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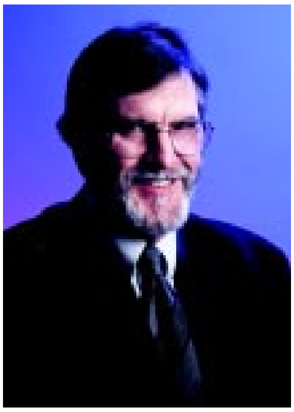


Welcome to the 1999 issue of Tephra. Over the years Tephra has developed a deservedly high reputation in New Zealand and around the world for its high quality of content and presentation. I am delighted to have the opportunity to introduce this issue, the last issue of the century, and coincidentally my first as Director.

Tsunami. The very name conjures up an image of a frightening force rarely seen and even harder to predict. It is unlike an earthquake fault line where we are able to measure movement and speculate on when it may next move, or a flood plane, where we can consider river patterns, land usage and weather patterns to assess future problems.

Tsunami may come from the other side of the ocean, allowing only a few hours warning that something could happen, the scale and location of which can only be roughly predicted. They may also be generated from local earthquakes or from massive submarine landslides. In these instances warning may be only a few minutes - as in the Papua New Guinea Tsunami in 1998.

There is evidence of large Tsunami (12-14m or more in height - the scale of which is difficult to comprehend!) in parts of New Zealand and some record of events over the past 150 years.

As with other hazards, it is possible to identify susceptible locations and to assess the scope and scale of potential impacts.

Vulnerable communities can be informed, land use planning choices made and simple explicit plans and instructions put in place to respond to the event should it occur.

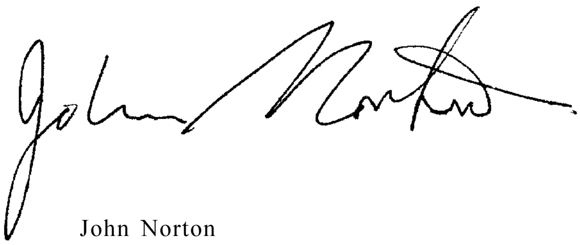
This of course is in line with the new risk management approach to reducing vulnerability and managing emergencies. It does however require a commitment of resource to understand the hazard and its potential consequences on particular communities. It is only through action that we can reduce vulnerability and develop resilient communities.

Tsunami are fascinating phenomena as I am sure you will find in this issue of Tephra. The challenge is to understand not just what they are, but what they can do to us and what we can do to minimise their impacts.

The increasing impact and cost of disasters around the world reinforces the importance of continuing the scientific understanding of hazards. The added dimension to give value to this work is the understanding of the potential consequences on communities and infrastructure.

It is this dimension we wish to develop, including a social and economic component, to better inform the planning and choices available in risk management.

I would like to thank all the contributors for their effort in making this issue possible.



John Norton
Director
Ministry for Emergency Management

OVERVIEW OF

Tsunami hazard in *New Zealand*

by Willem de Lange and Rodger Fraser

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Tsunami are relatively common coastal hazards in New Zealand. On average there are 12-13 tsunami exceeding 1 m every century around the New Zealand coast, a similar frequency to Indonesia and Hawaii, but about a third that of Japan (Figure 1). However, there has been only 1 death officially attributed to tsunami, although several hundred Maori may have been killed by tsunami last century before their deaths were recorded. The lack of fatalities and limited damage caused by historic tsunami leads to an impression that, despite the similar frequency of events, New Zealand tsunami hazard is considerably less than that of Hawaii and Indonesia.

New Zealand has experienced at least 38 tsunami since 1820 (Table 1). Several prehistoric tsunami events have also been identified from deposits at locations on the coast of Cook Strait. Considering the available data for tsunami since 1840, the annual exceedence probability f for the entire New Zealand coastline is given by

$$f = \exp(-0.213H - 1.83)$$

This equation defines the probability that a tsunami wave somewhere around the New Zealand coast will exceed the specified height H. Many people are more comfortable with the concept of return period, or the average number of years between events of a given size. The annual exceedence probability can be expressed as a return period by taking the reciprocal.

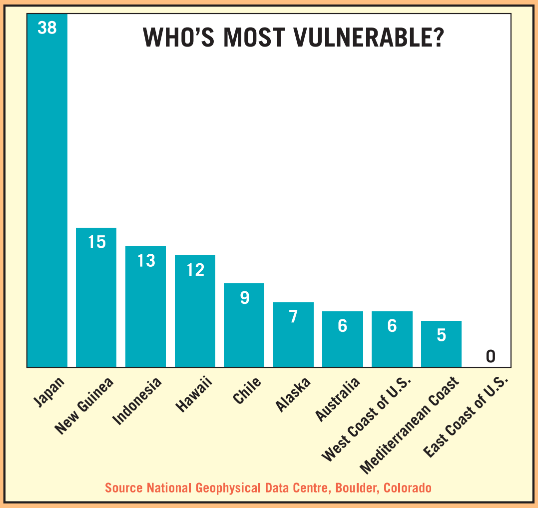

Figure 1 - Number of tsunami exceeding 1 m in height per century for various countries (Time magazine, 28 September 1998).

This has been done for a range of tsunami wave heights in Table 2. The expected return period of tsunami exceeding 1 m for New Zealand is similar to that for Hawaii and Indonesia and half that of Australia. Further the return period for catastrophic tsunami exceeding 10 m in height is a little over 50 years. Since 1840 catastrophic tsunami have only affected sparsely populated coastal regions, and so are not perceived as a major threat. Due to increased development of coastal regions since the last major tsunami event in 1964, this may be a dangerous

Date (UTC)	Source location	Maximum height (m)	Comments
1800 BP	Lake Taupo	unknown	Rissaga caused by eruption
AD 350	Cook Strait	unknown	Local earthquake
AD 1220	Cook Strait	unknown	Local earthquake
AD 1440	Cook Strait	unknown	Local earthquake
~AD 1820	Probably Fouveaux Strait	>10 m ?	Many killed at Orepuki, landslide?
~AD 1830	Kaikoura	unknown	Local landslide
15 Oct. 1848	Lower Wairau Valley	0.3	
23 Jan. 1855	West Wairarapa	9.1	
March 1856	Chatham Rise	Uncertain event	
13 Aug. 1868	Chile	3.1	Largest nationwide tsunami
18 Oct. 1868	Cape Farewell	?	Possibly coseismic seiching
10 May 1877	Chile	3.7	
27 Aug. 1883	Krakatau, Indonesia	1.8	Rissaga caused by eruption
22 June 1891	Waikato Heads		Uncertain event
7 Aug. 1904	Cape Turnagain	"Large wave"	Possibly due to landslide?
22 Feb. 1913	Westport	1.5	Landslide
1 May 1917	Kermadec Islands	?	Recorded in California
11 Nov. 1922	Chile	0.2	
25 Dec. 1922	Rangiora	0.4	Probably an immediate wave
1 Sep. 1923	Kwanto, Japan	<0.1	
1927-1928	Tolaga Bay	>4	3 large waves, landslide?
16 June 1929	Whitecliffs, Karamea	2.5	Landslide
2 Feb. 1931	Waikare, Hawke Bay	15.3	Landslide
2 Feb. 1931	Napier	~3	
25 Mar. 1947	Gisborne	10.0	2 large waves, landslide?
17 May 1947	Gisborne	6.0	Landslide?
14 Mar 1950	Bay of Plenty	Uncertain event	
4 Nov. 1952	Kamchatka	0.9	
22 May 1960	Chile	5.5	Largest teletsunami this century
28 Mar. 1964	Alaska	0.9	
4 Jan. 1976	Kermadec Islands	0.8	
22 June 1977	Tonga	<0.2	
25 May 1981	Macquarie Ridge	0.3	
Dec. 1982	Kermadec Islands	<0.1	
Oct.1986	Kermadec Islands	<0.1	
20 May 1987	Doubtful Sound	3.0	Landslide
June 1993	Kermadec Islands	<0.1	
6 Oct. 1994	Kuril Islands	<0.1	
16 Jan. 1995	Japan	0.1	Kobe Earthquake
25 Mar.1998	Balleny Islands	0.2	
17 Jul. 1998	Sissano, Papua New Guinea	0.3	

Table 1. Summary of prehistoric and historic tsunami events recorded around the New Zealand coast.

