Contents

Paul Officer

Most of us yearn for stability and seek wide areas of certainty in our lives, from where we venture out among the risks of our own choosing. Some of these areas of expected certainty are societal: that our money will not become worthless overnight, that the rule of law will continue and so on. Others relate to natural phenomena. Among these assumptions associated with the natural order is the one that the earth will not move beneath our feet.

This issue of *Tephra* explores the subject and would not have been possible without the excellent and authoritative contributions made by scientists from the Institute of Geological & Nuclear Sciences. The research and perspectives of Mark Yetton from the University of Canterbury and photographs and imagery provided by Landcare NZ and NASA have served to complete the publication. My sincere appreciation to all those involved.

Many New Zealanders trace their ancestry back to northern Europe and a large measure of the culture derives from that region. Earthquakes are by no means unknown in Britain, Holland and Germany but it must be admitted they do not form any significant part in the traditions or folklore there. Maori people, however retain memory of the 15th century earthquake, "the earth-swallower", but how many of us have heard of the 14 th century quake in England that brought down the spire of Norwich Cathedral?

For these and other reasons we take it generally for granted that the earth beneath our feet will stay in place, which makes a major earthquake emotionally as well as physically shocking. Yet we need to remember that the impact of a great earthquake could occur anywhere in New Zealand at any time. True, it is less likely to be focused under some spots than under others, but we must not allow ourselves to become blase about the fact. The 1964 earthquake in Alaska caused uplift and major damage over an area larger than the land mass of the South Island.

Our understanding of the causes of earthquakes has been much enhanced over this last generation with the developments in geophysics. We have learned that the land masses of our planet float, as it were, on the material beneath them, and drift, and collide and crumble. Through this endless agitation, material is both created and consumed as the tectonic plates are moved and drawn down by convection currents. This rhythmic cycle is potentially the most destructive natural hazard we face.

New Zealand has given earthquakes very serious consideration. Since the 1931 Hawkes Bay earthquake our building codes have been revised frequently; responding to lessons learned from experiences abroad, and also by applying the results of indigenous research on the subject. In general we believe we have rational cause to expect that contemporary houses (without the tall brick chimneys of earlier generations) as well as commercial buildings will resist earthquake forces better than even 20 years ago.

However, as we cannot expect to escape damage entirely, the sensible risk managing tactic of insurance has not been neglected. The Government's withdrawal from the automatic insuring of commercial risks has led to a more careful examination and to a more market related set of insurance options. The basic anxiety of all New Zealanders as members of households, needing reassurance that there will be a roof over the head again, as soon as practicable, continues to be treated with the seriousness it requires.

Studies in the early years of this decade, and intensified since the Northridge and Kobe earthquakes, have revealed the vulnerability of some of our urban lifelines. In response to these studies the utilities of several cities have been significantly strengthened to ensure if possible that services do not fail, and if they do, to have swift and appropriate remedial measures in place. The recent Auckland experience should offer a warning about some of the difficulties that could be expected, and projecting from the simple loss of electricity, house holders and businesses could imagine some of the other consequences of an earthquake.

As we approach the twenty-first century, the force of the earth movement itself is still beyond our control. Yet we respond to the idea that if we can make ourselves realise that the earth may reel beneath us, and make rational plans to reduce its effects and be ready to respond and recover, one of the most emotionally draining hazards we face may become more tractable.

Some personal reflection

.... As I prepare to leave after seven years as Director it is important to re-affirm that civil defence and emergency management embraces the entire fabric of New Zealand society. It is appropriate also to pay tribute to those whose work has particularly contributed to our safety in the face of emergency threats: to the professionalism of the emergency services and defence forces; to the indispensable input by territorial authority and regional council staff; to the expertise of scientists and technicians from the Crown establishments and Universities; to the contributions of major corporations and the healthy interest of the media; to the support of successive Ministers of Civil Defence, the Department of the Prime Minister and Cabinet, and of other key government agencies; and not least, to the loyalty and commitment of my own staff.

But as important as all the above roles are in times of adversity, the very essence of our system is based on the principle of self reliance and mutual support set on the bedrock of the voluntary sector. In a small isolated country of three and a half million people and limited resources no amount of reassurance or intervention by the authorities can supersede this fundamental fact.

............Within the Government sector we are accustomed to the concept of "the collective interest" - that obligation to think of the total concerns of the executive rather than the narrower ambitions, however commendable, of its component agencies.

However there is beyond that again our interdependence as New Zealanders, aware that all parts of our national community are diminished when any particular area or group is struck by adversity.

.... During my tenure I have been mindful that the statutory duties, functions and powers bestowed on me carry with them an explicit "Duty of Care" which manifests itself in strict accountability. In my view, all the participants in the process should be alert to and accept their moral obligations to exercise an appropriate level of care, whatever their role. This often misunderstood principle goes hand in glove with involvement at any level. Inevitably some things will go wrong and, human nature being what it is, there will be a readiness to engage the "swivelling finger" and apportion blame to someone else. In matters affecting public safety however, everyone must accept their measure of individual accountability. Although my statutory duty of care will end when I retire I shall always retain my personal obligations in this field.

There is always more to write about than space will allow, as indeed there is much more that could be done than funding will ever accommodate. Suffice to say I have enjoyed enormously the many and varied relationships with a splendid cross-section of professional folk. I shall miss the conversations and interplay of minds which have been a fascinating and stimulating feature of those relationships.

P N Officer
Director
MINISTRY OF CIVIL DEFENCE

Why does New Zealand have lots of earthquakes?

by
Martin Reyners

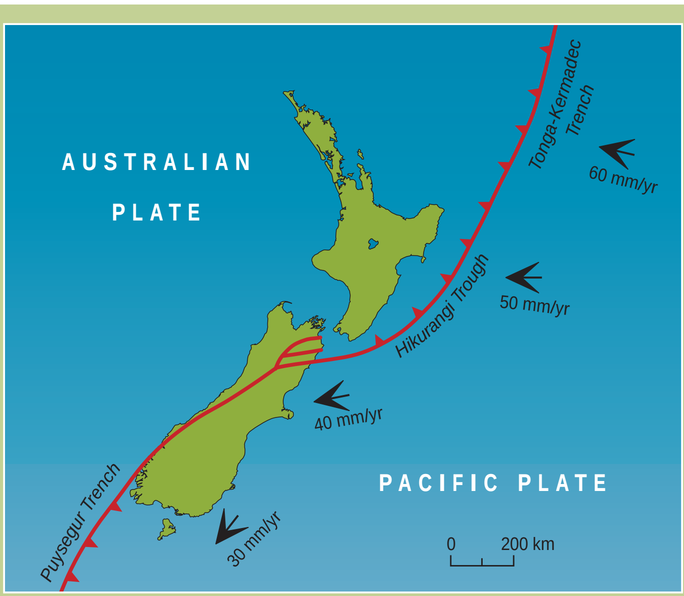


FIGURE 1

New Zealand has many earthquakes because it straddles the boundary between two of the earth's great tectonic plates – the Pacific Plate in the east and the Australian Plate in the west (figure 1). These two plates are converging obliquely at about 40 mm/year in Fiordland and at about 50 mm/year at East Cape.

While these rates are rather small, comparable to the rate your fingernails grow, the plates are about 100 km thick, so a large volume of rock is deformed as the plates collide. This rock can deform in two ways: either through straining elastically and eventually fracturing, which produces earthquakes, or by flowing. The situation is analogous to pushing the two ends of a Mars bar together. While the brittle chocolate layer deforms by fracturing in tiny earthquakes, the caramel interior deforms by flowing.

The convergence of the plates is taken up in differing ways along the plate boundary. In the North Island and the northern South Island, the Pacific Plate dips below the Australian Plate. This process is known as subduction, and earthquakes originating within the subducted Pacific Plate occur as deep as 600 km beneath Taranaki. In Fiordland and the region to the south, the situation is reversed, with the Australian Plate subducting beneath the Pacific Plate. Between these two subduction zones, the crust of both plates is too buoyant to subduct, so the convergence is accommodated by a combination of uplift, which creates the Southern Alps, and movement along the Alpine Fault.

Convergence of the plates is largely driven by the sinking of the relatively dense deeper part of subducted plates into the earth's mantle. This phenomenon of 'slab pull' often leads to tensional stresses in the subducted plate, and many of the earthquakes within New Zealand's subducted plates result from such stresses. Earthquakes also result from bending and tearing stresses which occur as the plates subduct.

If the interface between the subducted and overlying plates were smooth and well lubricated, such earthquakes within the subducted plate would be all that we would experience at our subduction zones. However, this is nearly always not the case, and the plate interface periodically locks up. This has two main consequences. First, the overlying plate deforms as convergence continues, resulting in faulting in the brittle crust. Good examples of such faulting are the Wellington, Wairarapa and Mohaka faults in the overlying plate. And secondly, the plate interface itself may eventually rupture in a large earthquake.

Because of these major tectonic processes operating at our plate boundary, New Zealand will continue to experience many earthquakes. The level of activity is comparable to that in California, but somewhat less than that in Japan. Living in New Zealand thus means living with earthquakes. The studies described in this Tephra are all aimed at quantifying the hazard that these earthquakes pose, so that engineers and planners, and the public at large, can prepare for them.

The *how, what* and *where* of an *earthquake*

by
Warwick Smith

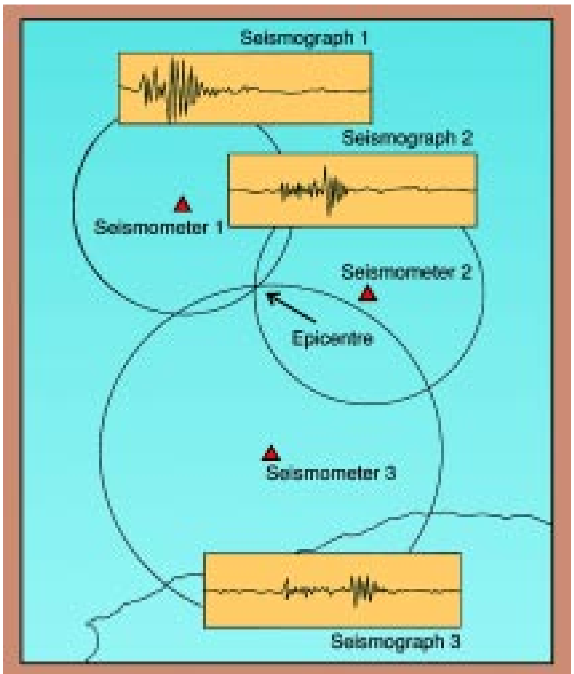
When an earthquake occurs, two things need to be determined quite quickly: the *epicentre* and the *magnitude*.

The actual source of an earthquake, called its *focus*, is usually many kilometres deep in the Earth, where the rock has been strained and has finally reached breaking point. The epicentre is the place on the Earth's surface directly above the earthquake's focus. The energy that is released travels outwards from the focus in waves – and people and buildings are shaken when those waves reach their area.

A seismograph makes a recording of how the ground moved, and from that recording we can tell exactly what time those waves reached the seismograph. Two different types of wave travel outwards from an earthquake at different speeds, rather like lightning and thunder in a thunderstorm. The different times that the two waves arrive at any particular seismograph let scientists know how far away the focus of the earthquake was from the seismograph. As there are several seismographs at different places, each one gives us a measure of the distance to the focus, so scientists can combine all the information to calculate exactly where the focus was, its depth within the Earth, and its nearest point on the Earth's surface – the epicentre.

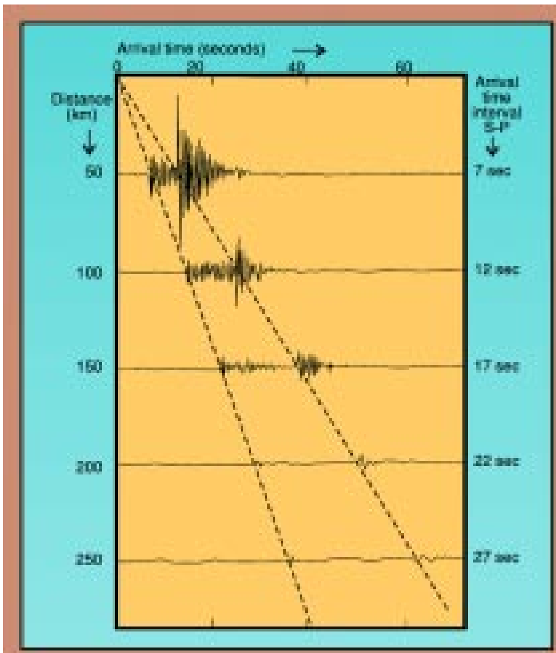
The magnitude number describes how big the earthquake is, not in the sense of how badly you get shaken, but of how big the disruption was within the Earth. If you are near a large earthquake, you will be shaken very strongly. If you are far enough away, you might not feel it at all. To calculate the magnitude, seismologists use a seismograph to measure how much the ground moved, and then they take into account the distance from the focus. This means that no matter what distance you are from the earthquake, you always get the same answer for the magnitude.

The intensity of an earthquake, on the other hand, is its *destructiveness* due to the amount of ground movement at a particular place. Intensity is measured



FINDING THE EPICENTRE. The maximum distance that the epicentre could be from each seismograph is calculated and circles are drawn with that distance as the radius. The circles will intersect at the epicentre.

according to the Modified Mercalli scale which has ten steps for New Zealand earthquakes. The intensity of an earthquake does not directly relate to the amount of energy released in an earthquake (which dictates its magnitude) because intensity decreases with distance from the event, and because natural features, such as rock type, soil type and the amount of water in the area, modify its destructiveness from place to place.



FINDING THE DISTANCE X The time interval between the arrival of the P. wave and the slower S. wave increases with increasing distance from the focus.

The magnitude scale was devised by Professor Charles Richter in 1935 to compare local Californian earthquakes. An earthquake of magnitude 4 is quite small, in fact you have to be quite close in order to feel it. Magnitude 6 is big enough to do quite a lot of damage within a distance of a few kilometres. The biggest that has occurred in New Zealand was of magnitude 8 in 1855. It was felt over almost the whole country and caused a lot of damage in central New Zealand.

It is important to understand that as you go up one step on the magnitude scale, you multiply the size by about 30. So an earthquake of magnitude 5 is 30 times as big as a magnitude 4 earthquake, and a magnitude 6 is 30 times as big again, or 30 times 30 = 900 times. This means that the magnitude 8 earthquake in 1855 was nearly a million times as big as an earthquake of magnitude 4.

Civil Defence needs to know the location of the earthquake and its magnitude quite quickly in order to judge whether the effects are likely to be very severe. If the magnitude is small, or if the epicentre is offshore, it is likely that no damage has been caused. But if the magnitude is large, and especially if the epicentre is near a city of any size, many people could have been affected and Civil Defence may need to be involved.

The Institute of Geological and Nuclear Sciences has developed procedures for providing this information, and expects that with further developments it will soon be able to perform this task more quickly and reliably.

NORTH CANTERBURY EARTHQUAKE 31 AUGUST 1888

Although Christchurch has been little affected by earthquakes this century, its cathedral suffered damage on several occasions last century. The 1888 magnitude 7.1-7.3 earthquake in northern Canterbury, some 100 km from Christchurch, caused partial collapse of the cathedral's spire. This North Canterbury earthquake originated at a shallow depth and ruptured to the surface along the Hope Fault, west of Hanmer Springs. During the earthquake, the two sides of the fault slid past each other with little or no vertical displacement. This type of faulting is called *strike-slip* faulting and recognition of it in 1888 was a world first.

Damage to houses during this earthquake was severe close to the fault rupture. Sand fountains and cracks in the ground (other than the fault crack) were common on the nearby river flats and there were numerous landslides on the surrounding hills.

The fault that ruptured in this earthquake is one of several nearly parallel faults that crosses the North Canterbury and Marlborough areas. These have formed due to the collision between the Pacific and Australian plates.

This fence across the Hope Fault broke during the 1888 North Canterbury earthquake. One part of the fence was displaced about 2.6 m from the other part, and the gap was later wired up as shown in this old photograph.

The *where* and *when* of New Zealand earthquakes

by
Lee Aitken, Terry Webb

Clever science and technology that accurately measure and locate earthquakes are too recent for our scientists to have detailed pattern books for earthquake behaviours in all parts of New Zealand. On top of this, our written history is too short for seismologists to sort out patterns of normally recurring earthquake activity from activity that heralds an earthquake emergency. Identifying the where and when of local earthquakes is absolutely fundamental to planning ways of lessening earthquake disasters in the future.

