into two separate guides – one for each island, and later another editor putting it back together and requiring Peter to morph it from a roadside guide into a general guide to the geology of New Zealand. This version is virtually unchanged since the format it was in when Peter passed away. It is because of this chopping and changing, however, that there is some duplication between text and boxes as Peter was required to rework the content for various editors. The present compilers have tried to remove the most glaring duplications, but recognise that there is still a good deal remaining. We have tried to be faithful to Peter's concept and have made no significant changes to its text or content and it is published here more or less as it was left in 2009, with just a few updatings of ages and facts that have been obtained since that time.

Soon after Peter's passing, the last publisher declared that the book was not commercial enough for them to publish. Attempts to find an alternative publisher were unsuccessful and the book has remained gathering dust since that time. In 2016, the Geoscience Society of New Zealand decided that rather than let it disappear for ever, they would have it brought together as an eBook and make it widely available on-line for all to use. This last phase of the book required a fair amount of organisation and material-gathering, which included finding 110 photographs from various sources that Peter had placed on his wish list, but never seemed to have sourced. At this time graphics specialist, Louise Cotterall, took the opportunity to fully colourise many of her original diagrams that had been prepared by her in two colours a decade or more earlier.

Bruce Hayward and Jill Kenny

ACKNOWLEDGEMENTS

On behalf of Peter Ballance, his family would like to thank all the New Zealand geologists who helped with this book in many ways over the years, especially staff and students in the Geology Department at the University of Auckland, and fellow members of the Geoscience Society of New Zealand.

Peter's wife, Queenie, was a tireless supporter of this project over the years that he was working on it, and accompanied him on many field trips, on foot and in the car.

Thank you to Susi Bailey for editorial input.

Heartfelt thanks to Bruce Hayward and Jill Kenny for all their work wrestling the manuscript into shape and sourcing photos so the book could finally see the light of day.

And an enormous thank you to Louise Cotterall, from the School of Environment at the University of Auckland, for producing the illustrations during a long collaboration with Peter.

Alison Ballance

The eBook editors would also like to thank:
Louise Cotterall and University of Auckland for updating and fully colourising her original figures and boxes.
Alison Ballance for coordinating Ballance Family wishes for the updated compilation of Peter's project into an eBook.
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Geoscience Society of New Zealand, for their generous sponsorship, which allowed this book to be prepared for publication as an eBook.

Bruce Hayward and Jill Kenny

Illustrating Peter Ballance's Geology 1995–2017

Peter started his anthology of everything he knew about New Zealand and its regional geology using a Mac and hand-drew pencil sketches of his desired maps and diagrams on recycled paper. The illustrator used CorelDraw, Adobe Illustrator and MS Word to translate these files to digital format. When Peter retired in 1998, he continued his treatise by sending his pencil sketch maps by snail mail from his home in Nelson to the illustrator still working in the University of Auckland's Geology Department. Peter's original concept was based on '*The Roadside Geology of Montana (Roadside Geology Series)*' using a two-colour publication - red and black - to reduce the cost of colour printing. This method was used to produce the first iteration. At the request of one prospective publisher, the book project changed shape and was divided into North and South Island sections, with some repetition of geological concepts in each.

Peter obtained sponsorship for his book from several organisations including the Geoscience Society of New Zealand, which became the repository for these funds. Thus, after Peter's death in 2009, the Society continued to seek a commercial publisher without success. In 2016 the Society and Ballance family accepted a proposal to make the book freely available as a downloadable eBook from the GSNZ website. Bruce Hayward coordinated plans, while Jill Kenny undertook to set it out using Adobe InDesign. The illustrator took this opportunity to convert the figures and boxes into full colour.

In reading the guide, one can still hear Peter's voice. His colloquial observations shine through in places dear to his heart, for instance Nelson Boulder Bank (the description of 'heart-shaped rocks' he credited to his observant and artistic daughter, Joyce) and Banks Peninsula volcanoes - quote 'shouldn't be there'.

Because Peter's knowledge was so broad, expansive and enlightening, I am eternally grateful to Bruce Hayward and Jill Kenny for all the work they have put into bringing this project together.

A last word for Peter: *Sedimentology encompasses the study of modern sediments such as sand, silt and clay, and the processes that result in their formation (erosion and weathering), transport, deposition and diagenesis.* Rather like the journey of his guide book. Thanks to Peter for writing it.

Louise Cotterall, Illustrator