Height (m)	Annual exceedence probability	Return period (y)
1.0	0.1296	7.7
2.5	0.0942	10.6
5.0	0.0553	18.1
10.0	0.0191	52.5

Table 2. Predicted tsunami annual exceedence probabilities and return periods for tsunami along the New Zealand coast. The values were determined using available data on historic New Zealand Tsunami since 1840. For most populated areas of the New Zealand coast tsunami exceeding 1 m in height are considered to be a significant hazard requiring mitigation or a Civil Defence response, while tsunami exceeding 10 m would be catastrophic.

perception.

The tsunami recorded in New Zealand are subdivided into two main groups for the purposes of hazard assessment:

- Teletsunami that have been generated beyond the New Zealand continental shelf. These tsunami have longer periods and persist for several days. They also affect most of the New Zealand coast; and
- Local tsunami that have been generated on the New Zealand continental shelf. These tsunami have shorter periods and do not last long. They also only affect a limited section of the New Zealand coast.

Teletsunami

Most historic teletsunami have been quite small (<0.5 m) with minimal impact on the New Zealand coast. However a few have been large enough to cause extensive damage. Combining the results from computer models with historical data, it is possible to assess which seismic regions around the Pacific Rim are likely to produce major teletsunami that can cause damage on the New Zealand coast. The National Geophysical Data Center in the USA maintains a global database of tsunami events. To assist with classifying events, the Pacific Ocean Basin and surrounding seas have been subdivided into a number of tsunami generating regions (Figure 2). For each of these regions, the characteristics of the minimum earthquake capable of generating a teletsunami affecting New Zealand can be defined. As more data become available, the likely coastal response for earthquakes exceeding this minimum threshold can be better characterised.

Some regions can be ruled out as potential generators of hazardous teletsunami affecting New Zealand. The areas of negligible hazard are Hawaii, New Guinea and Solomon Islands, Indonesia and Philippines. These areas are excluded either because the potential tsunami generating mechanisms are incapable of producing a hazardous teletsunami (viz. Hawaii) or, as for the other areas, the travel paths of any teletsunami reaching New Zealand are indirect with very high energy dissipation. Normally this occurs because the orientations of the subduction zones in these regions direct tsunami energy away from New Zealand. None of these areas have been associated with a teletsunami recorded in New Zealand.

Next there are regions that represent a minimal hazard. These are the New Zealand Exclusive Economic Zone beyond the continental shelf and the islands of the South Pacific, Japan and the Kuril Islands and Kamchatka. All these regions have generated teletsunami recorded in New Zealand. The south-west Pacific tsunami are generated by tectonic structures that direct the tsunami waves away from New Zealand, so that the resulting waves are also small in New Zealand. Shallow earthquakes larger than $M_w = 7.5$ near the Kermadec Islands have generated tsunami up to 0.75 m height in northern New Zealand. However, most of the tsunami energy from the Kermadec Islands seems to be directed towards the north-east Pacific Ocean. Hence California tends to experience larger tsunami from the Kermadec Islands than New Zealand. For sources further north, much larger earthquakes are needed to create a detectable wave.

Until recently the area south of New Zealand was not considered a significant tsunami source; indeed no region code had been established for the global tsunami database to cover Antarctica. However on the 25th March 1998 an earthquake near the Balleny Islands generated a small teletsunami that reached Australia, New Zealand and several South Pacific islands. Two other historic tsunami have been generated south of New Zealand. The 1981 Macquarie Ridge teletsunami was also small, and the remaining event was identified from early settlers' records of Maori oral traditions and may have been a local event. Therefore the historic record suggests that teletsunami from the south will probably be small. Certainly most known tectonic structures south of New Zealand would direct tsunami away from New Zealand, and the shallow continental platforms such as the Campbell Plateau disperse tsunami energy, further reducing their impact. Therefore, teletsunami from south of New Zealand are not considered a major hazard.

From historical data, very shallow earthquakes with moment magnitudes > 8.25 located on the east coast of Japan, Kamchatka, and the Kuril and Aleutian Islands are needed to produce a significant teletsunami (0.25 m) in New Zealand. Even though these subduction zones are a little more favourable for directing tsunami energy towards New Zealand, most energy is directed towards South America. Further, the travel paths of the north-west Pacific tsunami involve considerable energy dissipation due to the many shallow island groups encountered. Neither of the most destructive tsunami generated in this region, the 1896 Sanriku Tsunami and 1946 Aleutian Tsunami, appear to have reached New Zealand. Both of these events were produced by unusual tsunami earthquakes and may have been generated by landslides, which would cause them to dissipate rapidly.

From the available historical data, the west coast of South America represents the greatest teletsunami hazard. Four teletsunami have been recorded in New Zealand from this source, including three of the most destructive events (in 1868, 1877 and 1960). All the historical teletsunami have been associated with shallow earthquakes with Richter

Figure 2 - Map of the Pacific Ocean showing the tsunami generation zones defined by the National Geophysical Data Center, USA.

The source locations are: 80 - Hawaii; 81 - New Zealand and South Pacific islands; 82 - New Guinea and Solomon Islands; 83 - Indonesia; 84 - Philippines; 85 - Japan; 86 - Kuril Islands and Kamchatka; 87 - Alaskan (including the Aleutian Islands); 88 - West coast of North and Central America; and 89 - West coast of South America.

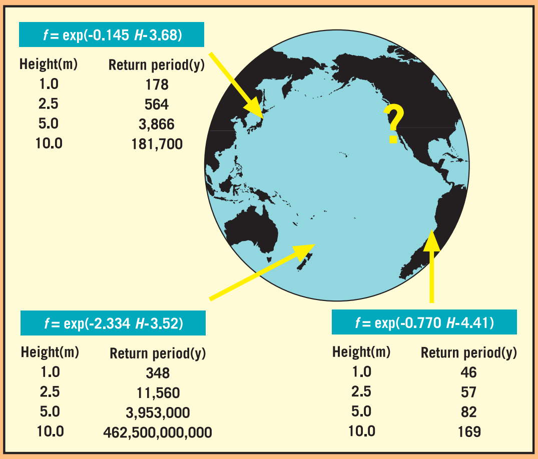

Figure 3 - Return periods for a range of tsunami wave heights from source regions that have produced more than 3 teletsunami affecting New Zealand since 1840. The equations define the annual exceedence probabilities f for the specified wave height.