It is possible that different parts of New Zealand have such different earthquake environments that identifying patterns for some areas may be decades away. And, of course, some areas may not be prone to repetitive earthquake patterns at all.

In the late 1980s, however, the national network of seismographs (instruments that record earthquake waves) was upgraded so that earthquakes could be located more accurately and efficiently. This has confirmed some previous observations and interpretations about where earthquakes originate under different areas of New Zealand. New patterns that had not been previously identified, however, have also shown up. More research is needed to understand what causes these patterns of seismic activity, so that we can better estimate the future hazard posed by earthquakes.

The link between earthquakes and the tectonic-plate boundary that runs diagonally across the country is clearly demonstrated by plotting the depths of earthquakes.

One of the clearest pictures of the role of plate tectonics in New Zealand's earthquake history comes from plotting the depths of earthquakes along a line from Hawke's Bay to Kawhia (figure 1). The earthquakes are relatively shallow in the east, but become progressively deeper towards the west. This almost exactly follows the dip of the Pacific Plate as it dives under the Australian Plate – on which the North Island sits. Earthquakes can occur all along the surface of the diving plate as it travels beneath the North Island into the interior of the planet. Earthquakes more than 600 km deep have been recorded off the Taranaki Coast at which depth the diving plate loses its rigidity and ability to fracture and create earthquakes.

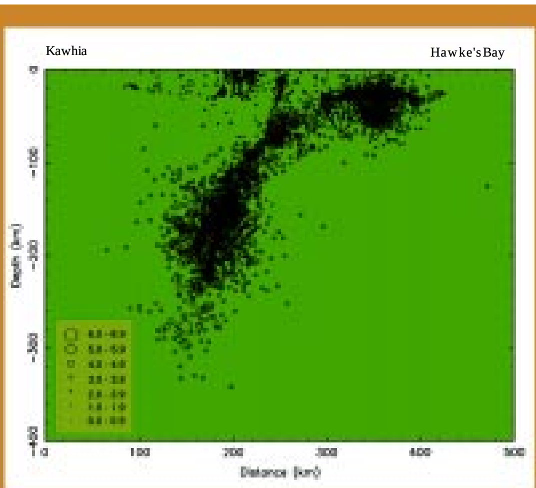


FIGURE 1 Earthquakes under the North Island from Hawke's Bay to Kawhia.

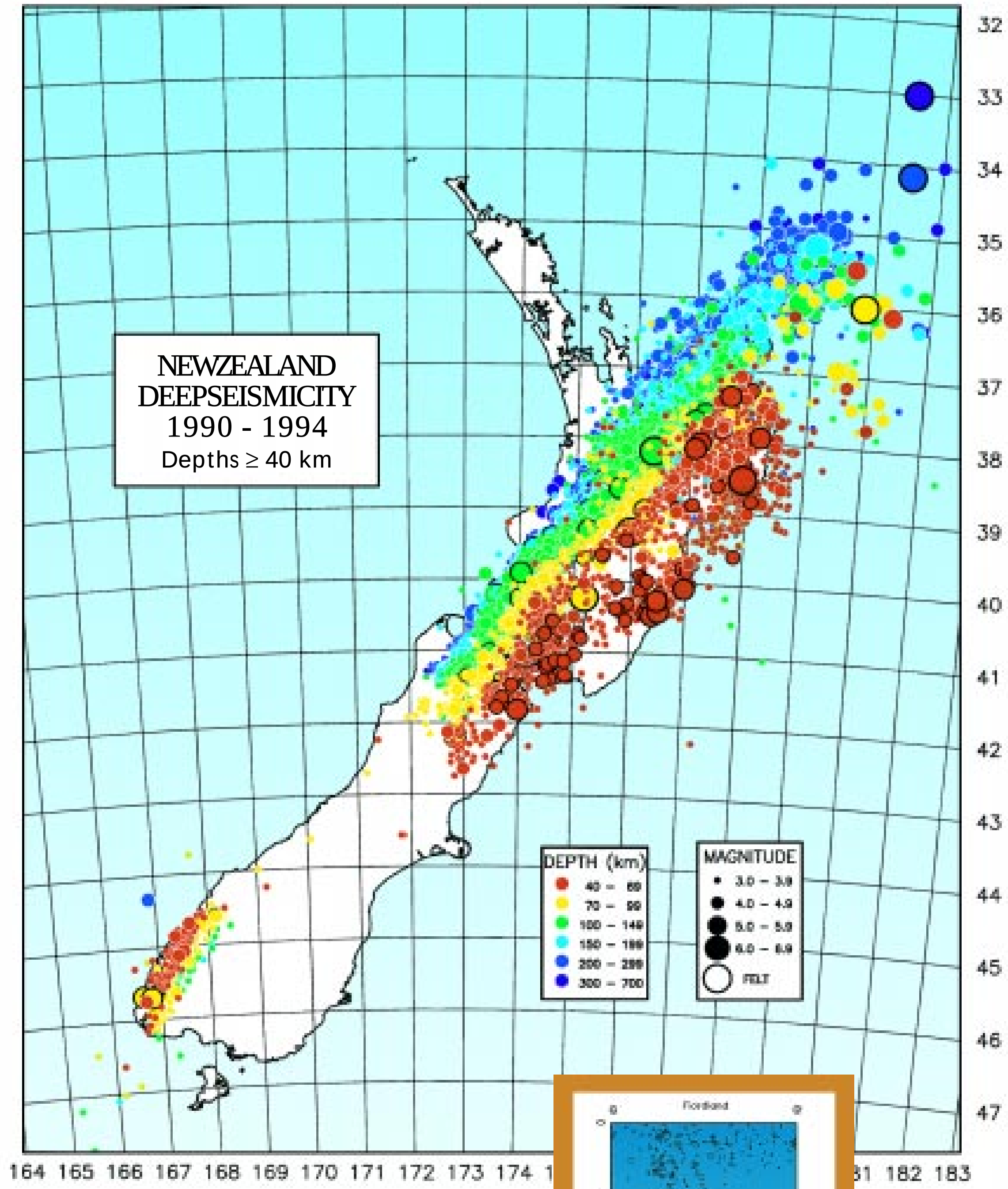


FIGURE 2 New Zealand's deep earthquakes from 1990-1994 become deeper from east to west.

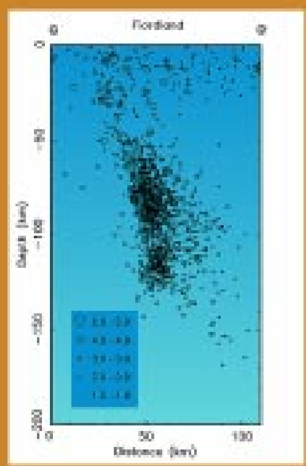


FIGURE 3 Earthquakes beneath Fiordland show that the subduction zone dips towards the east.

The tectonic relationship is also clearly seen in the map of deep earthquakes (figure 2), where under the North Island the earthquakes get deeper to the northwest. The distribution of all depth bands is parallel to the Hikurangi Trough along which the Pacific Plate dives. Deep earthquakes also occur in Fiordland and to the southwest of New Zealand, where the Pacific and Australian plates swap their diving and overriding behaviours. In this area, the Australian Plate

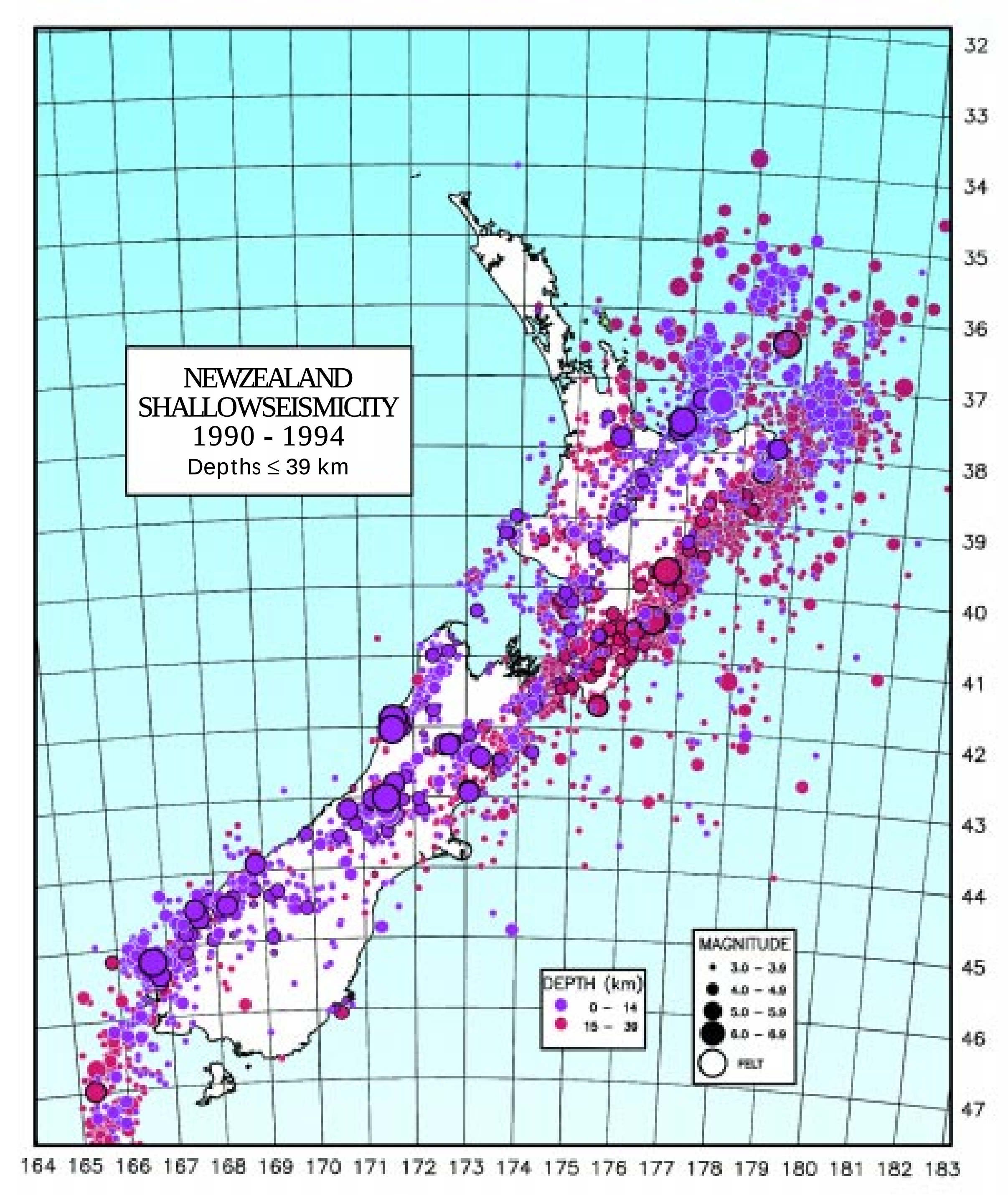


FIGURE 4 New Zealand's shallow earthquakes include those triggered by the crustal deformation and tectonic stresses along the Alpine Fault.

does the diving and the Pacific Plate, topped by south and east of the South Island, is the overriding mass.

This change of the diving and overriding roles is graphically illustrated by the locations of earthquakes under Fiordland (figure 3). Here the earthquakes became deeper towards the east. It can also be seen that the diving angle of the Australian Plate is much steeper than that of the Pacific.

Between the two zones of deep earthquakes, there is gap in deep earthquake activity which coincides with the Alpine Fault. This marks the boundary between the Pacific and Australian plates where the two plates push past each other sideways, rather than one plate diving under the other. They are also pushing together with such force that the Southern Alps are squeezed up along the seam. The sideways

Looking south along the trace of the Alpine Fault, New Zealand's onshore plate boundary. Milford Sound in the far distance.

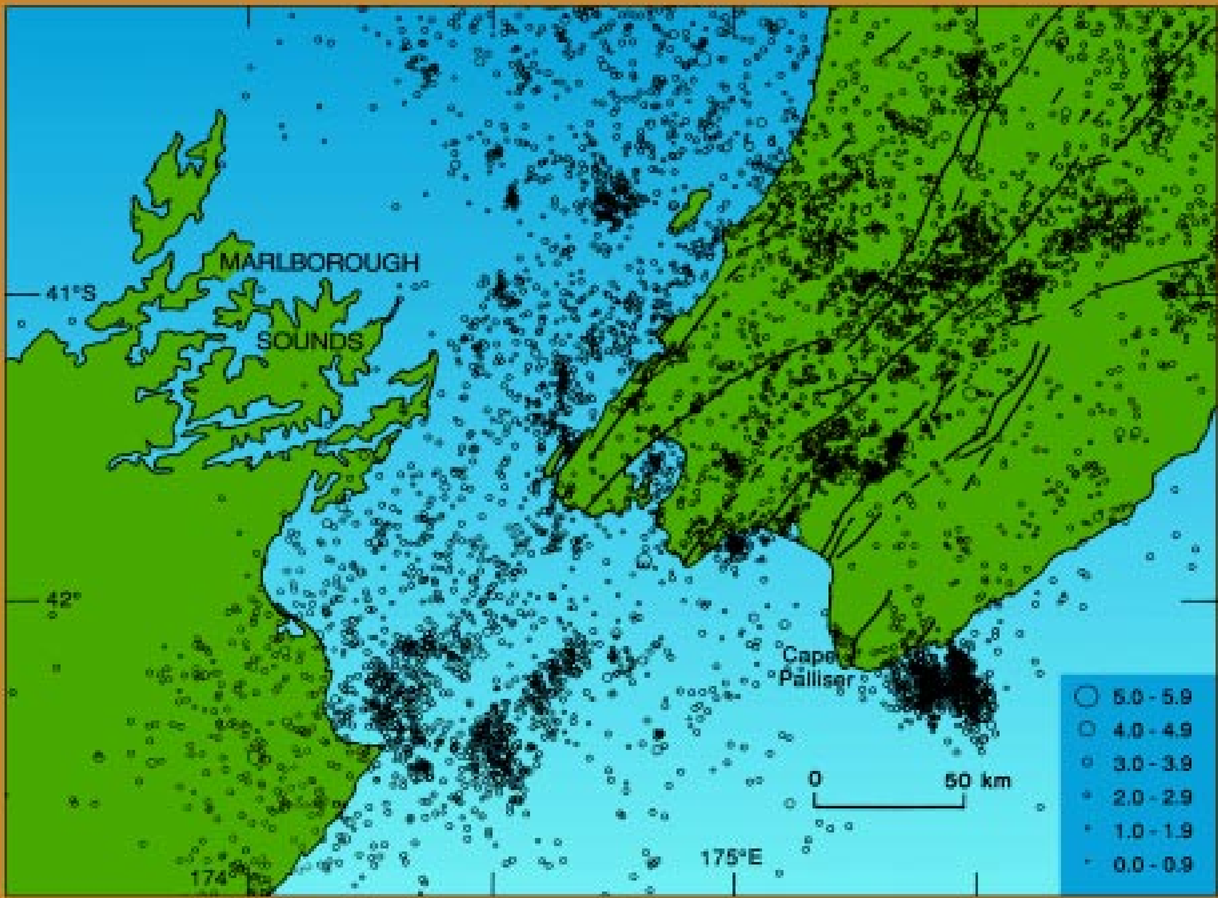


FIGURE 5 Earthquakes shallower than 40km between January 1990 and February 1993.

movement means that parts of Nelson and west Otago that were adjacent to one another five million years ago are now 450 km apart, and that in another 10 million years, Haast on the Australian Plate and Christchurch on the Pacific Plate will be close neighbours.

The lack of deep earthquakes along the Alpine Fault zone is because neither plate is diving below the other. The rocks on both sides of the boundary are too buoyant, and when there is no cold and rigid plate edge descending to great depths, there are no earthquakes generated along the interface.

The Alpine Fault zone is also relatively free from earthquakes between 10 and 40 km deep. This is due to the pressures from the ongoing sideways grind between the plates creating so much heat that the rocks deform by flowing out of shape, rather than fracture by snapping out of shape through earthquakes. The distribution of shallow earthquakes, however, shows a somewhat diffuse pattern (figure 4), so clearly the strain can be released by rocks in the top 10+ km snapping in a brittle fashion.

The map of shallow earthquakes also shows that the colliding plates cause the main earthquake type in New Zealand. A zone of shallow earthquakes along the east coast of the North Island and in Fiordland reflects the subduction boundaries.

Other minor zones lie between Buller and Taranaki, offshore from Wanganui, and in the volcanic zone from Ruapehu north into the Bay of Plenty. There have been earthquakes in Northland (1963 particularly) and eastern Otago (1974), but these places are much safer from earthquakes than the rest of the country.

New Zealand has about 200 earthquakes a year that are big enough to be felt by people nearby, but these are mostly quite small (figure 5). Most earthquakes that are big enough to damage buildings are of magnitude 6+, and we generally only have one of those a year. A magnitude 7 quake occurs about once every ten years, and a magnitude 8 once a century.

Unfortunately, earthquakes do not occur evenly through time: they may happen in clusters. Return-periods are averaged out over many years (with the help of paleoseismologists and scientists in countries with centuries of earthquake records), so they are no cause for complacency. The averaging procedure that declares a magnitude 6+ ‘about’ every year really only means that within a hundred years, there is likely to be about a hundred of them.

How much will it shake where you live?

by
John Taber, Mike Kozuch, Graeme McVerry

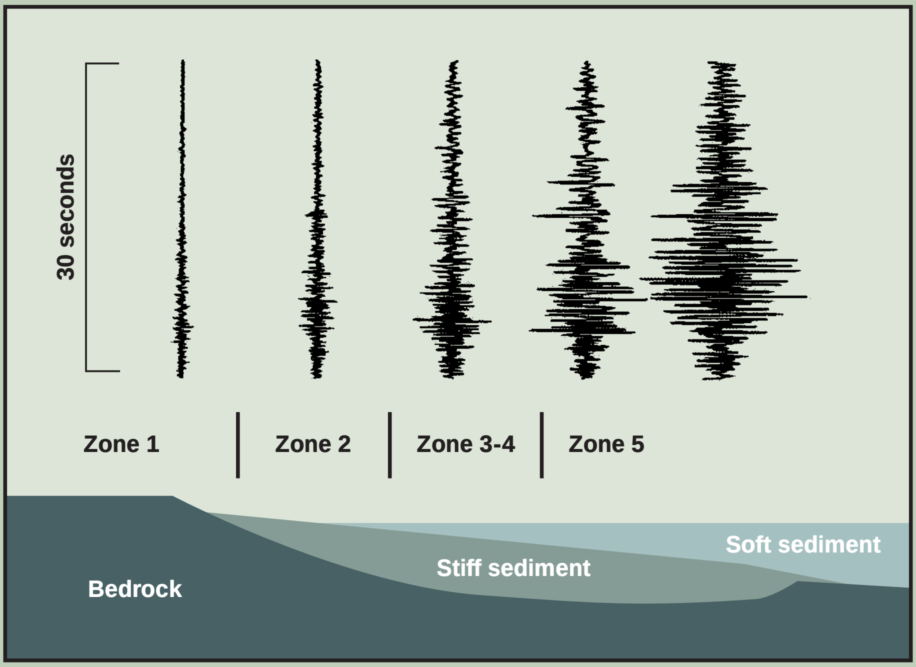


FIGURE 1 Seismograph traces illustrating the influence of underlying rock types on ground shaking

In Mexico City in 1985, more than 8000 people died in an earthquake of magnitude 8.1, which was centred more than 400 km away. In the same earthquake, the damage in Acapulco, 100 km nearer the epicentre, was relatively minor. This earthquake taught us an important lesson which has been repeated many times since: the composition of the ground directly beneath a building can be just as important in determining the amount of shaking the building will experience as the distance from the earthquake.

The question most frequently asked by earthquake engineers is: "You can tell me something about where earthquakes will occur, and how big they will be, but how much will the ground move under this building?" We are now beginning to answer that question, but it takes a detailed knowledge of how seismic waves are generated by earthquakes, and how they travel through the Earth. And it takes a computer analysis to put it all together to make predictions for a particular location.

Damage in an earthquake depends on three factors: how large the ground motion is (the amplitude), how long it lasts (duration), and how rapid (frequency). The larger the amplitude, the greater the force on the building. The longer the duration, the more likely the building is to be damaged because continuous flexing may cause columns to crack and then fail. The frequency determines what size structures are likely to be affected. Tall buildings tend to be damaged by slow vibrations whereas short buildings are vulnerable to rapid motion. The buildings damaged in Mexico City in 1985 were mostly between 10 and 14 storeys tall.

The main reason for the large amount of damage in Mexico City was the fact that the amplitude of the seismic waves was magnified by the ground conditions. The city is founded on a dry lake bed of soft sediments. This caused the seismic waves to slow down as they travelled, and as they did so they built up in amplitude. Not only that, but the particular size and shape of the old lake bed meant that the waves were amplified at a frequency that matched some of the buildings. So it was like pushing a child on a swing: if you push at exactly the right frequency the amplitude becomes large. This had disastrous effects for these buildings.

The geological structure near the surface can have a significant effect on the level of shaking during an earthquake (figure 1). Thick dry sediments can amplify the shaking. Shaking can be amplified in steep hilly areas because the waves are focussed by the shape of the land surface. The study of these effects is known as microzoning.

There have been some clear examples in New Zealand. In the 1942 Masterton earthquake, damage in Wellington was much more severe in locations on soft sediments than in the hill suburbs. But being able to predict the level of shaking in any particular place is not as simple as that. Current research involves measuring ground motions on a variety of ground conditions, identifying the types of waves that are amplified, developing new computer techniques to predict ground motions and testing their results against actual recordings, and comparing ground motions recorded in deep boreholes with those from the surface to understand how seismic waves are amplified as they reach the surface.

The Wellington Regional Council commissioned a study by the Institute of Geological and Nuclear Sciences and Victoria University to produce a set of maps of ground-shaking hazard in selected parts of the Wellington region (figure 2). The set shows which areas can expect the greatest shaking in an earthquake.

If sandy or silty soil is loosely packed and saturated with water, it can behave like a liquid when it is shaken strongly during an earthquake. It loses its strength, so that cars and even buildings can sink into the ground. The soil changes from solid to liquid abruptly, as one eye-witness observed; "... the ground boiled, fissures opened, and the earth shook with violence ..." The most recent occurrence of liquefaction in New Zealand was during the Edgecumbe earthquake in the Bay of Plenty in 1987. In Wellington, low-lying areas near the harbour, particularly the reclaimed land, are most at risk.

FUTURE DIRECTIONS

Our understanding of the factors affecting ground motion is growing rapidly. Geographic Information Systems (GIS) are being used to piece together a vast amount of information, and to combine geological information, topography, recorded ground motions and other data on to maps of land use.

The Wellington Regional Council has an all-hazards map which uses this technology to combine the hazards due to ground shaking, liquefaction potential, fault movement, tsunami, and landslip potential. The map allows city planners and the Wellington Earthquake Lifelines group to identify critical areas that are particularly vulnerable to earthquakes and to establish the importance of lifeline facilities in those areas.

It may also be possible to use this information about areas subject to strong ground motion, at the time of an earthquake. When an earthquake strikes, the seismic waves travel about 35 km in 5 seconds, whereas radio waves are almost instantaneous. It may be possible to develop rapid warning systems so that automated systems shut down gas lines or turn on critical facilities in the few seconds between the beginning of an earthquake and the arrival of the most damaging earthquake waves. Even a few seconds' warning may be enough to reduce the likelihood of fire, or stop high speed trains or lifts, and thus save lives.

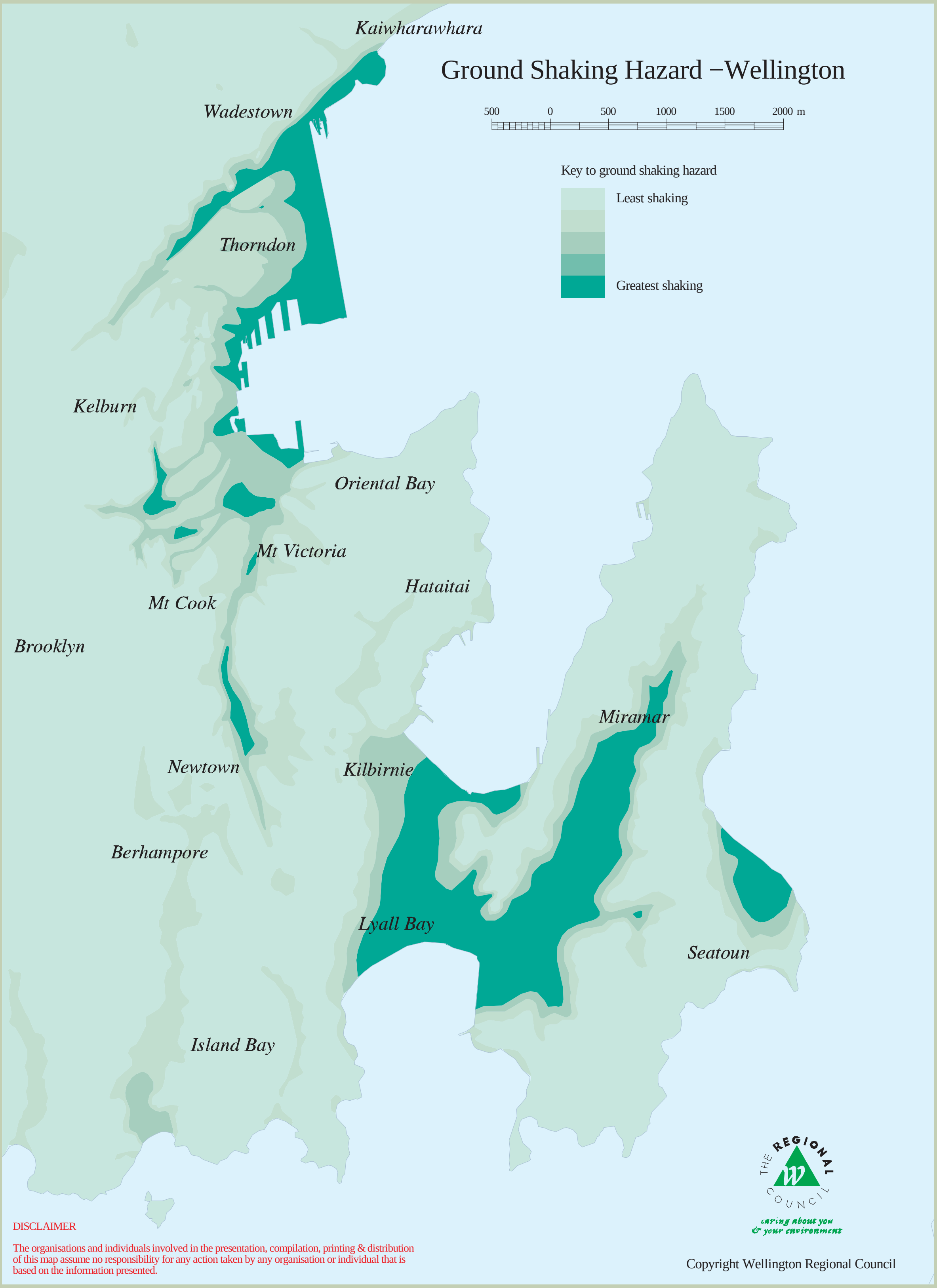


FIGURE 2 Cities in the Wellington region are the first in New Zealand that have been surveyed to assess ground conditions to predict which areas will be prone to the most severe shaking. Soft sediments from reclaimed land and ancient streams and rivers will shake the most severely in an earthquake, and bedrock will shake the least.

The hazard from ground shaking is shown on this map. This gives public planners an overall view of the danger zones in a region.

Landslides and ground damage caused by earthquakes

By
Graham Hancox

Landslides and ground damage are some of the most serious hazards caused by earthquakes. Since 1840, at least 22 earthquakes have resulted in damaging landslides in New Zealand. The damage to buildings and other structures by earthquake-induced landslides is second only to that caused by strong shaking.

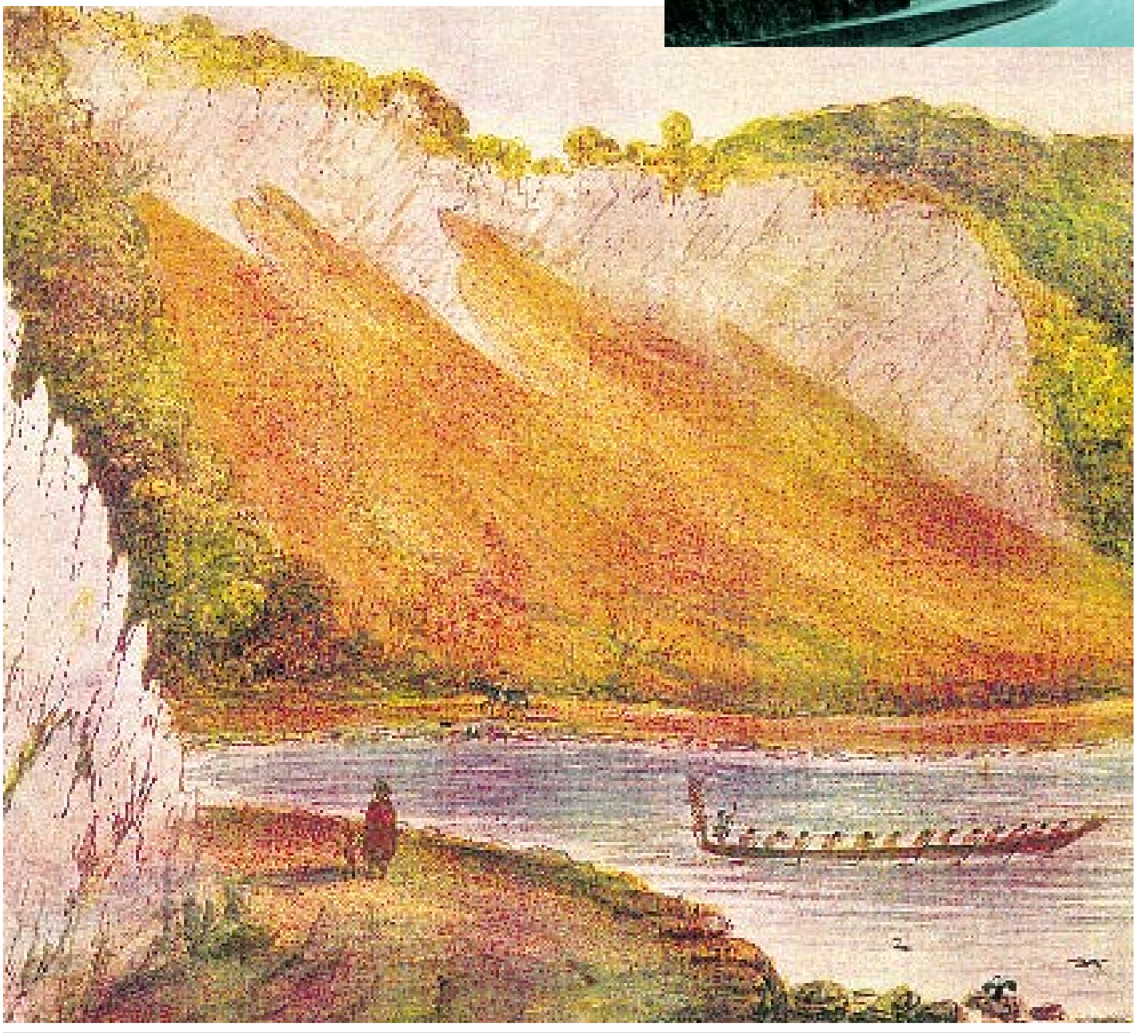


FIGURE 1 Top: A recent photo of the Gold's Slide area, now beside the Hutt Valley-Wellington motorway, demonstrates the vulnerability of main roads and railway lines to earthquake-induced landsliding. Bottom: The water colour Gold's Slide showing the rock avalanche triggered by the 1855 Wairarapa Earthquake (Turnbull Library painting by Charles Gold, 1803-1871). About 300,000 m^3 fell from the 120 m high coastal cliff into Wellington Harbour and blocked the old Wellington-Hutt road.