Table 3. Tsunami return periods (years) for the specified heights determined for a selection of New Zealand major and minor ports. The return periods were calculated using the annual exceedence probability distributions of Fraser (1998).

	Tsunami height (m)			
Location	1.0	2.5	5.0	10.0
Whangarei	179	930	14,500	3,510,000
Auckland	85	427	6,280	1,360,000
Tauranga	80	322	3,300	345,000
Gisborne	44	67	135	556
Napier	56	97	243	1,540
Wanganui	79	147	414	3,260
Wellington	40	119	728	27,200
Lyttelton	35	52	101	376
Timaru	63	130	439	5,010
Dunedin	125	1075	39,000	51,000,000

magnitudes greater than 8.2, and have produced wave heights >5 m. Assessments of tsunami hazard undertaken by Regional Councils have all identified South American tsunami as the most credible threat, with the 1868 event representing the worst scenario for many east coast locations. Derek Todd discusses the assessment of these events for the Canterbury and Otago regions elsewhere in this issue.

Figure 3 summarises the probability distributions and return periods for the three main tsunami source regions identified above. The data presented in this diagram highlights the historic impact on New Zealand of tsunami from the west coast of South America. However, there is one obvious region of uncertainty: the west coast of Central and North America. Only one historic teletsunami has been recorded from this region, the 1964 Alaskan Tsunami, which provides insufficient data to determine the exceedence distribution. The tectonic structures in Central America would tend to direct tsunami energy along the Equator and into the North Pacific Ocean, so that New Zealand would not be greatly affected. The 1992 Nicaraguan and 1996 Peru Tsunami do not appear to have been detected in New Zealand. However these events involved unusual tsunami earthquakes, and may not be representative of tsunami from this region.

However, further north in the Cascadia region, the tectonic structures may direct tsunami towards the South-west Pacific, and New Zealand. Research in British Columbia, Washington and Oregon suggests that tsunami are generated in this region by very large earthquakes. Dating of tsunami deposits indicates that large tsunami occur with return periods of 300-400 years, and the last event occurred in December 1700 based on historical data from Japan. Tsunami from this area would also tend to concentrate energy on New Zealand due to the curvature of the Earth. Therefore it is likely that this area represents a moderate to high teletsunami hazard.

The historical record indicates that although the impact of teletsunami varies around the coastline, some trends are evident that allow an overall assessment of teletsunami hazard. The first conclusion that can be drawn is that the east coast is more susceptible to tsunami than the west coast. This mainly occurs because the main tsunami sources lie to the east of New Zealand, and sources to the west are blocked by islands and shallow seas. Tsunami waves from the eastern Pacific Ocean do reach the west coast of New Zealand, mostly by reflection off the Great Barrier Reef of Australia. Therefore they are smaller, and arrive a lot later than the direct waves reaching the east coast of New Zealand.

Some regions along the east coast of New Zealand show consistently higher than average responses, and others are consistently lower than average. There are two main regions that tend to have a higher than average response to teletsunami: Banks Peninsula; and Poverty Bay. Banks Peninsula consistently has a large response, probably due to resonance in Pegasus Bay amplifying the tsunami waves. It is also possible the Chatham Rise may also concentrate tsunami energy on the Peninsula. Increased tsunami wave heights were observed around Banks Peninsula during the 1868, 1877 and 1960 Chilean tsunami.

Numerical modelling of Poverty Bay shows that resonance may occur in Poverty Bay, amplifying the tsunami waves. Simulations of the 1960 Chilean tsunami also show that the East Pacific Rise tends to focus South American tsunami in this area.

There are also more localised areas where refraction of tsunami waves around offshore islands concentrate wave energy producing larger tsunami (Omaha Bay), funnel shaped embayments concentrate the tsunami energy at the head of the bay (Mercury Bay), or the tsunami causes estuaries to seiche and amplify the waves (Wellington Harbour). Cook Strait, particularly Tasman and Golden Bays, tends to have significantly smaller tsunami waves, probably due to high dissipation of the tsunami energy by shallow water.

Local tsunami

Local tsunami in New Zealand have been caused by a variety of mechanisms including earthquakes, landslides and volcanic eruptions, and include the largest tsunami recorded in New Zealand (Table 1). Local tsunami can be generated anywhere around the New Zealand coast. However three regions are of particular interest: Poverty Bay to East Cape; Cook Strait; and the West Coast.

The continental shelf off Poverty Bay and East Cape appears to produce a large number of tsunami earthquakes, where the tsunami produced is much larger than would be expected from the earthquake magnitude. These may be due to landslides triggered by the earthquake. Seismic and side-scan sonar surveys of the continental margin in this region have shown many landslide features as discussed elsewhere in this issue by Keith Lewis and others. However, no landslides have yet been identified in the vicinity of the likely source of the large 1947 tsunami north of Gisborne.

An alternative source for some of the large local tsunami is mud volcanism along the offshore Ariel Bank. In this area old sea-mounts and guyots on the descending Pacific Plate are causing extreme compression of soft sediment on the overlying Indo-Australia Plate. The sediment is forced into large folded structures called diapirs, and every so often the overlying rock ruptures and mud, gas and fluid is forced out. This may involve the sudden uplift of overlying rocks, or the violent eruption of material at the surface as a mud volcano. Several large events have been recorded this century ~20 km north of Gisborne in the Mangaehu and Waimata Valleys. The first occurred on July 25, 1908, and produced an eruption column ~120 m high, depositing ~150,000 m^3 of material around the vent. Another occurred on May 6, 1930, and deposited ~270,000 m^3. Neither eruption was accompanied by seismic activity. At the coast, ~100,000 m^2 of sea floor in Sponge Bay near Gisborne was uplifted 2 m within a few minutes on 17 February, 1931. Again no seismic activity was observed.

Diapirs are also found beneath the continental shelf, particularly near the large shore parallel shoal known as Ariel Bank. There is no conclusive evidence that there have been any mud eruptions associated with the offshore diapirs. However patches of disturbed muddy water have been observed near the offshore diapirs in 1877 and 1947 that may be the result of the eruption of gas and mud from the sea floor. Numerical modelling also has shown that mud eruptions are the most likely cause of the large tsunami of 25 March and 17 May 1947 that struck the coast between Mahia Peninsula and Tolaga Bay. The same mechanism probably also generated the tsunami waves that damaged the Tolaga Bay Wharf during construction between 1927 and 1928.

These tsunami are particularly hazardous because they are associated with either no seismic activity or with quite small earthquakes, and the waves are quite large (5-10 m). This in combination with the short travel times between the source and the coast means that there is very little warning of their arrival and their impact is severe.

Several large fault zones associated with the Alpine Fault cross Cook Strait. One of these, the West Wairarapa Fault, was associated with the largest historic earthquake in New Zealand and the largest earthquake generated tsunami. The Wairau Fault also may have generated a tsunami 7 years earlier. Two historic tsunami from Cook Strait may not appear to represent a significant hazard. However it is possible to identify prehistoric tsunami from the deposits they leave behind. Cores of wetlands along the Cook Strait coast have identified at least 5 tsunami during the last 2,000 years (Table 1). The important feature of the tsunami deposits is that the large 1855 historic tsunami is poorly preserved in the sediment record. This suggests that for the sites examined, the prehistoric tsunami were probably larger. Three of the prehistoric tsunami have radiocarbon dates that closely match known movements on the local faults, and the remaining event has a date similar to the large Taupo Eruption around 1800 BP.