FIGURE 2 Soil landslide triggered by the 1987 Edgecumbe earthquake on a steep coastal cliff. This landslide is typical of the slope failures that affect roads during moderate and large earthquakes, disrupting communications and emergency relief operations.

Many roads and railway lines have been severely damaged and closed by landslides and subsidence failures during earthquakes (figures 1 and 2), and at least 18 people are known to have been killed. Landslides killed 16 people during the 1929 Murchison earthquake, including two by coal falls in mines, and one person was killed by a rock fall that destroyed a house during the 1968 Inangahua earthquake (figure 3).

In New Zealand, small landslides up to 1000 m^3 may follow magnitude 5 earthquakes, but significant landsliding only occurs at magnitude 6+ events. Earthquakes over magnitude 6.2 that are shallower than 45 km are responsible for most of the rock and soil slides. The shaking intensity on the Modified Mercalli intensity scale that causes landslides is MM6, but the most common shaking intensities for significant landsliding are MM7 and MM8. Very large landslides

FIGURE 3 Rock fall triggered by the 1968 Inangahua earthquake involved many large limestone blocks and destroyed a farm house not far from the cliff, killing one occupant. It illustrates the high hazard posed by steep cliffs adjacent to houses and other buildings. Such sites are best avoided unless slope stabilisation is undertaken.

FIGURE 4 The Falling Mountain landslide involved 72 million m^3 of rock. It is 10 km from Arthur's Pass, and was triggered by the 1929 Arthur's Pass earthquake (magnitude 7.1). The rock fell about 1200 m, and flowed 4.5 km rapidly down the valley trimming off the bush as it went.

FIGURE 5 The Lower Lindsay Landslide, 80 km from the epicentre, involved 2.5 million m^3 of rock and was triggered by the 1929 Murchison earthquake (magnitude 7.8). The scar extends 900 m vertically and 1.3 km laterally, and a small landslide-dammed lake was formed in Lindsay Creek.

(about 1 million m^3 or larger, see figures 4 and 5) occur mainly at MM9 and MM10. In the absence of building damage, landslides may be used to define felt intensity zones for such strong shaking.

Very large landslides in narrow mountain valleys often create landslide-dammed lakes, like those formed on the Karamea River during the 1929 Murchison earthquake (figure 6). The sudden collapse of landslide dams may cause a serious flooding hazard downstream, as occurred south of Karamea after the Murchison earthquake: a landslide dam burst causing a damaging flood.

FIGURE 6 The Garibaldi Landslide, 64 km from the epicentre, was triggered by the Murchison earthquake. Its debris still ponds the Karamea River.

FIGURE 7 Typical sand boil formed during moderate and large earthquakes. Such deposits were ejected from the ground in many places on the Rangitaiki Plains during the 1987 Edgecumbe earthquake. Sand boils are an interesting, but not generally a damaging, soil liquefaction phenomenon.

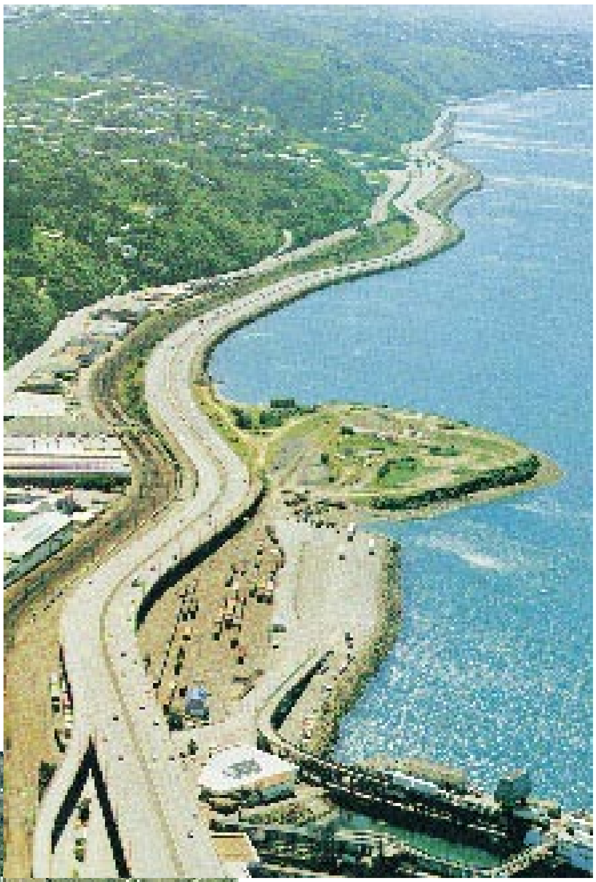
FIGURE 8 Typical lateral spreads involve ground fissures and ejection. These occurred along the banks of the Whakatane River during the Edgecumbe earthquake. The effects shown here are only minor (2-5 cm cracks), but in places they can be much more severe, and include slumping and collapses into rivers, lakes, and harbours. Lateral spreading is a potentially damaging soil liquefaction event that can lead to ground subsidence, flow failures, and severe damage to road and rail embankments, bridges, buildings, and other structures.

Landslide size depends on the magnitude, shaking intensity, and distance from the earthquake source. The maximum landslide area in historical times related to the largest historical earthquake – about 20,000 km^2 was affected by landsliding during New Zealands largest earthquake, the magnitude 8.2 Wairarapa earthquake in 1855. The magnitude 7.8 Murchison earthquake in 1929 caused widespread and damaging landsliding over about 7000 km^2 of northwest Nelson.

Landslides can form up to 300 km from the epicentre of a magnitude 8+ earthquake, but large slides generally occur within 100 km of the epicentre of magnitude 6 and 7 earthquakes, if shaking intensities are MM8 to MM10. The very large landslide in figure 5 is 80 km from the epicentre of the Murchison earthquake and resulted from MM9 shaking on a high mountain ridge.

If other factors are about equal, landsliding in New Zealand is likely to be more severe and widespread during winter when the ground is often wetter. During magnitude 6+ earthquakes, strong or prolonged ground shaking may also cause liquefaction, where sandy waterlogged soils lose strength and flow like a liquid. In appropriate materials, liquefaction effects can include sand boils where sand and water are ejected, and ground fissuring with sand and water ejections along river banks and embankments. Ground damage due to liquefaction is most common at shaking intensities of MM8 to MM10, some 10-100 km from the epicentre.

Sand boils are interesting, but generally not damaging. Ground fissure and lateral spreads, however, cause widespread damage to river banks,



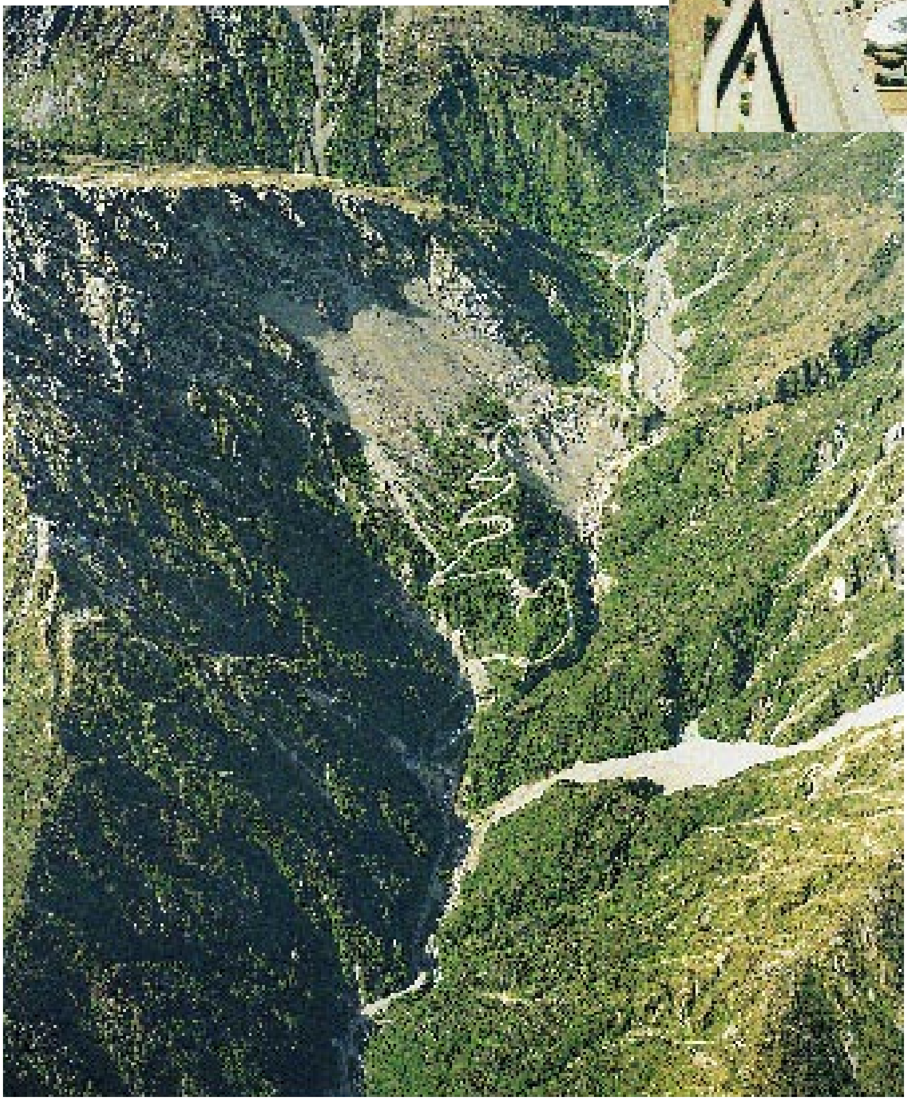


FIGURE 9 Reclaimed areas and lifelines around Wellington Harbour are at risk during a future, nearby, large earthquake. Extensive damage to landfill can be expected from liquefaction-induced flow failures and subsidence. The resulting ground damage is likely to greatly affect both buildings and lifelines such as roads, railway lines, water, power, gas mains, and telephone cables.

FIGURE 10 The Zig-Zag on the Arthur's Pass highway is over a very large prehistoric rock avalanche that gave rise to the debris scree slope and very steep bluffs rising 400-500 m above the road. During the 1994 Arthur's Pass earthquake, the highway was damaged by rock falls on the Otira side of Arthur's Pass, one of which blocked the road and dammed the Otira River for several days. Further very large earthquake-induced landslides are possible in this area, which could severely damage the planned new viaduct and road to be located at the foot of the slope. An alternative option involving a tunnel would have avoided the high hazard zone, but it was considered too expensive.

wharves, bridges, road and rail embankments. Extensive damage can be expected from liquefaction and subsidence in landfill and reclaimed areas. The resulting ground damage is likely to greatly affect both buildings and lifelines of cities, such as roads, railway lines, as well as water, power, gas mains, and telephone lines.

Next to earthquake magnitude and shaking intensity, landslides depend on topography, and rock and soil types, and failures mostly occur on 20-50° slopes. The most common landslides during earthquakes are rock and soil falls on steep cliffs (figures 1 and 3), gravel banks, and high unsupported construction cuts. Rock defects, low strength, and topographic amplification of shaking make things worse. Very large rock avalanches are caused by earthquakes of magnitude 6.5+ on slopes steeper than 25-30° and more than 100-200 m high. Strongly shaken high ridges are particularly susceptible to rapid failure (figures 4 and 5).

Historical earthquake records show that shallow earthquakes that trigger damaging landslides are more likely in the steep hilly areas of northwest Nelson, the central Southern Alps, Fiordland, Marlborough, Wellington, Wairarapa, Hawke's Bay, and East Cape areas. The central North Island, Auckland, Central Otago and Southland are regarded as lower hazard areas because fewer potentially damaging earthquakes are likely in those areas.

Studies of earthquake-induced landsliding are used to assess hazard and risk during future earthquakes in New Zealand. Areas of high to low landslide susceptibility are identified according to combinations of factors such as slope angle and height, weak soil and rock materials, natural slope erosion, and modification by people.

Areas below steep high slopes and unsupported cuts are the most hazardous. Many main roads are flanked by high cliffs that are prone to failure during a large earthquake (figures 1, 2, 9, 10), and numerous buildings in cities are exposed to a similar hazard. The risk to people working or living below steep unsupported slopes is much greater because of the increased time which they are exposed to the hazard.

HAWKE'S BAY EARTHQUAKE 3 FEBRUARY 1931

The Hawke's Bay earthquake (often referred to as the Napier earthquake) caused the largest loss of life and the greatest amount of damage ever recorded in New Zealand. The effects of the earthquake were greatest in Napier and Hastings, but other towns in the Hawke's Bay area suffered major damage also. The official death toll was 256. Fires in the business areas of Napier and Hastings following the earthquake became uncontrollable as water pressure dwindled because of broken water mains.

The earthquake was felt throughout New Zealand, except for the far north and the far south. It was followed by many aftershocks, one at magnitude 7.3. The magnitude of the mainshock was 7.8 and its epicentre was 30 km deep. As a result of the earthquake, a 90-km long dome formed from southwest of Hastings to northeast of the Mohaka River mouth. The maximum height of the dome was 2.7 m. The only surface faulting was a 15-km stretch southwest of the dome.

This earthquake is one of the three largest shocks to have occurred in New Zealand: it is matched in magnitude by the 1929 Buller earthquake and exceeded by the magnitude 8.1-8.2 Wairarapa earthquake in 1855. The Hawke's Bay earthquake demonstrates the potential of large earthquakes near high density population areas to cause many deaths and injuries as well as extensive damage to residential and commercial property.

EARTHQUAKE FORECASTING

by
David Rhoades, Frank Evison, Mike Kozuch

The greatest challenge facing seismology is learning how to forecast major earthquakes. Several countries including China, Japan, Russia, USA, Greece and New Zealand have research programmes with this goal. The New Zealand programme is relatively small, but promising ideas are being developed and tested here.

HAZARD CHANGES WITH TIME

An earthquake forecast is a statement about how earthquake hazard changes with time. Seismic zoning maps give us an average measure of future activity but not the location, magnitude and time of occurrence of individual earthquakes. The aim of earthquake forecasting research is to extend and improve zoning, especially by including the time variable.

PRECURSORS

Many seismologists hold that the build-up process for a major earthquake lasts for a decade or more, so there are plenty of opportunities for observing earthquake precursors if they occur. The problem is whether precursory patterns can be recognised. Many different types of precursors have been proposed. From our understanding of the way the Earth behaves, the two most appealing are patterns of earthquake activity, and changes in the way the Earth's crust deforms.

The New Zealand earthquake catalogue lists all the earthquakes in the New Zealand region. It is one of the better earthquake catalogues in the world, and contains all earthquakes of magnitude 4 and greater during the past 35 years. About two hundred such earthquakes occur every year, and they provide a suitable database for trying to identify precursors to larger earthquakes.

New technology for precise surveying, using global positioning satellites, has great promise for measuring deformation of the Earth's surface. Within a few decades, this database will contain a large amount of precise information on changes in crustal deformation, which may be useful for earthquake forecasting.

Other earthquake precursors are changes in water levels in wells, changes in the chemical composition of groundwater (especially concentrations of radon) and unusual electromagnetic signals in the ground. New networks of measuring instruments, however, would have to be set up to detect them systematically. Meantime, earthquake patterns are the easiest to study, and most of the present research in New Zealand is focused on them. These patterns involve foreshocks, earthquake swarms, magnitude variations, seismic quiet periods, and changes in the rate of occurrence.

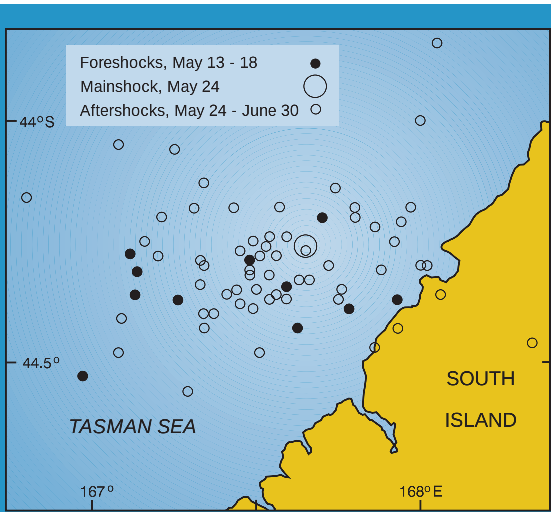


FIGURE 1 Foreshocks leading up to the 1960 magnitude 6.3 Fiordland earthquake. A prominent series of foreshocks was observed before the magnitude 7.3 Haicheng, China, earthquake of 4 February 1975 and was treated as indicating a high probability of a following major event. Only a small proportion of major earthquakes have such a prominent sequence of foreshocks, but a larger proportion (perhaps 25%) have one or more. Their value for forecasting is presently limited because most minor earthquakes are not foreshocks and we have not yet learned how to recognise foreshocks before the mainshock occurs.

Foreshocks are short-term precursors, and precursory swarms are long-term (Figures 1,2 and Box 1). The length of time between the occurrence of the precursor and that of the major earthquake tends to increase with earthquake magnitude for long-term and intermediate-term, but not for short-term precursors.

STAGES OF RESEARCH

The pattern of events leading up to major earthquakes tends to vary from one earthquake to another. Identifying any particular type of precursory pattern and using it for forecasting is a difficult process that involves working through several different research steps.

The first is the anecdotal stage. During this, examples of apparent occurrences of the proposed precursor are noted through studying the sequence of events before particular major earthquakes. After a number of such examples have been gathered, common features are looked for, and rules to define the pattern may be set up.

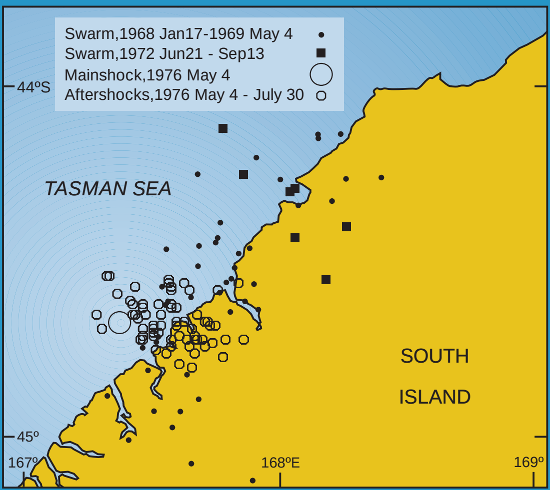


FIGURE 2 Precursory swarms of minor earthquakes before the 1976 magnitude 6.5 Milford Sound mainshock event. Some 17 examples of swarms followed by mainshock events have been identified in systematic studies of the New Zealand and Japan earthquake catalogues. In certain regions of these countries, most mainshocks have been preceded by swarms, and most swarms followed by mainshocks. In volcanic regions, swarms are not precursory to mainshocks, and in continental regions distant from tectonic plate boundaries, mainshocks usually occur without accompanying swarms.

TYPE	TIME-SCALE	EXAMPLES
Long-term	Years or decades	Precursory swarms Crustal deformation Migration of seismic sources
Intermediate -term	Months or years	Seismic quiescence Accelerating seismicity Changes in magnitude distribution
Short-term	Hours or days	Foreshocks Electromagnetic anomalies Fault creep Triggering events Change in radon concentrations Change in well levels

BOX 1 A classification of some proposed earthquake precursors

Systematic correlations of the precursor pattern and major earthquakes mark the beginning of the experimental stage. A method for estimating the variation of the hazard with time is developed, and tested against future events. If such tests show that the new method offers better hazard estimates than the existing one, the operational stage begins and the method is adopted for practical use.

The greatest practical benefits from forecasting will be achieved when several different precursors reach the operational stage. By combining information from several independent types of precursor (for example, a long-, an intermediate-, and a short-term precursor), it is theoretically possible to achieve greater refinements of the hazard, than by using only one. The value of a short-term precursor is much improved when it is used in conjunction with longer-term precursors.

It takes considerable time and effort to advance a proposed precursor through the anecdotal and experimental stages. Many precursor studies are still at the anecdotal stage, and some are well into the experimental stage, but few have yet been developed into methods for estimating the ways that the hazard varies through time. Even fewer have been tested against future data.

One precursor pattern currently being tested in New Zealand is the earthquake swarm. This is perhaps the easiest type of seismicity precursor to recognise. New Zealand and Japanese earthquake catalogues have many examples in which one or more swarms were followed in the long term by one or more mainshocks. An attractive feature of this type of precursor is that the magnitude of the swarm earthquakes can be used to estimate both the magnitude and timing of the largest mainshock. Tests of the swarm precursor have included a successful long-range forecast of the magnitude 7 East Cape earthquake of 5 February 1995. It has still to be demonstrated, however, that the time-varying hazard model based on this precursor will improve earthquake hazard estimation overall, given the possibility of false alarms and of earthquakes that occur without precursors.

FORECASTING AND COUNTERMEASURES

When earthquake forecasting enters the operational stage, a new approach to planning countermeasures becomes possible. Since earthquake hazard is currently viewed as constant through time, countermeasures are maintained at a constant level. They consist of permanent design requirements aimed at controlling the vulnerability of people, structures and lifelines to earthquake hazard, coupled with a preparedness to respond after an earthquake disaster to restore normal life as quickly as possible. With forecasting, a new range of options would be available.

Long-range forecasting shows that the hazard of strong earthquake shaking in any given locality could rise and fall markedly over the course of a few years to a decade. Different localities would face a high level of hazard at different times and, at any given time, the hazard would be concentrated in a few well-defined areas. This presents choices for temporary civil-engineering countermeasures.

- Efforts in strengthening or demolishing old buildings with poor earthquake resistance could be concentrated at any particular time on areas where the forecast indicates a high hazard.
- Temporary strengthening might be a possibility for some buildings until the hazard had subsided.
- New buildings could be designed with provision for elaborate earthquake-resistant devices to be installed temporarily if needed.
- Relief planning and public education could be intensified and a relief organisation could be kept at a high state of readiness in areas of high hazard.
- Home owners would be alerted to reduce the vulnerability of private dwellings, and everybody encouraged to take preparatory measures which mitigate the worst effects of an earthquake.

Intermediate-range forecasts, in which an area's hazard rises and falls over the course of months or a year or two, would encourage similar types of temporary countermeasures, but the shorter time-scale of hazard variation necessitates a faster response. An intermediate-range precursor occurring in a region of already elevated hazard would raise the hazard to a higher level again. This would provide a stronger justification for intensifying countermeasures.

Short-range forecasts, in which an elevated hazard lasts only for a few hours or days in a particular locality, call for a different type of response. The time-frame is too short for most of the measures mentioned above, but other options such as evacuating vulnerable structures, and closing down sensitive equipment would be considered if the hazard became high enough. Such measures require a much higher level of hazard than could possibly be generated by long- or intermediate-range forecasts. A sufficiently high level could perhaps be realised if the short-term precursor occurred when the hazard was already greatly elevated.

The study of earthquake precursors is a difficult one, but progress is being made. A successful outcome will improve estimates of hazard changes over time, and greatly benefit the mitigation of earthquake risk.

Paleoseismo

digging up past earthquakes

by
Russ Van Dissen, Kelvin Berryman

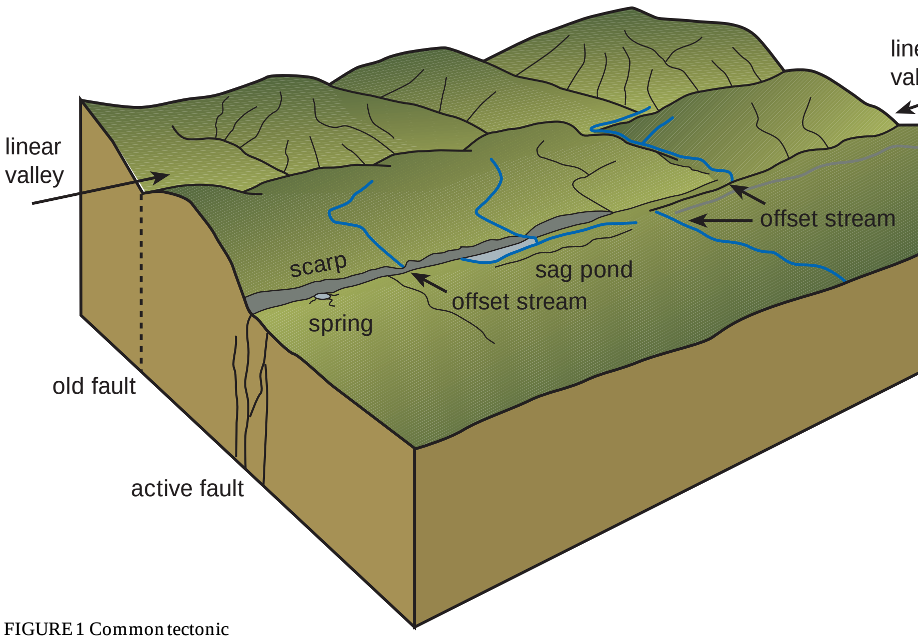


FIGURE 1 Common tectonic landforms along an active, sideways-moving fault.

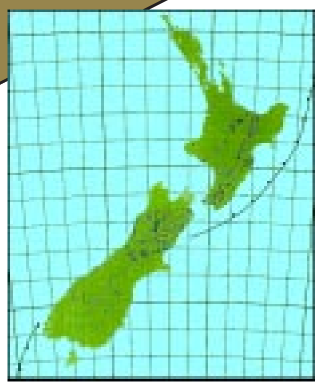


FIGURE 2 New Zealand geological faults that have been active during the last 5000 years.

INTRODUCTION

Paleoseismologists study large, prehistoric earthquakes – their location, size, and timing. The motivation behind these studies is society's need to assess the likelihood and size of future earthquakes. Paleoseismologists interpret geological evidence of ancient earthquakes from various sources, such as natural exposures of active faults, trench excavations dug across faults, offset landforms, and uplifted beaches.

Most of New Zealand's earthquake-generating faults have been quietish since historical records began about 150 years ago, so the only way to learn about previous large earthquakes and their likely recurrence is through the geological fieldwork. Such investigations indicate that the average time between large earthquakes is 3-300 times longer than our written records. Consequently, the earthquake history of active faults has to be detected using the techniques of paleoseismology.

Paleoseismology generally only covers earthquakes greater than magnitude 6.5 because geological evidence of small and moderate-sized earthquakes is rarely created or preserved near the ground surface. Geological evidence of large earthquakes includes:

- local deformation such as fault scarps, sag ponds, and offset stream valleys (figure 1)
- indicators of sudden uplift or subsidence over a broad region over a buried fault (elevated shore lines, and drowned tidal marshes)
- geological indicators of strong ground shaking some distance from the causative fault, such as landslides, liquefaction features, and tsunami deposits.

ology:

ACTIVE FAULTS IN NEW ZEALAND

Tectonic movement of the Pacific and Australian plates is the driving mechanism behind most of New Zealand's earthquakes – the plates move relative to each other at an average rate of 35-45 mm/yr. The strain and deformation of the Earth's crust that leads to earthquakes are concentrated along the boundary zone between these plates.

New Zealand's currently active faults (figure 2) represent the sources of the largest earthquakes that New Zealand has experienced. They have deformed the ground surface within the last 100,000 years, mostly because of movement during large earthquakes.

FIGURE 3 The Long Gully section of the Wellington Fault.

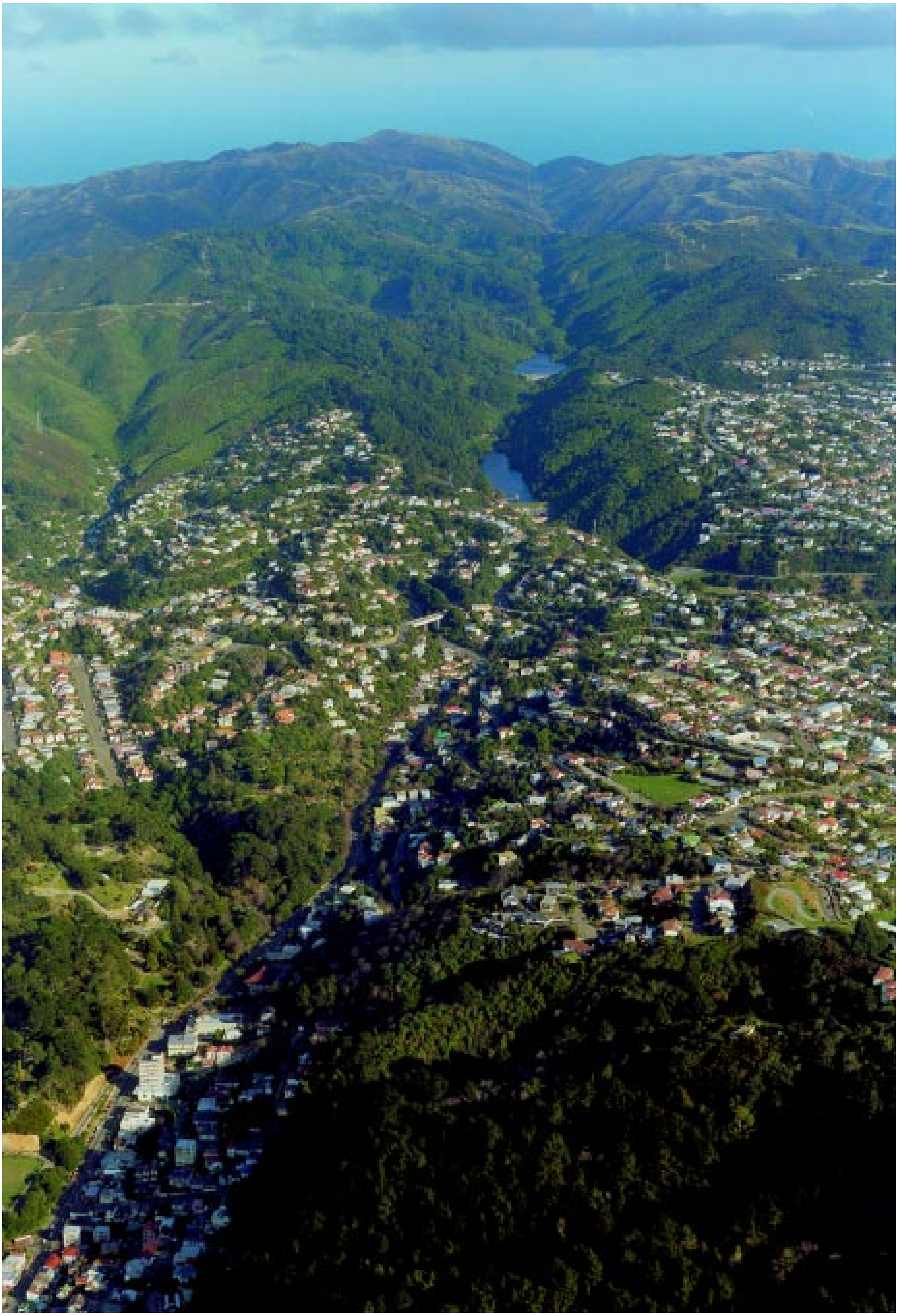


FIGURE 4 The Wellington Fault through the city's suburbs of Kelburn, Northland and Karori. The Karori Reservoirs are on the fault.

The location of an active fault tells where large earthquakes have occurred in the past, but detailed paleoseismic studies are needed to determine when they occurred, and how big they were. This information is then used to estimate the size of future earthquakes on that fault, and to assess how often the fault is capable of generating these earthquakes.

THE WELLINGTON FAULT

The Wellington Fault extends the length of the North Island. It goes through downtown Wellington and presents one of the most serious natural hazard scenarios facing New Zealand. In Long Gully near Wellington's southern coast, streams have been offset across the fault by about 50 m and are ponded behind the fault scarp (figures 3, 4). Paleoseismologists have dug trenches across the fault in the ponded areas to uncover evidence of past surface-rupturing earthquakes along the fault.

In one trench, the youngest geological unit deformed by faulting is 790-930 years old, and it is overlain by unfaulted peat dated at 560-670 years. This demonstrates that the ground surface ruptured there sometime between 560 and 930 years ago, although there may have been other ground-rupturing earthquakes since then whose evidence is not sufficiently preserved.

In another Long Gully trench, an unfaulted deposit of material eroded from the fault scarp buried the exposed fault plane and a tree aged 350-450 years. This probably indicates the time when the burying material was eroded from the fault scarp, and consequently, the age of the earthquake that triggered the ground rupture.