The sedimentary record therefore indicates that the Cook Strait region may occasionally experience large tsunami (>10 m) generated by local earthquakes. However since there is little data on the prehistoric events, it is difficult to determine whether the tsunami hazard is greater than the earthquake hazard for coastal areas. It is certain, however, that the tsunami will exacerbate the earthquake impacts. Given the apparently low frequency indicated by the sedimentary record for these events it may be difficult to justify the expense of building protective structures.

Although the west coast of New Zealand has been largely unaffected by historic tsunami, the West Coast and Fiordland have experienced several local tsunami generated by landslides. This is likely to continue due to the ongoing uplift of the Southern Alps. Landslide tsunami can be very large, particularly in confined waters such as the fiords. However, due to the sparse population in these areas, the risk associated with local tsunami is low.

Besides the landslide features observed near Poverty Bay, the Hikurangi Trough and adjacent continental shelf margin as far south as Kaikoura contains numerous landslide scars and deposits. These indicate that there is the potential for landslide tsunami. However as yet there is no data on the age and frequency of these slope failures.

There have been no historic volcanic tsunami

Figure 5 - Exceedence probability distributions for major ports around the New Zealand coast. Lyttelton and Gisborne clearly have a higher hazard than other ports.

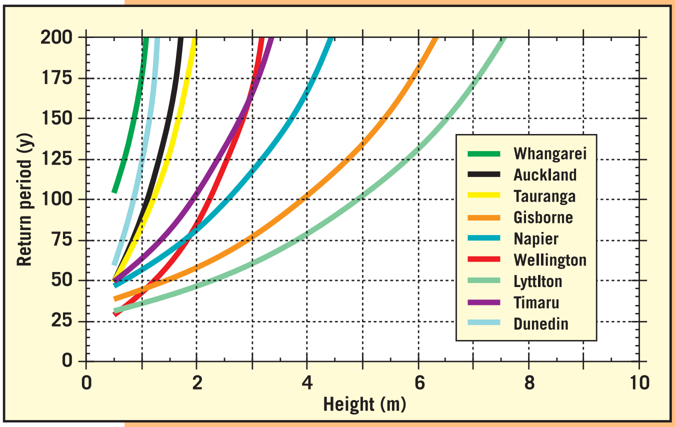

Figure 4 - Map of the New Zealand region, showing locations of points in Figure 5.

observed in New Zealand. However New Zealand is a tectonically country and volcanic activity is relatively high. Three main areas capable of generating a volcanic tsunami have been identified: Auckland Volcanic Field; offshore Taupo Volcanic Zone; and onland Taupo Volcanic Zone.

The Auckland Volcanic Field consists of about 50 monogenetic basaltic volcanoes. This type of volcano is not normally considered to be a significant source of tsunami. However it is possible for tsunami to be generated by phreatomagmatic eruptions during the formation of maars, when the rising basalt magma comes into contact with water under suitable conditions. This style of eruption is prevalent in the Auckland Volcanic Field. Numerical modelling shows that, given a suitable vent location, a sequence of closely spaced explosions may generate 1-2 m tsunami waves along the eastern beaches of Auckland. However, it is more likely that any tsunami generated by eruptions in this Field will be quite small (<0.5 m).

There are several offshore volcanic vents in the Bay of Plenty associated with the Taupo Volcanic Zone, including White and Mayor Islands. Public concern over tsunami produced by an eruption of White Island triggered by sea water entering the crater led to the first numerical modelling of tsunami in New Zealand. Two independent studies concluded that White Island would not generate large tsunami affecting the coast. The main reason is that the volcano is in deep water beyond the continental shelf and most of the energy of any tsunami produced is reflected away from the coast. It was also found that explosions are very inefficient at generating waves, so an extremely large explosion would be required, and there is no evidence that White Island would be capable of producing such and explosion. White Island also is unlikely to produce pyroclastic flows, which are the most efficient generation mechanism. However, Mayor Island is on the continental shelf and has undergone several episodes of major eruptions producing pyroclastic flows. Numerical modelling shows that the largest credible eruptions at Mayor Island could produce 20 m high tsunami along the Bay of Plenty coast.

There are also a series of volcanoes further offshore between the Bay of Plenty and the Kermadec Islands. The Kermadec Islands are an active source of seismic teletsunami, but are not known to have generated a volcanic tsunami. However the eruptive style of the Kermadec Island volcanoes is capable of tsunami generation. Until recently the volcanoes between White Island and the Kermadec Islands were considered to be andesite volcanoes similar to White Island. With the collection of detailed side-scan sonar images and better hydrographic data, several large caldera volcanoes similar to Lake Taupo have now been identified. These may represent a significant threat, but little is known of their eruptive history which makes it difficult to make a sensible assessment of the likely hazard.

The remaining volcanoes of the Taupo Volcanic Zone are on land, which suggests that they are incapable of generating a tsunami. However, some of the volcanic centres are capable of producing very large eruptions (particularly Okataina and Lake Taupo). These may be capable of generating rissaga by atmospheric coupling, or tsunami through large pyroclastic flows. Coastal cliffs near Matata in the Bay of Plenty were formed by pyroclastic flows reaching the sea from the Okataina Volcanic Centre 30-40,000 years ago. One of the prehistoric tsunami deposits identified around Cook Strait may also represent a rissaga, or meteorological tsunami, generated by the 1800 BP Taupo Eruption by coupling between the atmospheric waves produced by the eruption and the ocean. The 1883 Krakatau Eruption generated rissaga up to 2 m in height around the New Zealand coast. A local eruption is expected to produce much larger waves, probably around 5-10 m in height as indicated by the Cook Strait deposits.

Regional tsunami hazard

The tsunami annual exceedence probability distributions have been for several major and minor ports around New Zealand (Figure 5). The resulting distributions were used to determine the return periods summarised in Table 3. These data indicate the relative tsunami hazard around the New Zealand coast, although they should be treated with caution as the data used to derive the distributions are of poor quality. This suggests that the greatest hazard occurs around Banks Peninsula (Lyttelton Harbour) and Gisborne.

The high hazard around Banks Peninsula is due to amplification of teletsunami by several processes, including: resonance of Pegasus Bay; refraction over the Chatham Rise; and resonance within the harbours, particularly Lyttelton Harbour. The 1998 Saundaun teletsunami experienced a 300% amplification between Kaikoura and Lyttelton. Historical data suggest that greater amplification occurred for the 1868, 1877 and 1960 tsunami from South America.

The high hazard around Gisborne is partly due to amplification of teletsunami. This has been attributed to excitation of continental shelf edge waves between Banks Peninsula and Gisborne. However, it is more likely due to local resonance involving Poverty Bay and the adjacent continental shelf. The hazard is also increased at Gisborne by the frequency of large local tsunami, such as the two events of 1947.

Unfortunately, despite the high incidence of tsunami events, the data available to make hazard assessments are very limited. Hence these data need to be supplemented by numerical simulations. Several numerical tsunami studies are discussed in other articles in this issue. One difficulty with numerical simulations is that ideally there should be a calibration with known events and their effects. Due to the paucity of data, most numerical simulations for New Zealand have not been rigorously calibrated. To assist with future work, we would appreciate any additional data on historic tsunami that readers may have.