The conclusion from a range of paleoseismological work along the Wellington-Hutt Valley segment of the fault is that the youngest surface-rupturing earthquake occurred 300-450 years ago, and the next youngest about 700-800 years ago. An average recurrence interval for surface rupture earthquakes on this segment seems to be between 400-800 years.

At Te Marua in Upper Hutt, a series of Hutt River terraces less than 10,000 years old record progressive offsets across the Wellington Fault. The smallest offset is 3.7-4.7 m which is taken as the displacement associated with the most recent surface-rupture earthquake at this locality. It is estimated that ground displacements average 4-5 m per earthquake, and that the length of the surface rupture averages 75 km for major events on this section of the Wellington Fault.

Comparing these estimates of surface-rupture length and ground displacement with earthquakes around the world suggests that the Wellington-Hutt Valley segment of the Wellington Fault is capable of producing earthquakes in the order of magnitude 7.5.

FIGURE 5 The northern section of the 50 km long Ostler Fault west of Lake Pukaki.

THE OSTLER FAULT

The 50-km long Ostler Fault in South Canterbury (figure 5) has been investigated for the timing of past fault movements. A trench was dug in a ponded area in the Twizel River headwaters. Two wedges of material eroded from the fault scarp after surface-rupture earthquakes were found. Radiocarbon samples from below and above the older wedge gave ages of 7780-7437 years and 6183-5737 years, respectively. These ages give maximum and minimum constraints on the age of the earthquake that lead to the erosion.

This lower wedge is faulted, so there must have been a younger earthquake – the one that created the upper wedge. Radiocarbon samples from below and above the base of this unfaulted wedge have ages of 4407-3924 years and 3324-2851 years, which bracket the age of the most recent surface fault movement.

Differences in sediment type at a lower level may have resulted from an even older surface rupturing earthquake about 10,000 years ago. The geomorphology of the area around the trench is consistent with three faulting events, and information from the trench excavation indicates that the three most recent surface-rupture earthquakes occurred approximately 10,000, 6000, and 3600 years ago. This suggests a recurrence interval of surface-rupturing earthquakes of about 3000 years.

MARINE TERRACE INVESTIGATIONS

The Mahia Peninsula in northeastern Hawke's Bay is the closest land to the Hikurangi Trough, where

FIGURE 6 View northwest across Mahia Peninsula shows four of the five uplifted marine terraces whose surface offsets and fossil dates provide information for assessing the time and size of prehistoric earthquakes.

the Pacific Plate dives under the North Island (figure 6). Different levels of estuary and beach deposits around the peninsula are raised and deformed landforms that developed because of prehistoric earthquakes. Coastal environments are rich in fossils that can be radiocarbon dated, and five distinct ages have been found for the marine terraces.

Results of paleoseismic studies of these uplifted marine terraces indicate that earthquakes with magnitudes greater than 7.5 occurred about 250, 1600, 1900, 3500 and 4500 years ago. This interpretation is supported by the 3 m of coastal uplift in the Wairarapa that occurred due to the 1855 magnitude 8 earthquake in the area, and the 2.7 m uplift that was associated with the magnitude 7.8 Napier earthquake in 1931.

The 250-year-old earthquake-uplift was relatively small, and paleoseismologists estimate that it will be followed in the not-too-distant future by another uplift event. No earthquakes greater than magnitude 6 have occurred within 40 km of the Mahia Peninsula since historic records began, but if the whole length of the local fault ruptured in a single event, a magnitude 7.5 could be expected.

MODELLINGSEISMICHAZARD

HAWKE'S BAY

A seismic hazard model for Hawke's Bay predicts earthquake intensity (on the Modified Mercalli Intensity scale, MM), and peak ground acceleration for a 500-year return period (figure 7). It is based on the catalogue of earthquakes since 1840, major prehistoric earthquakes inferred from paleoseismology, and geological studies of faults. The model indicates that the variation in strong ground shaking across the region for the 500-year return event may be MM Intensity 8 to greater than 9.

The model uses geologically-determined locations, frequencies, and magnitudes for large earthquakes on active faults, and earthquake statistics from the historical catalogue. The hazard is assessed by integrating rates and strengths of earthquake ground motion from all seismic sources. Characteristics of 40 active fault or fault segments in the Hawke's Bay region were used to constrain the size and frequency of major earthquakes.

The behaviour of the subducting Pacific Plate in locking and storing energy that is released in a major earthquake is allowed for by the model, which assumes that at least part of the plate movement will be released by major subduction earthquakes. The most likely subduction earthquakes are of magnitude 8.1, and they recur every 2500 years. Within a 500-year return period, however, parts of the Hawke's Bay near 'fast-moving' active faults can expect an earthquake with a damaging intensity greater than 9 on the Modified Mercalli scale.

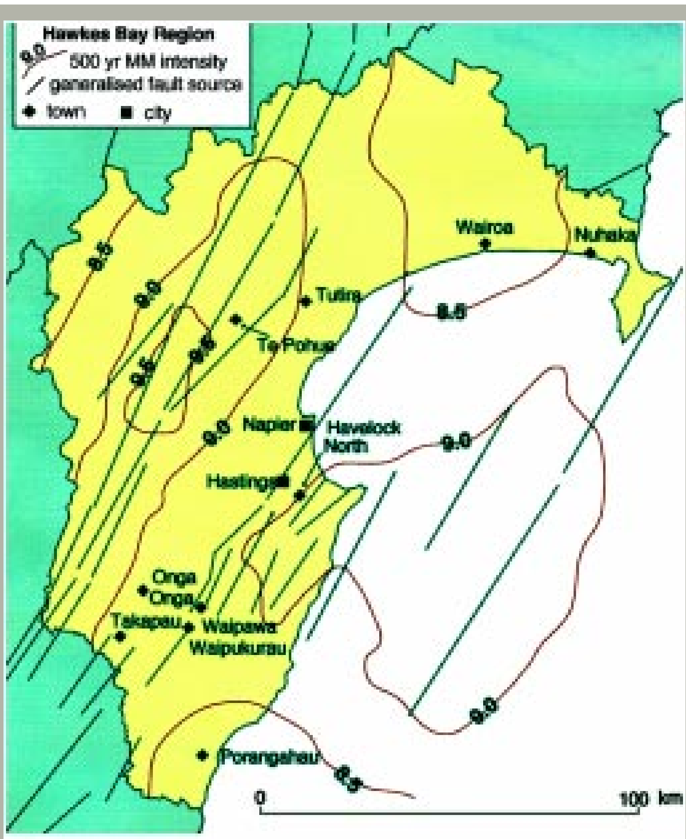


FIGURE 7 The predicted distribution of intensity for the earthquake that occurs once in every 475 years on average. There is a 10% chance that this earthquake will occur within the next 50 years.

CONCLUSIONS

Results of paleoseismology research show that major geological faults produce large earthquakes at time intervals measured in hundreds and thousands of years. Consequently, historical measures of seismicity and ground deformation are too short to assess earthquake hazards, and paleoseismology is vital to estimating the hazard in any particular region.

This practical research is applied around New Zealand by various organisations:

- local, regional and territorial authorities who are required by the Resource Management Act to understand hazards to which their regions are vulnerable
- major utility companies, such as ECNZ and Contact Energy, who need to know the exposure of their businesses to earthquake risk
- insurance and reinsurance industries who need to establish client premiums and know their exposure to claims.

The need for Public Good Science Funding to support the underlying research is recognised by the increasing levels of funding awarded by the Foundation for Research Science and Technology for research in this area since Foundation funding started in 1990.

THE ALPINE FAULT

Mark Yetton is a director of Geotech Consulting Limited, a private consulting and research company based in Christchurch. He has completed this work while enrolled for doctoral study at the University of Canterbury Natural Hazards Research Centre (Department of Geological Sciences). Andrew Wells, a doctoral student in the Plant Science Department at Lincoln University, carried out the forest and tree analysis. Nick Traylen (Geotech Consulting Ltd) assisted with the probability methodology.

by
Mark Yetton

This article arises from an Earthquake Commission Report 95/193. This three year project has been funded by the Earthquake Commission, the New Zealand National Society of Earthquake Engineering, and numerous local authority and infrastructure providers. It evaluates the probability and consequences of a future earthquake on the Alpine Fault in the central South Island.

The Alpine Fault is the largest active fault in New Zealand and extends over 650 kilometers from Milford Sound to Blenhiem. The Southern Alps are a consequence of uplift on the fault, but by far the greatest component of fault movement is horizontal, with an estimated offset of matching strata of around 470 kilometres. The evidence suggests the offset is episodic and each movement of several metres is accompanied by a large earthquake.

The most active part of the fault is the central section forming the western boundary of the Southern Alps from Haast to the Taramakau River at Inchbonnie. Further north the fault becomes progressively less active as movement is transferred to the numerous faults within Marlborough. This project has focussed on the seismic hazard associated with the central and north section, while other researchers from Otago University and IGNS are currently investigating areas further south.

To evaluate the probability of a future earthquake the history of past earthquakes must first be established. This has been done by a combination of four methods, many of which have been applied to the Alpine Fault for the first time.

The first and most direct method is the excavation of trenches and pits across the most recent area of fault rupture. By defining and dating older sheared strata, and overlying younger post earthquake sediments, the timing of past fault ruptures and associated earthquakes can be estimated. Dating requires the presence of organic material to allow the use of 14 C radiocarbon methods but fortunately organic material is relatively common in the forested areas of Westland. However the resolution possible with radiocarbon dating is limited and the timing of the last earthquakes can only be estimated approximately.

The other three methods applied are made possible because previous earthquakes in rugged forested terrain in New Zealand and overseas have demonstrated the profound effects of earthquakes on forests and slopes in the epicentral area. Earthquakes may;

(1) trigger landslides on sloping ground,
(2) cause liquefaction of alluvial areas, and
(3) shake the trees until some fall.

Some of the landslides can be directly dated from radiocarbon dates of buried logs. In addition following a large earthquake new forest will simultaneously re-establish in the clear areas leaving a potential record of the timing of the disturbance in the age structure of the forest. Forest age can be estimated by carrying out ring counts on large numbers of living trees and combining the data to establish modes of forest age.

Some trees also survive the earthquake but still suffer root damage, broken branches and tilting. This is often recorded in their growth rings which potentially provide a very accurate way to estimate the timing of earthquake disturbance.

All four of these methods have been applied to the Alpine Fault between the Paringa River (south of Mount Cook) and the Rahu Saddle near Reefton. They produce a consistent record from which we infer two recent earthquakes on the Alpine Fault in the last 500 years. Figure 1 summarises the data for the two most recent earthquakes.

The most recent event appears to have taken place in 1717 AD, and the rupture extended in length from Milford Sound to the Haupiri River, a distance of at least 375 kilometres. Approximately 100 years earlier, at around 1620 AD, another earthquake occurred in the north section of the fault and extended at least as far south as the Paringa River. Prior to this, another earthquake at around 1450 AD is suggested by the data, but this has yet to be recognised in trenches.

The implied pattern of earthquake recurrence is not regular, but averages around 200 years and varies from 100 years to at least 280 years, which is the lapsed time since the most recent inferred earthquake.

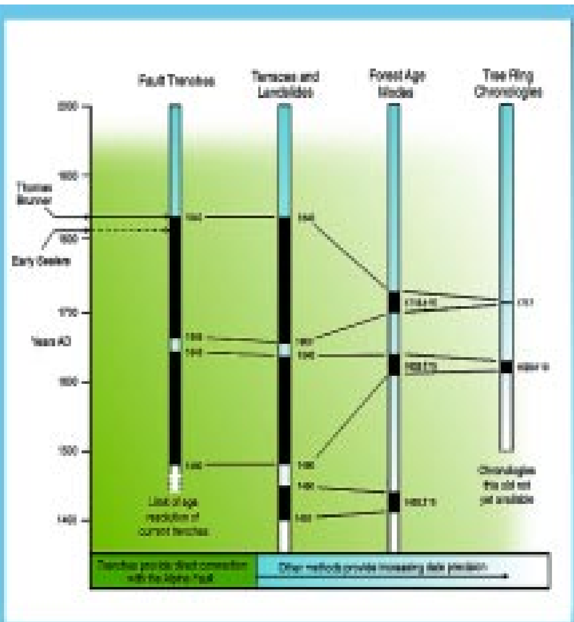


FIGURE 1 Summary of the four methods used to establish the timing and extent of the last two Alpine Fault earthquakes.

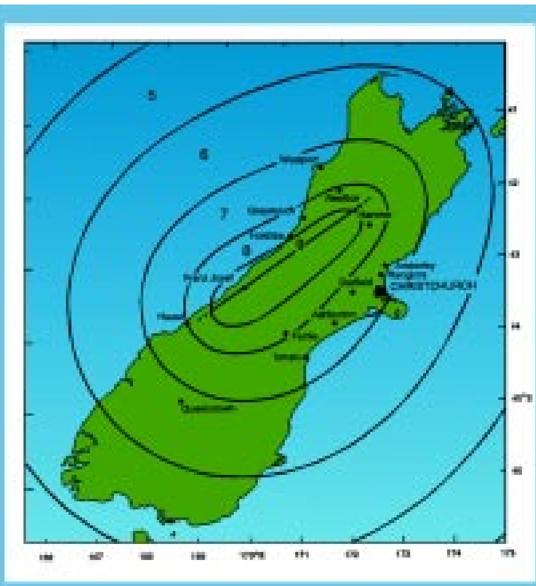


FIGURE 2 Estimated Modified Mercalli Intensity isoseismals (lines defining equal shaking intensity) for the Alpine Fault earthquake around 1620 AD. Modelled using the method of Smith (1995a & 1995b) and reproduced courtesy of Warwick Smith, Seismological Observatory (pers comm., 1997).

Probability estimates can be made using the record of earthquake recurrence derived from a combined analysis of earthquake timing on other plate boundary faults around the world. Other faults also exhibit a wide range in recurrence behaviour, and probabilty estimates are model sensitive, but for the Alpine Fault the probability estimates of the next earthquake are consistently high. A commonly used method suggests a 50 year probability of 65 ± 15% and a 100 year probability of 85± 10% (refer Table 1)

YEARS HENCE FROM 1998	AVERAGE PROBABILITY	RANGE
30 years	45%	30 - 60%
50 years	65%	50 - 75%
100 years	85%	75 - 95%

TABLE 1 Probability estimates for the next Alpine Fault earthquake on the central section of the Alpine Fault using an updated version of the method of Nishenko & Buland (1987).

Based on the rupture length, we estimate both of the most recent earthquakes were around Magnitude 8 on the Richter Scale and reconstructions can be made of the most likely pattern of earthquake shaking intensity. Those earthquakes which also rupture the more northern portion of the fault, like the one around 1620 AD, have generally more impact on the main population centres and Figure 2 shows the estimated shaking pattern.

The next Alpine Fault earthquake is likely to produce very strong shaking in locations close to the Southern Alps. In particular locations such as Arthurs Pass, Otira, Mount Cook and Franz Josef will be seriously

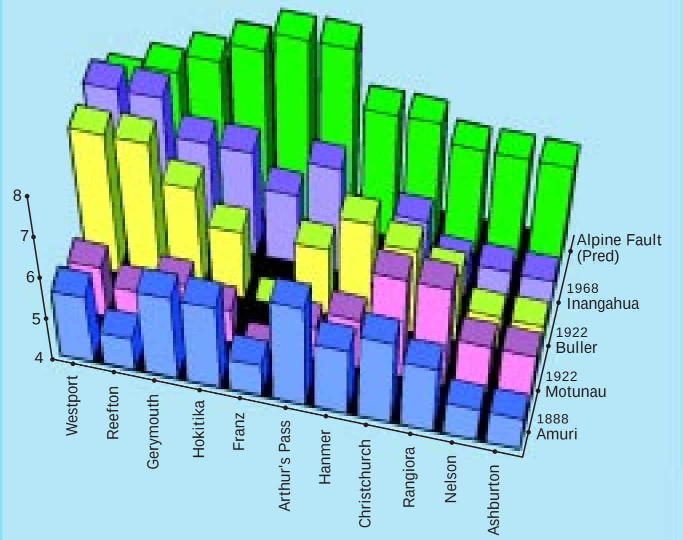


FIGURE 3 A comparison of predicted shaking intensifies for the next Alpine Fault earthquake with those experienced in other large earthquakes this century for a range of locations. The vertical scale shown here from 4-9 is the Modified Mercalli Intensity scale.

affected. Hokitika and Greymouth will also be strongly shaken. Predicted intensifies are generally less on the east coast but in virtually all central South Island locations the next Alpine Fault earthquake will be stronger than any other earthquake experienced there in the last 100 years. Figure 3 summarises the predicted intensifies and compares these to other recent earthquakes.

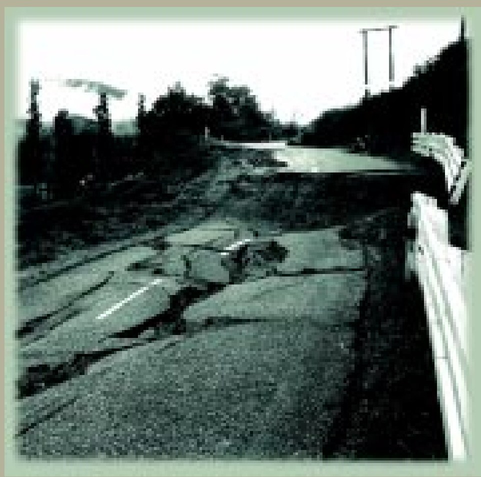
Direct effects of the next earthquake will include landslides and liquefaction. Landslides will be most severe in and around the Southern Alps. It is likely some large rock and debris avalanches will be triggered but the majority of landslides will be relatively shallow failures of weathered soil and rock. Temporary landslide dams are likely to be created. Landslides will also be triggered in sloping ground in Greymouth and the east coast foothills. They are unlikely to be serious in locations as far away as the Port Hills of Christchurch however the greater density of housing in this area may still result in significant property damage.

In Christchurch liquefaction is likely to cause more damage than landslides, because the city has susceptible soils, and it is well within the likely range of liquefaction. Liquefaction will also be widespread in Westland.

One of the most profound long term impacts will be to the river regimes of catchments which drain the Southern Alps. Increased sediment load from landslide material entering the rivers will result in aggradation and channel shifting, particularly in the upper catchments. This has implications for river control, bridging and hydro-electric generation.

INANGAHUA EARTHQUAKE 24 MAY 1968

The Inangahua earthquake is the largest onshore earthquake to have occurred in New Zealand in the last thirty years. Originating at a depth of about 15 km, this magnitude 7.1 earthquake was felt over much of New Zealand, and three people died. Near the epicentre, some 15 km north of Inangahua, most structures were damaged, and damage to bridges, railway lines and underground pipes was extensive. Of particular note were large landslides, one being responsible for two deaths and another temporarily blocking the Buller River. Chimneys were damaged up to 150 km away.



The earthquake ruptured up to the earth's surface in three places, but none of the fault traces, where the ground surface was broken, was more than 2-3 km long.

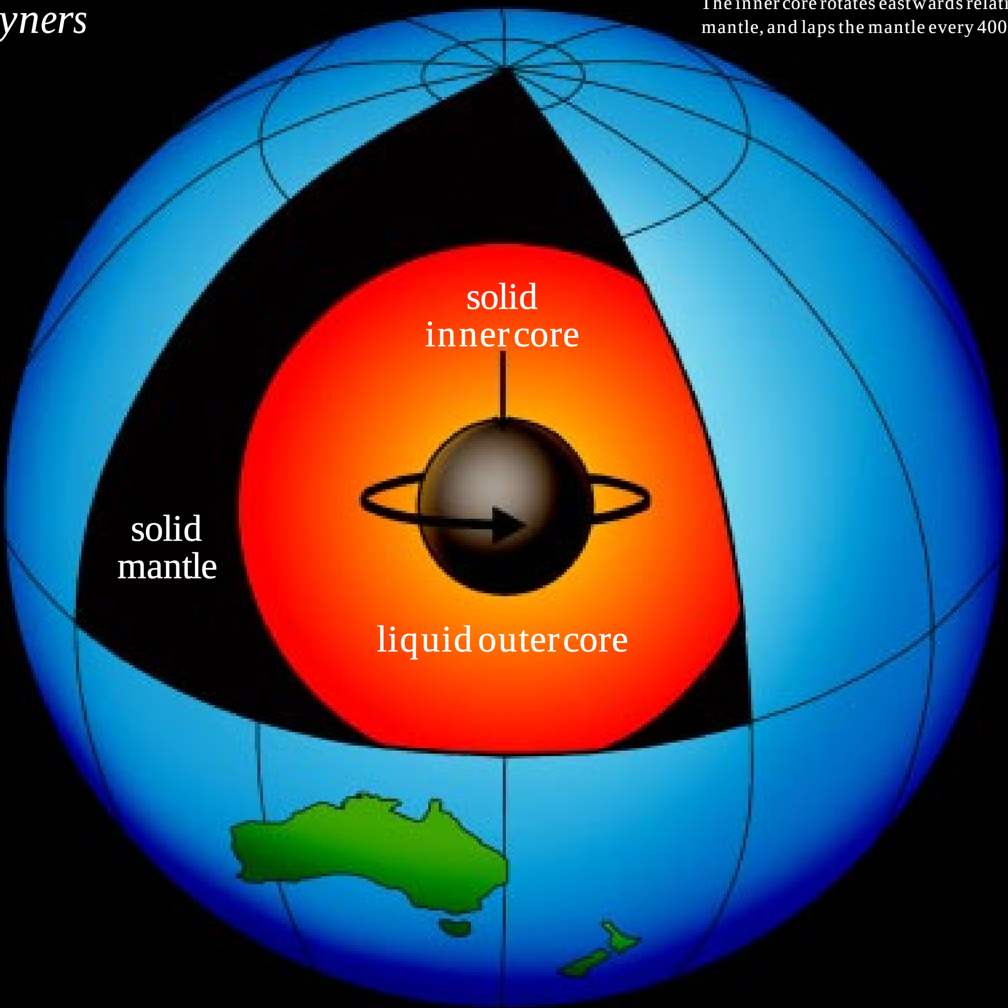
In the last 150 years since European settlement, the west coast of the South Island has experienced several large shallow earthquakes. None of these, however, has occurred on the Alpine Fault. This prominent feature along the western side of the Southern Alps forms the boundary between the Australian Plate and the Pacific Plate where one plate slides past the other.

The Inner Core

THE NEW ZEALAND CONNECTION CONTINUES

by
Martin Reyners

The inner core rotates eastwards relative to the mantle, and laps the mantle every 400 years.



The magnitude 7.8 Murchison earthquake of 1929 was notable not only for the major damage which occurred in the Buller region, but also for the fact that it was well recorded by seismographs worldwide. Indeed, recordings of this earthquake in Europe provided the Danish seismologist Inge Lehmann with evidence for the existence of a solid inner core, with a seismic velocity larger than the liquid outer core of the earth.

Now, nearly seventy years later, the inner core is in the news again. US seismologists have recently found that the inner core spins faster than the earth itself. Their results suggest that the inner core laps the Earth's mantle and crust by a complete revolution every 400 years. Differential rotation of the inner core has major geophysical implications, particularly in generating and maintaining the geomagnetic field.

In studying this differential rotation, it is best to use seismic waves which travel in a north-south direction. This is because the inner core has a "grain" like a piece of wood - seismic waves travel through it faster when following a north-south path and slower when traveling east-west. The rotation of the inner core has been picked up by tracking the orientation of this grain with time using 30 years of seismic records.

This is where the New Zealand connection continues. A very useful dataset to use in this study is large nuclear explosions detonated in the Russian Arctic, and recorded on seismograph stations in Antarctica. New Zealand has operated a seismograph at Scott Base since 1957, and records from this station have proved invaluable in tracking down the differential rotation of the inner core.

mitigate

mitigate v.t. Alleviate (pain, grief); reduce severity of (punishment); moderate (heat, cold, severity etc.); appease (anger etc.).

Lead-Rubber Bearings:

A New Zealand Invention

by
Jim Cousins

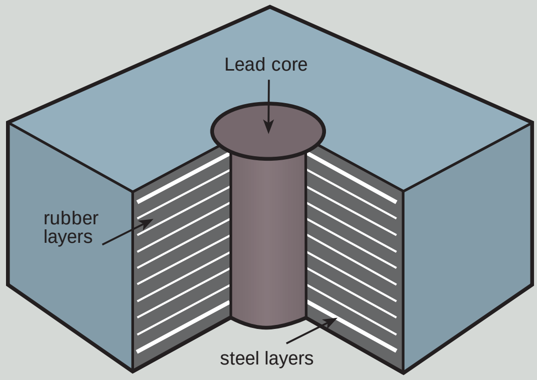
This New Zealand invention is now being used worldwide, in more than 200 buildings and 1000 bridges. Prime local examples are the new Museum of New Zealand, the old Parliament buildings, and the building housing the printing press of Wellington Newspapers at Petone.

A crazy idea - put blocks of rubber between a building and the ground - to isolate the building from strong earthquake shaking? Not really. It's been done, and it works. The trick is to use the right kind of rubber, actually a combination of rubber and steel sheets, and to put a plug of lead down the middle to absorb energy.

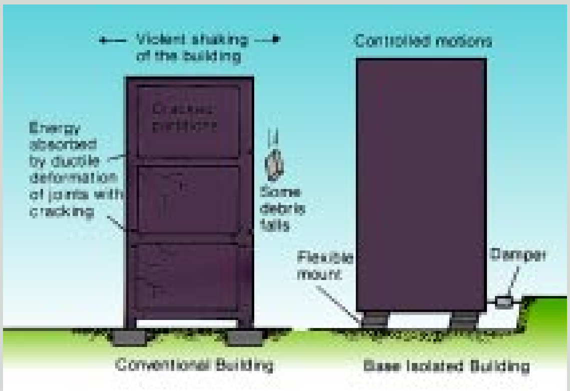
The heart of the matter is the alternating layering of rubber and steel. Each rubber sheet is about 15 mm thick and each steel sheet about 3 mm. Typically there are 20 of each to each bearing (block) giving an overall bearing height of a little less than 500 mm. Bearings for a large building measure 1 m by 1 m.

This is not your ordinary rubber band! Such a bearing needs a fork-truck to lift it, and can carry a load in excess of 2000 tonnes.

The final item needed to turn an ordinary bearing into a seismic-isolator, is a vertical plug of lead to absorb seismic energy during a large earthquake. This was the critical breakthrough made by Dr Bill Robinson of the DSIR in 1976.



The bearings are placed in a layer between the base of the building and the ground. If a strong-earthquake shakes the ground, the bearings are able to flex in shear and greatly reduce the seismic forces on the building.



This crazy idea works. It has been thoroughly tested in the laboratory using enormous, specially constructed test machines and there have been some real-life tests in earthquakes. There were 10 Los Angeles hospitals in the epicentral region of the 1994 Northridge earthquake. Only one remained functional after the earthquake, the one on lead-rubber bearings. Buildings and bridges on the bearings were subjected to strong-shaking during the 1995 Kobe earthquake in Japan. All survived nicely. And finally, the Te Teko bridge in the Bay of Plenty was right at the epicentre of our own Edgecumbe Earthquake of 1987. It too survived nicely.

What are the largest subduction zone earthquakes we can expect?

by
Martin Reyners, John Beavan

Most of the world's great earthquakes and tsunamis involve thrusting between the plates in the shallow part of subduction zones (figure 1). The largest earthquake ever recorded instrumentally, the magnitude 9.5 Chile shock of 1960, was such an event. During this earthquake, the Pacific Plate thrust under the South American Plate by an average of 20 metres along a 1000 km-long section of the coastline. The magnitude 9.3 Alaskan earthquake of 1964 also involved such thrusting between the plates, as did the magnitude 7.9 Kanto earthquake of 1923, which caused major destruction and loss of life in Tokyo. Subduction zones occur in New Zealand from Kaikoura to East Cape (the Hikurangi subduction zone), and also in Fiordland and the region to the south. So what is the potential for such huge events here?

HISTORICALEVIDENCE

In subduction zones with a long historical record, such as in Japan, it has been found that distinct segments of the plate interface rupture repeatedly in large thrust earthquakes. For example, a 300 km-long segment of the plate interface off Shikoku ruptured in large thrust earthquakes in 1707, 1854 and 1946 – sometimes on its own, and sometimes at the same time as a neighbouring segment to the northeast. Because the magnitude of such earthquakes scales with the area that ruptures, the identification of these segments provides seismologists with an estimate of how large future events will be.

Because of our relatively short historical record, we cannot use the same approach in New Zealand. Indeed, historical evidence for large subduction thrust earthquakes is sparse. In 1826 sealers reported a large shallow earthquake in Fiordland that caused uplift and subsidence in the region between Milford and Dusky Sounds, and was followed by a sequence of large aftershocks. This was possibly a thrust earthquake at the shallow part of the plate interface which underlies this region. The only other candidate for a large subduction thrust event in the historical record is the magnitude 6.7 Cape Turnagain earthquake of 1904. This caused sand fountaining, landslides and other ground damage above the shallow part of the plate interface from Napier to south of Masterton, and a large wave (possibly a tsunami) was observed at Mohaka.

THE EFFECT OF PLATE STRUCTURE

With such a dearth of historical information, we need to combine seismological, geodetic and geological techniques to decipher the potential for large subduction thrust earthquakes in New Zealand. What we need to determine is the extent of the 'seismogenic zone' – that part of the plate interface which periodically locks up, causing strain to build up until it is finally released in a large earthquake (figure 1). Research worldwide on the seismogenic zone shows that the ability of the interface to lock up depends on the structure, not only of the interface, but also of the overlying and subducted plates. If the plate interface is smooth and well-lubricated, so that it slips easily, there will be no locking and thus no large earthquakes. Similarly, if the overlying plate is weak, it will deform in response to continued plate convergence if the plates are locked, and again there will be no large thrust earthquakes at the plate interface.

In recent years we have undertaken major studies of the structure of the shallow parts of our two subduction zones using dense deployments of portable seismographs. We can use the numerous small earthquakes recorded during these deployments to

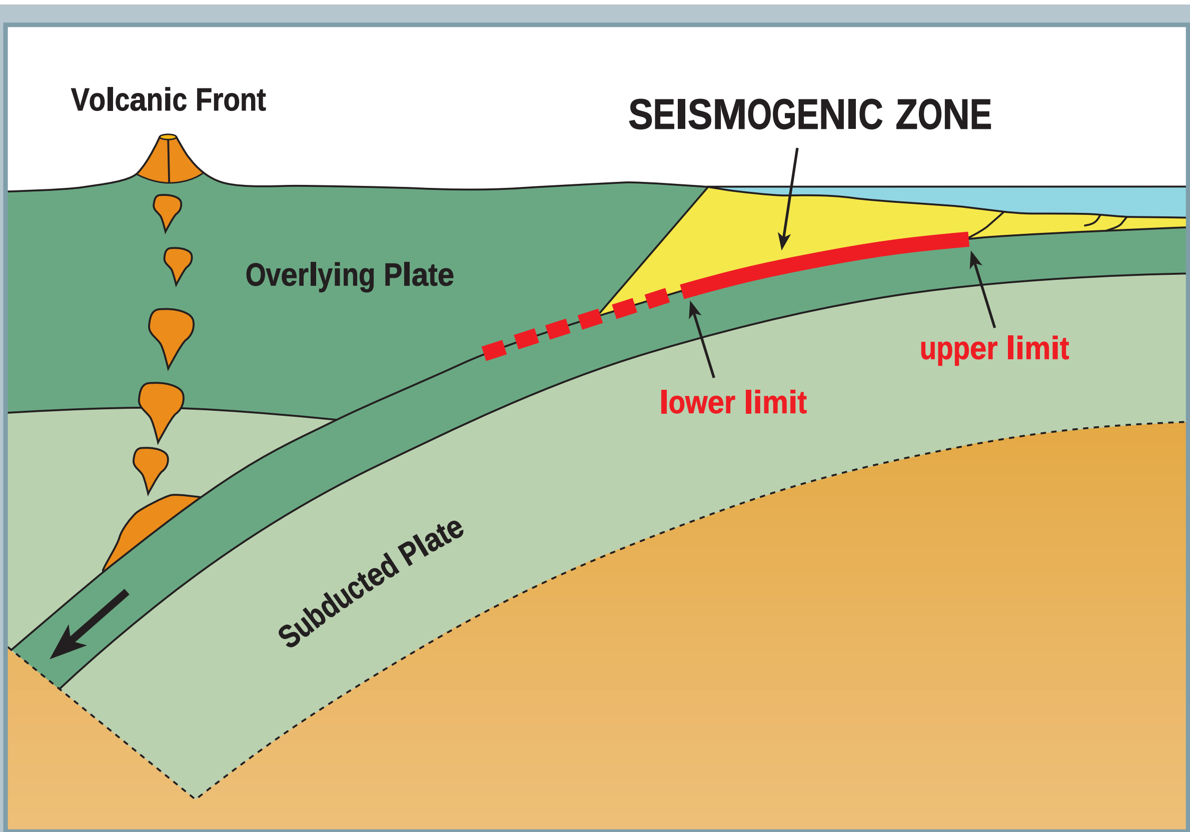


FIGURE 1 Depth section of a subduction zone showing the location of the seismogenic zone.