Tsunami hazard mitigation in New Zealand

The Ministry of Emergency Management and Regional Councils are responsible for the mitigation of natural hazards in New Zealand. For teletsunami events the Ministry of Emergency Management is the contact for the PTWC and they receive all bulletins issued concerning tsunami. Only the Tsunami Watch and Warning Bulletins require further action.

When a Tsunami Watch Bulletin is received the information is relayed to Regional Councils and territorial authorities, the Police and New Zealand Defence Force, and other organisations concerned with hazard mitigation. The public are not alerted unless the tsunami arrival time is less than four hours (which is only the case for tsunami generated within 3000 km of the coast).

A Tsunami Warning Bulletin indicates that a tsunami has been generated. If the arrival time is greater than 4 hours, the Ministry of Emergency Management consults with its' scientific advisers to determine the necessary response. Depending on the level of threat, the Ministry can issue tsunami warning bulletins to affected regions and the necessary authorities. The Ministry will also advise the public of the hazard. If the travel time is less than 4 hours, the Ministry will normally issue tsunami warning bulletins without scientific advice. The Regional Councils and territorial authorities have responsibility for evacuation procedures.

PALEOTSUNAMI: Now you see them, *now you don't*

by Catherine Chagué-Goff[1] and James Goff[2]

1. GeoEnvironmental Consultants
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Paleotsunami are a vital link between our past and our future. They are only known through a distinctive arrangement of geological deposits because they mostly occurred before written records - undoubtedly as far back as millions of years ago. If we can understand the nature of past tsunami, then we can start to understand how they will affect us in the future.

This is a noble aim, but the reality is somewhat different. When dealing with the past we always have to remember that we have an incomplete record. The event, be it catastrophic or everyday, has passed and all that remains is the memory preserved in the landscape. The carnage and devastation left behind after a tsunami or a cyclone might appear to be a significant "memory", but it tells us all too little about the physical properties of the event such as how long it lasted. Furthermore, these memories deteriorate with time and we are eventually left with what is just an erosional and depositional record. Subtle indicators such as debris up trees to mark wave height have long gone and the detective work begins. The further back in time, the less likely one is to have any contemporary accounts - the page is blank.

Relating a geological deposit to a paleotsunami is a delicate exercise. For example, there is considerable worldwide debate as to whether it is possible to differentiate between paleotsunami and paleocyclones in the record of old sediments. While it is easy enough to identify the difference between a large, well documented tsunami and a small, well documented cyclone, there is a significant grey area in the middle that is likely to remain unresolved for many years.

In New Zealand, less however, we have made considerable progress in addressing the issue. This is important because in the past 150 years or so since written records began there have been over 50 tsunami, big and small. Before this there are few definite tsunami that can be interpreted from the Maori oral record, so past ones have to be deciphered geologically.

Researchers have developed a series of diagnostic criteria that can be used to identify paleotsunami deposits. Individually, these are not mutually exclusive to tsunami but, as a suite of features, the more that are identifiable, the more robust the interpretation.

Diagnostic characteristics

• Each wave (a tsunami normally consists of a train of 3-11 waves) can form a distinct deposit, although this is not often recognised in the sediment.

• The deposits generally taper inland like a wedge and the sediments become finer both upwards and inland as wave energy decreases. Particle orientation often indicates flow directions - lower parts of the unit are orientated inland in the direction of wave run-up, upper sections are orientated seawards by wave backwash.

• Distinct upper and lower sub-units of the sediment representing wave run-up and backwash can be identified, but investigation of recent tsunami deposits indicates that there is still considerable uncertainty about when and where most deposition occurs.

• Each tsunami deposit is separated from the surrounding material by a marked horizon that is normally the result of erosion. The bottom of the deposit has often overloaded and deformed underlying sediments. Particle sizes range from boulders to coarse sand to fine mud. Sediments reflect the nature of the material transported by the tsunami, not necessarily that of the area of deposition. Most paleotsunami deposits are recognisable as anomalous sand units in peat sequences, although they can occur as mud units in sand, so care must be taken in interpretation.

• Tsunami sediments tend to have higher concentrations of some or all of sodium, sulphur, (chlorine), iron, calcium and magnesium. These are indicative of saltwater inundation.

• Paleotsunami sediments often include increased numbers of marine to brackish water microfossils compared to sediments above and below the event, but reworking of underlying estuarine sediments may simply produce the same assemblage. The sudden occurrence of deep water microfossils is also characteristic of paleotsunami deposits.

• Individual marine shells and shell-rich units are often present. Sub-tidal species are a useful indicator of deposition by a catastrophic event. Paleotsunami deposits are often associated with buried vascular plant material and/or buried soil.

• Shell, wood and less dense debris deposited last from the waning backwash flow of the wave are often found near the top of the sedimentary sequence.

• Dating of tsunami sediments is problematic, but this is also a useful indication of the catastrophic nature of the event. Best results for dating are from units above and below the tsunami deposit to "bracket" the event. Radiocarbon ages are often equivocal because older reworked material is incorporated in the deposit. Age dating of introduced marine shells is preferred. Optical dating (OSL) is the best method available assuming that the sediments were exposed to daylight during reworking by the tsunami.

• If the event is in the recent past, wave height estimates can be based on indicators such as broken branches (low reliability), stripped bark (reliable- figure 1), and debris/human artifacts caught up in trees/buildings.

Example 1: Wellington Region - Wairarapa Coast, 1855

Sediments exposed in a stream bank to the east of Lake Ferry, southern Wairarapa show many classic tsunami features.

Three fining-upward sequences, each separated by an erosional lower contacts are visible. Large cobbles and pebbles are orientated in the direction of water movement - inland near the base of the unit and offshore near the upper part (backwash). The grain size becomes finer inland and

Figure 1: Warapu village site, Papua New Guinea - Bark stripped off palm tree by sand transport in the tsunami.

Figure 2: Wairarapa Coast. Paleotsunami deposit, fining up sequence from cobbles to silt.

Figure 3: Malol villages area, Papua New Guinea - Patchy veneer of sand overlying vegetation in the foreground (also some dead vegetation on surface). Poles were structural support for a house - these now lean inland in the flow direction of the incoming wave.

Figure 4: Warapu village site - Strip of open ground between trees was the site of the main village destroyed by the tsunami. Rafted debris at the end of the spit was deposited by the waning flow of the backwash passing through the lagoon entrance.

there is a marked change in clast shape at the edges of the deposit.

Pumice clasts and wood fragments are found at the top of each sequence and sub-tidal marine shells are present in the deposit. The general microfossil assemblage is mainly the same as that in the bracketing deposits because the material transported by the tsunami is similar to that found in the area of deposition. Some individual microfossils indicate, however, that they were carried inland by a saltwater inundation.

Dates above and below (and within) the deposit bracket the event, and tie it to the 1800s. Contemporary evidence indicates that the tsunami generated by the 1855 rupture of the Wairarapa Fault involved three massive waves, and this is illustrated by the sedimentary record. It is unclear why this particular section of coastline was so badly hit. The nature of the deposits indicates that wave height was most probably larger than that recorded along other parts of the coast. It is possible that either the morphology of Palliser Bay or some submarine canyon preferentially focused the waves on this section of coastline.

Example 2: Papua New Guinea - 17 July 1998

The immediate effects of this recent tsunami were well researched and photographed, but because the tsunami deposit is represented by one patchy sand unit overlying more sand, there is no clear differentiation at a macro scale. Local people reported that the tsunami comprised three waves but this does not show up in the sedimentary evidence. However, grain-size generally becomes finer inland and sub-tidal marine shells and macrofauna were found on the ground surface. Microfossil data are unavailable at present.

There was a considerable amount of buried plant material and/or buried soil, but much of this material was also found at the very top of the tsunami deposit, resting on the ground surface, as a result of rafting and deposition during waning flows of the backwash. Debris caught up in branches indicates that the maximum wave height was about 17.5 metres above sea level. The tsunami was generated in part by subsidence on land. This is evident from the recently drowned vegetation, and the sudden subsidence of a large block of land that helped to focus the waves onto the affected area.

These Wellington and Papua New Guinea examples offer an interesting comparison of the state of the science with respect to identifying paleotsunami. There are numerous diagnostic criteria in use, but evidence from recent tsunami indicates several problems. For example, sediments are not necessarily deposited, they may be eroded, and secondly the deposit may be indistinguishable from other events unless microfossil and more detailed sedimentary research are undertaken.

Microfossil and detailed sediment work are becoming increasingly important in the identification of paleotsunami. Microfossils in particular appear to hold the key to a better differentiation between paleotsunami and paleocyclones. With paleotsunami, the microfossil signal is purely marine or consists of redeposited nearshore material. Paleocyclone signals comprise significant elements of both marine and terrestrial input since cyclonic activity is normally associated with high rainfall. These differences have been successfully identified in New Zealand at different locations and scientists are now working at a single site to differentiate between paleotsunami and paleocyclone deposits.

It is clearly important to understand the environmental conditions of the study area at the time of inundation, and to realise that some diagnostic criteria or 'signatures' are likely to vary from site to site. For example, if a long-term record of paleotsunami at one location is required for regional planning or risk management, then a more sheltered study site such as a wetland is best. The sedimentary environment of a wetland will retain a record of numerous events without the most recent inundation destroying the evidence of previous ones. On the other hand, if the aim is to record either the last event or the maximum energy and erosional capability of a paleotsunami, an exposed coastal site is required.

Compared with paleoseismic research, paleotsunami studies are in their infancy. Depending upon one's point of view, New Zealand is fortunate in having a wealth of possible paleotsunami deposits and, in the end, that is the point - we need a better understanding of paleotsunami because we are so exposed to the tsunami threat. New Zealand is at the forefront of research in this field, and it is important for its coastal communities that our scientists keep their focus in these areas.

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Tsunami and tsunami hazard

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Tsunami is a Japanese word meaning '*harbour wave or waves*' (the plural is the same as the singular). The word is now used internationally for long period gravity waves generated by a sudden displacement of the water surface. The cause of the sudden displacement is normally a submarine earthquake, but may also include mass flows travelling along the ocean floor, large explosions and sea floor collapses, and the impact of bolides in the ocean. All of these source mechanisms produce an impulse that drives the tsunami. Therefore the term tsunami is strictly confined to long period waves generated by an impulsive source.

This definition excludes meteorological tsunami (rissaga or seebär) and storm surges: phenomena included in the original Japanese definition of tsunami. However meteorological tsunami have very similar characteristics to tsunami, and represent similar hazards along the New Zealand coast. They are long period shallow water waves that behave as tsunami, but they are not generated by disturbances near the ocean floor, or even within the ocean. Instead they are generated by disturbances in the atmosphere through a process called phase coupling. This occurs when the atmospheric disturbance travels at the same speed as a tsunami wave in the ocean, allowing energy to be transferred from the atmosphere to the ocean.

The collapse of the eruption column that forms pyroclastic flows during large volcanic eruptions, also causes the displacement of a large volume of air. This displacement generates atmospheric pressure waves that travel at 220 $m.s^{-1}$, close to the phase velocity of tsunami in the deep ocean. Therefore the pressure waves cause the ocean to resonate and generate tsunami-like waves. The 1883 Krakatau eruption generated pressure waves that travelled around the Earth several times. The pressure waves generated tsunami in many places, including one observed around the coast of the English Channel; and another that reached heights of 1.8 m around the New Zealand coast. The Krakatau pressure waves also caused large lakes in New Zealand to seiche (by up to 20 cm in Lake Taupo).

Another type of wave that is often associated and confused with tsunami, is the immediate wave or surge produced directly by the displacement of the water during an earthquake, mass flow, explosion or impact. This may continue to travel away from the source region as a solitary wave, in which case it is called a tsunami. However, it may just inundate any adjacent coast and not travel at all. This is often the case in confined areas such as harbours. In these places the shoreline is often very steep; usually vertical along wharves and seawalls. If the steep shore moves horizontally, an immediate wave can be formed. These waves are very hazardous. In May 1983 a class of Japanese school children was killed by an immediate wave as they were fishing from a wharf. The wave also lifted the boats alongside the wharf and swept them ashore.

Tsunami characteristics

Tsunami are long period shallow water waves with typical periods ranging from 15 to 60 minutes. Due to their long period they behave as shallow water gravity waves. Hence their velocity is solely a function of water depth as given by:

$$C=\sqrt{gh}$$

where C is the wave phase velocity, g is gravitational acceleration, and h is the water depth. This equations means that the tsunami travels faster in deep water than related to the phase velocity and the tsunami period by

$$C=\frac{L}{T}$$

Due to their high velocities and long periods, tsunami also have wavelengths of hundreds of kilometres in deep water. The energy associated with a tsunami is distributed through the whole water column, regardless of depth, and the entire tsunami wave train. Since tsunami are shallow water waves, the energy also travels at the same velocity as the waves.

Consequently, since the period remains constant, a tsunami travelling into shallower water slows down and increases in size. In the deep ocean the maximum height of a tsunami is less than 0.5 m. Tsunami may be tens of metres in height in shallow water. However, most tsunami are less than 1 m in height at the shore.

It may be difficult to recognise tsunami waves at the shore without the aid of instruments that record the water level over time. This is because the tsunami waves do not occur in isolation, but add to the variations in water level caused by shorter period wind-generated gravity and infragravity waves, and longer period tidal waves. The resulting interactions result in complicated water level changes that may mask the true character of the tsunami waves. It also makes it difficult to define what the tsunami wave height is. Several different measures are used to characterise the height of tsunami waves:

• *peak-to-peak amplitude*, or the difference in elevation between a consecutive wave crest and trough. This may sometimes be referred to as the double-amplitude, which is a more correct terminology if the waves are periodic;

• *zero-to-peak amplitude*, or the difference in elevation between the expected water level (normally expected tide height, ignoring the effect of short period waves) and the crest;

• *vertical runup*, or the difference in elevation between the height reached by the tsunami wave at the maximum inland extent, and the expected tidal elevation. Depending on how the tsunami behaves on reaching the shore, the maximum runup may occur any where between the shoreline, and the maximum inland extent of the tsunami runup (*tsunami inundation*);

• *maximum elevation*, or the difference in elevation between the expected tide level and the highest elevation reached by the tsunami (which need not be the same as the vertical runup);

• *draw-down*, or the difference in elevation between the minimum water level caused by tsunami-induced recession (the wave trough) and the expected tide level. The draw-down may have a greater magnitude than the runup, as occurred in Whitianga and Mangawhai Harbours in response to the 1960 Chilean Tsunami.