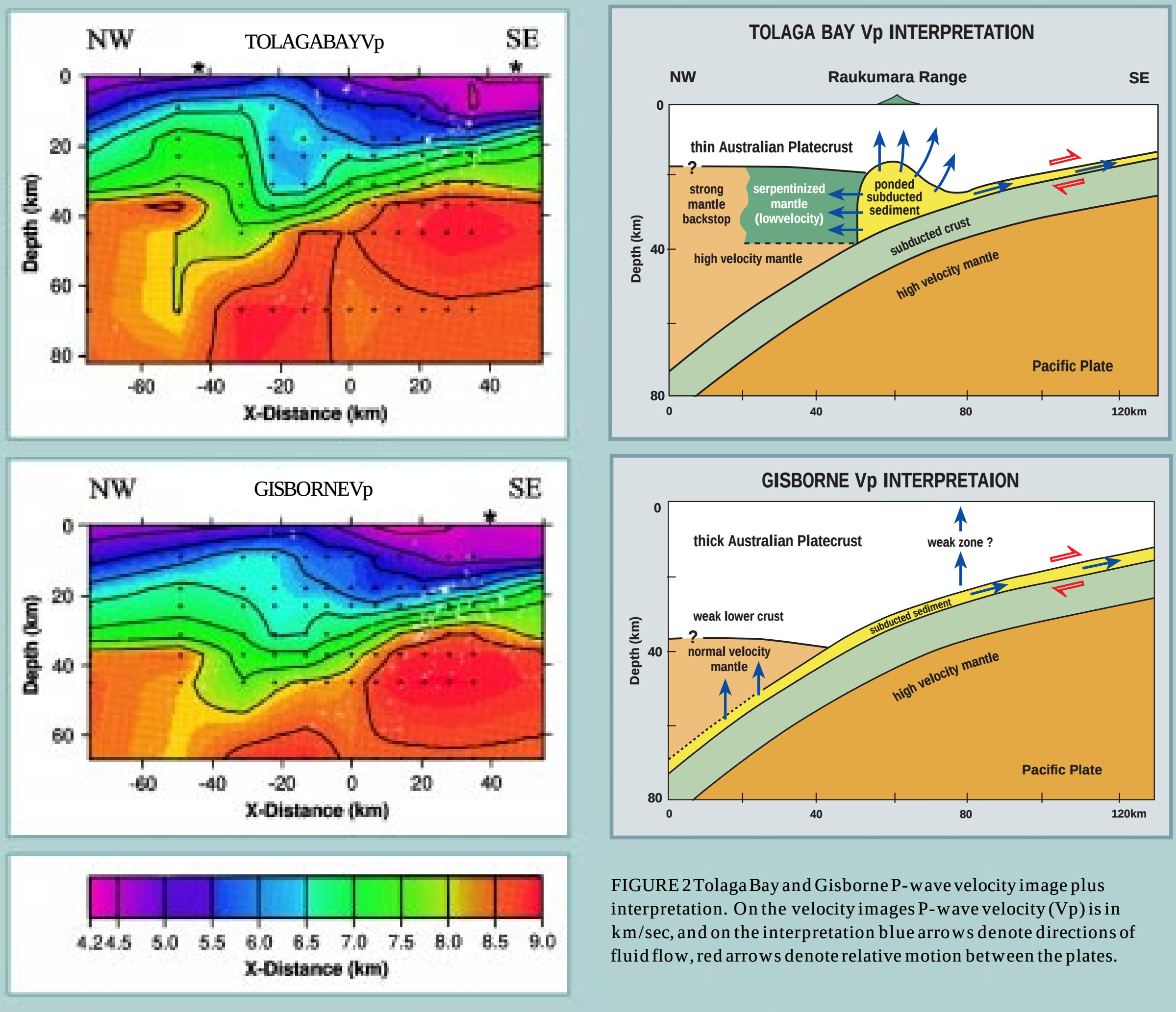


FIGURE 2 Tolaga Bay and Gisborne P-wave velocity image plus interpretation. On the velocity images P-wave velocity (Vp) is in km/sec, and on the interpretation blue arrows denote directions of fluid flow, red arrows denote relative motion between the plates.

obtain an image of the seismic velocity structure of the crust. The technique is termed seismic tomography, and is similar to the techniques used in radiology. While a medical CAT scan uses X-rays to build up a 3-D image of parts of the body, we use seismic waves to build up a 3-D image of the earth's crust.

The results of applying this technique to small earthquakes recorded in the Raukumara Peninsula are shown in figure 2. Two depth sections of the shallow part of the subduction zone are shown, one through Tolaga Bay and the other through Gisborne. On both sections, the subducting Pacific Plate clearly dips towards the northwest. However, the structure of the overlying plate is quite different in the two sections. This difference can be related to a change in thickness of the crust of the Australian Plate, as illustrated in the interpretations of the seismic velocity models (figure 2). In the Tolaga Bay section, this crust is thin, and low-velocity subducted sediment ponds up against the relatively strong upper mantle rocks of the Australian Plate. This ponding of sediment leads to rapid uplift of the Raukumara Range. In contrast, in the Gisborne section the crust is much thicker, and sediment on top of the Pacific Plate subducts to greater depth.

These structural results have implications for the potential for large earthquakes at the plate interface beneath the peninsula. Work at other subduction zones worldwide has shown that the lower depth limit of the seismogenic zone depends on the thickness of the overlying plate. When this crust is thick, as in the southwest of the Raukumara Peninsula, the lower depth limit is temperature controlled, with the transition from seismic to aseismic movement at the interface occurring at about 350°C. When the crust is thin as in the northeast, however, the lower depth limit corresponds to the base of this crust. Thus the seismogenic zone is probably shorter in the northeast than in the southwest. As the magnitudes of large thrust earthquakes generally scale with the width of this zone, we might expect that such events would be smaller in the northeast.

A similar tomographic analysis has been completed for the shallow part of the subduction zone in the Marlborough and Wellington regions. Here we find that changes along the subduction zone are caused by changes in thickness of the subducted Pacific Plate, which thickens appreciably towards the southwest across Cook Strait. As the plate becomes thicker it becomes more buoyant, and thus harder to subduct. This leads to strong coupling between the subducted and overlying plates, with little subduction of sediment between them.

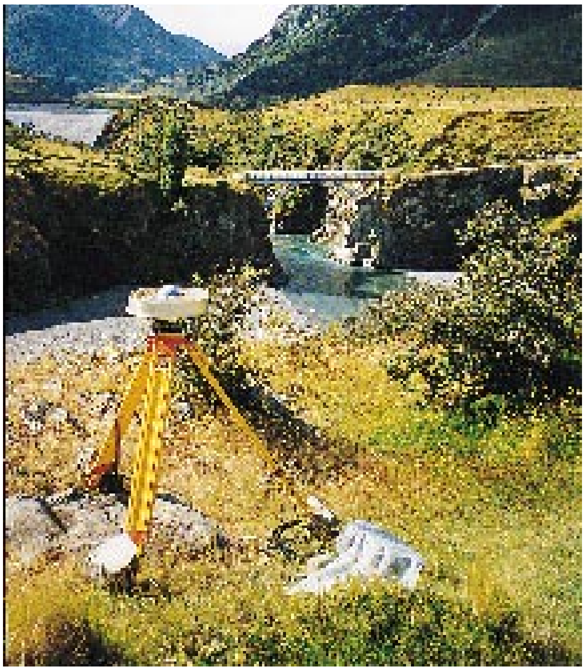
MONITORINGSEISMICSTRAIN

All shallow earthquakes involve shear movement on a fault (albeit a small one for small quakes) in response to stresses within the earth. Thus by monitoring the sense of faulting from many small events recorded in our dense seismograph deployments, we can build up a picture of how the crust is straining seismically in response to stresses generated by any locking of the plates at the seismogenic zone.

When we do this for this for the Raukumara Peninsula, we find that the uppermost part of the Australian plate is characterized by extension towards the trench off the east coast. This can be related to extension and gravity sliding of surficial rocks due to uplift of the Raukumara Range, and is consistent with the plates being weakly coupled in this region. In contrast, small earthquakes in the Wellington and Marlborough regions indicate that the overlying plate is being compressed, suggesting that the plates are strongly coupled there.

USINGGPSTOMONITORDEFORMATIONDIRECTLY

Earthquakes can only give us information on the strain which has been released seismically. As well as measuring this strain release, it is important to measure – if we can – the amount of strain that is building up and that might be released in a future major earthquake. Over the last few years it has become possible to do this with unprecedented speed and accuracy by using signals from Global Positioning System (GPS) satellites. GPS surveying is in some ways similar to conventional surveying using theodolites and electronic distance-measuring devices, in that GPS instruments are set up precisely over survey monuments fixed in the ground. But one great advantage of GPS over conventional methods is that, over any distance greater than about 100 m, it is more accurate. Over typical distances of 10-50 km measured in high-accuracy conventional surveying, GPS is 100 times more accurate – so that motions that would take 100 years to detect by conventional surveying can be measured in just one year by two GPS surveys spaced a year apart.



The GPS station at Poulte River in North Canterbury.

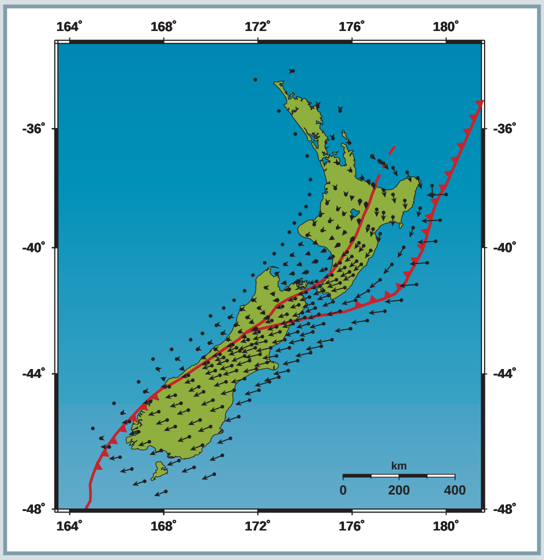


FIGURE 3 First order GPS velocity map for New Zealand.

Repeated high-accuracy GPS surveys have now been done at several hundred survey monuments throughout New Zealand, mainly by the Institute of Geological and Nuclear Sciences but also by Land Information New Zealand, Victoria and Otago universities, and overseas collaborators from the USA and England. Institute scientists have used the year-to-year displacements of some 260 of these points to generate a 'velocity map' of New Zealand. This map (figure 3) shows the velocity at which any point in New Zealand is moving, relative to points within the interior of the Australian Plate; these velocities can be as high as 40 mm/yr or 4 metres per century.

One main feature of the velocity map is the motion away from the Taupo Volcanic Zone, which is an indication that the Taupo region is widening. Another important feature is seen if we examine the velocity arrows as they cross the Hikurangi trench from the Pacific Plate into the Australian Plate. In the northern part of the North Island there is a big change in velocity across the trench, with the velocities on the Australian side of the boundary being in quite different directions to those on the Pacific side. This indicates that the motion of the Pacific Plate is not being transmitted into the overlying Australian Plate, so that the Pacific Plate must be subducting fairly smoothly beneath the

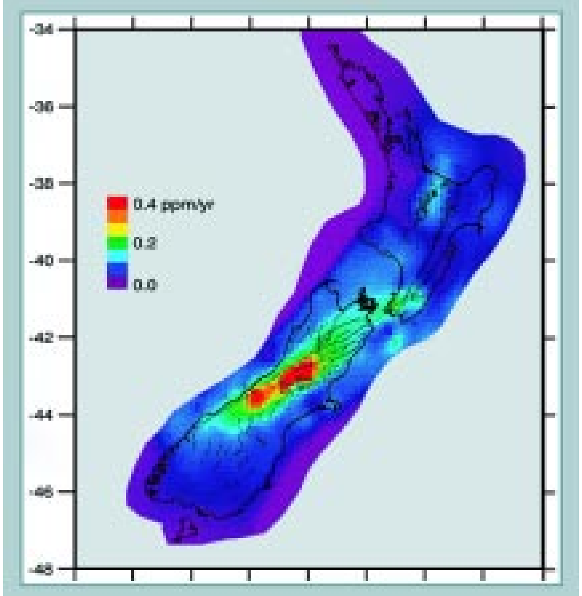


FIGURE 4 New Zealand shear strain map.

northern North Island without causing too much strain or deformation in the overlying plate.

In the southern North Island the situation is different. There is little change in velocity as the trench is crossed, which means that much of the Pacific Plate motion is being transmitted into the overriding Australian Plate, causing it to squash and deform throughout the region. This deformation must eventually be released in major earthquakes, some of which will rupture the plate interface beneath the southern North Island.

So the geodetic measurement of the deformation of the Earth's surface tells us that the Pacific and Australian plates are tightly coupled together under the southern North Island, but only loosely coupled further to the north. This is exactly the same result that was worked out by studying the small and medium earthquakes recorded by the dense arrays of seismometers. The fact that these two completely different ways of looking at the plate boundary provide the same result, gives us confidence that our interpretation is correct.

Where the velocity changes only a small amount between nearby points, not much deformation is occurring. Conversely, where the velocity changes a lot between nearby points, there is a lot of deformation going on. We can convert the velocity map into a related map (figure 4) that shows the rate of straining throughout the country. The most prominent feature of this map is the high shear strain rate that runs under the Southern Alps to the east of the Alpine Fault, and which also extends at a lower level through Marlborough and into the southern North Island. The elevated strain in the southern North Island and the small amount of strain across the central North Island are another manifestation of the change in plate coupling.

The strain rates under the Southern Alps are the highest in the country, even though they are not a result of subduction processes. At first sight it appears that the high rate implies a build-up of strain that will eventually be released in one or more very big earthquakes. But there is another possible explanation. Because of the collision processes between the Pacific and Australian plates that have caused the rise of the Southern Alps, we know that the temperatures of the rock beneath the Alps are raised significantly above their normal levels. It is possible that in these high temperatures the rocks beneath the Alps are able to slowly flow, so that the strain we observe at the surface is largely a result of steady deformation without the occurrence of really big earthquakes. These two different explanations of the data have very different implications for earthquake hazard in the central South Island, and are a subject of vigorous debate between scientists. We hope to discriminate between the two possibilities through experiments, measurements and computer modelling over the next few years – and the results will be the subject of a future story ...

FUTURE WORK

Our work so far has given us insights into how the coupling between the plates changes along the shallow part of the Hikurangi subduction zone, from strongly coupled in the Marlborough region to relatively weakly coupled at East Cape. We have also started to quantify the width of the seismogenic zone which might rupture in a large thrust earthquake, using both seismic and geodetic data. The challenge now is to determine which segments along the plate boundary will rupture in a single event.

We will meet this challenge in the next five years with further focused seismological, geodetic and geological studies. Existing data from seismograph deployments in Fiordland, northern Wairarapa and southern Hawke's Bay will be analysed, and a further deployment is planned for northern Hawke's Bay. Also, a transect across the subduction zone from Hawke Bay to off the west coast at Kawhia, using explosives on land and air-gun shots at sea, will provide a finely detailed image of the structure of the plate boundary in this region. We will also increase our network of GPS stations, so that the shallow parts of our subduction zones will have been surveyed at least three times. An exciting new development will be the installation of some continuously recording GPS stations, which will allow us to monitor strain accumulation in real time. And we will search for geological clues to previous subduction events, principally by tracking regional subsidence through the study of cores from lagoons, estuaries and lakes.

NOTES