Tsunami may be assigned a magnitude to define their relative size. Tsunami magnitude m_t is determined from the non-dimensional vertical runup height near, as given by:

$$m_t=\log_2\left(\frac{R}{R'}\right)$$

where R is the vertical runup height measured near to source, and R' is the reference runup height (1 m). Tsunami magnitude is therefore a logarithmic scale with unequal increases in runup height between steps.
Very few historical tsunami have exceeded magnitude 5 (32 m).

Tsunami generation

Tsunami generation requires the abrupt displacement of a large volume of seawater. The tsunami waves form as the displaced water mass, which acts under the influence of gravity, returns to equilibrium. Theoretically, most of the kinetic energy of the tsunami is derived from the horizontal displacements of the water column and not the vertical displacements. Hence any process that displaces a large volume of seawater is capable of generating a tsunami, particularly if the water is displaced horizontally.

Earthquake tsunami generation

Most tsunami are generated by tectonic earthquakes; a particular type of earthquake associated with crustal deformation and the movement of tectonic plates. Subduction earthquakes are particularly effective in generating tsunami. A tectonic earthquake normally generates a tsunami through seafloor displacement caused by fault rupture. However a few tsunami are produced indirectly by earthquake-triggered secondary mechanisms, such as mass movement or gas hydrate explosion.

Seawater is compressible and unconfined in the deep ocean. Therefore the impulsive stress exerted by the earthquake can dissipate unless the stress is applied rapidly over a large area. This requires that the earthquake be large, and that the hypocentre is located close to the seafloor. The minimum Richter magnitude (M_L) required for a tectonic earthquake to generate a tsunami is 6.3, and there is a general tendency for the size of the tsunami to increase with increasing earthquake magnitude. However predictive equations for tsunami size, based on the Richter magnitude (or related magnitude measures) are not very useful for very large earthquakes, and long duration or slow earthquakes (*tsunami earthquakes*). Tsunami earthquakes produce tsunami that are significantly larger (typically 5-10 times) than would be predicted by the magnitude of the seismic waves.

This difficulty arises because the Richter magnitude does not correctly define the energy available for tsunami generation, because it measured at a constant seismic wave period. Hence the Richter magnitude tends to become saturated (reach a constant value) when the rupture duration is comparable to the period. The total energy available is better defined by directly the seismic moment , which can also be related to the fundamental faulting processes of the rock.

The moment magnitude can also be measured directly from the seismic Rayleigh wave energy in the 50 to 300 s period range, providing a quick method of assessing the seismic moment and the earthquake fault parameters.

Using these improved measures of earthquake magnitude, tsunami risk can be defined (Table 1). It is also possible to predict the tsunami amplitude for specific locations for given earthquake magnitudes. One method commonly used is the Abe tsunami magnitude scale, given by:

$$M_t = \log_{10} a + b = \log_{10} a + C \log_{10} R + D$$

where M_t is the tsunami magnitude, a is the maximum tsunami amplitude (half the trough to crest distance), R is the distance from the earthquake epicentre to tsunami observation site, and b, C and D are site specific constants.

Normally the constant b is chosen so that the tsunami magnitude equals the moment magnitude of the generating earthquake. This allows the tsunami amplitude to be predicted for future earthquakes. Unfortunately the available tsunami amplitude data for New Zealand is very sparse, making it very difficult to determine the values of b. Further, historical tsunami databases, although providing useful information concerning likely future tsunami hazard, are often too short to provide reliable predictions. This is the case for the New Zealand tsunami database.

Numerical modelling of tsunami generation by earthquakes can provide an alternative method of assessing tsunami hazard. This requires an understanding of the processes that affect tsunami generation and their relative importance. The methods discussed above suggest that the critical parameters affecting the amount of energy transferred from the earthquake to the tsunami, and hence the tsunami magnitude, and the focal depth, the area of the ruptured fault plane, and the average displacement. However, a review of seismic parameters affecting tsunami amplitudes found this is not necessarily true at a distance from the source, simplifying the problem.

Three basic types of fault motion are considered to generate tsunami:

• Dip-slip, involving only vertical motion along a near-vertical fault plane;

• Strike-slip, involving only horizontal motion along a near-vertical fault plane;

• Thrusting, involving a combination of vertical and horizontal motion along an inclined fault plane (usually reverse faulting).

All known tsunami generating earthquake focal motions can be defined by a combination of one or more of these three basic types. Most tsunami are associated with predominantly dip-slip fault movement due to a greater efficiency of tsunami generation by a vertical impulse. However thrust faulting may be responsible for many tsunami earthquakes if the faulting displaces the continental slope, or the rupture predominantly occurs in soft sediment. Tsunami earthquakes have tsunami magnitudes considerably larger than indicated by Table 1, so that catastrophic tsunami may be associated by Richter magnitudes as low as 5.

Large earthquakes are normally associated with long, roughly linear rupture zones. Therefore they may be considered as linear wave generators that produce long sequences of waves. This is consistent with the observation that most earthquake generated tsunami consist of a sequence of waves, or wave train, that may persist for 3-5 days. The initial tsunami wave motion at any location depends on where it is relative to the rupture zone. Dip-slip focal mechanisms typically produce both an uplift and down-drop at the seabed, so that the initial wave motions can be either up or down. Normally the initial wave is quite small, and the largest wave occurs later in the wave train. Large earthquakes may also generate very stable tsunami known as solitons, that can propagate across the ocean with negligible energy loss. Tsunami that cause damage at great distances from source (*teletsunami*) are largely composed of solitons.

Landslide tsunami generation

Landslides entering water have always been known to generate large waves and numerous examples have been documented world-wide. Only a few of these examples have been considered to be true tsunami formed without seismic activity, so that landslide-generated tsunami are considered rare. However, there are also a significant number of cases where co-seismic landslides have generated tsunami, often in additional to the earthquake generated tsunami. For example, both the 1931 Napier Earthquake and 1964 Alaskan Earthquake generated a tsunami directly by fault displacement and a larger, but localised, tsunami by landslides induced by the earthquake. The larger than expected tsunami produced by tsunami earthquakes may also be due to co-seismic submarine landslides.

Landslide generated tsunami can be extremely large at source. For example the 1958 Alaskan earthquake triggered a medium rock slide (0.03 km^3) in Lituya Bay, Alaska. The rock slide pushed the water up to a height of 525 m above sea level on the opposite shore of the fjord, resulting in the generation of a tsunami ~30 m high at the entrance of the fjord. Large landslides (1000-5000 km^3) from the flanks of the volcanoes of the Hawaiian Ridge have generated tsunami waves 300-400 m along the coasts of the adjacent islands. Numerical modelling indicates that the resulting tsunami would have had a zero-to-peak amplitude of ~4 m on reaching New Zealand. Keith Lewis discusses evidence for similar landslides along the New Zealand coast in this issue.

Predicting the magnitude of a tsunami produced by a landslide is complicated by the lack of understanding of the processes involved, and a considerable international research effort is now being focused on this problem. Numerical and physical simulations indicate that landslides generate a limited number of waves that are not periodic, that is they propagate independently. Three main waves are recognised as being produced by landslides. The first is a solitary wave (crest) that propagates offshore from the landslide. This wave is followed by a forced wave trough that propagates at the speed of the landslide front. The third wave was a trough that propagates shoreward as a leading depression wave. A leading depression wave is a common feature of historical landslide tsunami. Most models of landslide tsunami indicate that the largest waves travel offshore, which disagrees with many observed events. However it appears that the initial submergence (depth over the head of the landslide) may affect the distribution of energy, so that shallow submarine and subaerial landslides produce a large offshore wave, while deeper submarine landslides produce a large onshore wave.

Landslides normally are much smaller in area than earthquake-induced deformation, and are sufficiently small to act as point sources. Therefore the tsunami waves produced are radially dispersive (like the waves produced by a pebble thrown into a pond) are dissipate rapidly. The tsunami wave train normally consists of only a few waves (3-6), and tends to be aperiodic.

Volcanic tsunami generation

Almost a quarter of the deaths directly caused by volcanic eruptions have been attributed to tsunami generated by the eruptions. This is mainly due to the distance over which the tsunami can propagate compared to other volcanic processes. The tsunamigenic processes associated with volcanism are discussed in more detail elsewhere in this issue, but that can be grouped into four basic tsunami generation mechanisms: earthquake-induced deformation; mass movement (pyroclastic flows, avalanches, lahars and lava flows); cratering (submarine explosions and caldera collapse); and phase coupling (basal surges, shock waves and atmospheric pressure waves).

Normally volcanic earthquakes do not release sufficient energy to generate a tsunami. However, if magma is actively migrating towards the surface, the amount of deformation that occurs at the sea bed may be greater than would normally be associated with a tectonic earthquake. This occurred on the flanks of the island of Hawaii in November 1975, when a large section of the flank of the volcano moved sideways during a volcanic earthquake, generating a large local tsunami, the Kalapana Tsunami.

The mass movement mechanism is the same as that discussed above for landslide. The main differences are that pyroclastic flows move considerably faster than most other forms of mass movement, which will affect the magnitude of the tsunami produced. There is a growing body of evidence that pyroclastic flows can enter water and produce water supported mass-flows that generate tsunami. This is discussed elsewhere in this issue.

The large tsunami (>35 m) produced by the 1883 Krakatau Eruption are often attributed to a large submarine explosion, or to caldera collapse. Both these processes generate tsunami by initially producing a depressed crater-like region at the ocean surface. The initial disturbance forms tsunami waves as the water rebounds. The area affected is usually quite small compared to earthquake displacements, so the "explosion" acts as a point source. Only a short sequence of tsunami waves is produced, and the largest wave is normally one of the first two waves, with the height decreasing rapidly with subsequent waves. The wave period is quite short (<10 minutes), and the first waves are solitary waves. This mechanism is very inefficient at generating tsunami, so a very large amount of energy is required to produce a hazardous tsunami.

Cratering of the ocean surface can also be produced by nuclear and other artificial explosions. During the Cold War a considerable research effort was expended on investigating the use of explosives to trigger tsunami waves. The research showed that generating waves with explosions is extremely inefficient and the resulting waves are small. Further increasing the size of the eruption decreased the size of the waves produced; once the explosion is sufficiently large to create craters that exposed the ocean floor the volume of water available to generate a tsunami decreases. It has been determined that the largest known artificial explosion was still 2 orders of magnitude too small to reach the energy threshold for earthquake tsunami (Table 1).

Gas hydrates have also been suggested as possible causes of cratering. They are unstable mixtures of water and natural gas. Under pressure they form an icy matrix in ocean floor sediments that can collapse explosively to release gas, water and mud. Explosions involving gas hydrates have been suggested as the cause of some large prehistoric tsunami in the North Sea. However, the tsunami may be generated by landslides triggered by the decomposition of gas hydrates.

Finally, large volcanic eruptions can generate meteorological tsunami or rissaga by phase coupling between the ocean and the atmosphere. It has been suggested that basal surges may generate quite large tsunami. However, except for the 1965 eruption of Lake Taal, most of the waves produced by basal surges have been small. The Lake Taal eruption occurred in a confined water body, and the large wave inferred from sedimentary deposits is most likely the result of seiching and ground motions.

Volcanic tsunami have small source regions, so effectively come from point sources. Therefore they are radially dispersive and rapidly dissipate. The tsunami wave train consists of a limited number of waves (1-5) and does not have a strong periodicity.

Impact tsunami generation

The Earth receives a constant rain of material from outer space, equal to a rate of ~100 thousand tonnes per year. Most of this material is very small dust particles, but some occurs as larger bolides (100-300 bolides > 1kg per year). The larger objects may impact the Earth's surface with considerable energy. Depending on the mass of the bolide, an impact in an ocean could produce tsunami waves with heights of several kilometres.

Since an impact is a point source, the waves dissipate rapidly with a corresponding reduction in wave height. However, numerical models indicate that stable solitary tsunami waves with heights of 50-100 m could be also be formed by bolides as small as 200 m in diameter. Bolides of this size strike the Pacific Ocean with return periods of 24,000 to 43,000 years. This means that there is at least a 0.002% chance of an impact in any year. There is good evidence to show that a bolide impact near Yucatan 66 million years ago produced a tsunami that was at least 90 m high along the coast of Texas, and of comparable height along the coast of Brazil.

Tsunami propagation

There are two basic types of tsunami wave train to consider regarding tsunami propagation: the long sequence, periodic and stable tsunami produced by large linear source regions (tectonic earthquakes); and the radially dispersive, unstable short sequence tsunami produced by point sources (most other mechanisms). Except for very large initial tsunami wave heights (»100 m), tsunami produced by point sources will not be hazardous after propagating more than 1000 km from source. However tsunami produced by linear sources can be destructive over much larger distances. There are several processes that can affect the propagation of the tsunami over these large distances, resulting in an increased magnitude and hazard.

Some point sources initially produce an even distribution of wave energy around the generation region. However most mechanisms involve a directed distribution of energy. With mass flow mechanisms the highest energy is along the axis of the flow, and with linear earthquake sources the maximum energy is roughly normal to the rupture zone. Therefore locations along the direction of maximum energy will experience wave heights. For large tectonic earthquakes associated with subduction zones around the Pacific Rim, most energy tends to be directed towards the coast and the centre of the Pacific Ocean.

Refraction occurs when different parts of the tsunami wave train travel at different speeds, and involves the wave crests bending towards the region of slowest phase velocity. The lowest phase velocity for a tsunami occurs in shallow water, so a tsunami will bend towards shallow water. This means that shallow areas can act as lenses to focus the tsunami energy. This occurred during the 1960 Chilean Tsunami when the shallows of the East Pacific Rise focused tsunami wave energy on the New Zealand coast between East Cape and Gisborne.

Shallow ridges may also trap tsunami energy by refraction, so that much of the energy propagates along the axis of the ridge. The Chatham Rise on the east coast of New Zealand seems to act in this way to focus additional tsunami energy on Banks Peninsula. Refraction can also disperse the tsunami and reduce the wave height. Refraction close to shore can produce extreme variations in wave height along short distances of the shore.

Tsunami wave energy can also be reflected by steep gradients on the sea floor, which will further reduce the tsunami energy. However the energy can also be reflected by the shoreline, to be trapped by refraction to form a dispersive edge wave that travels along the coast. This is most likely to occur close to source when the tsunami is generated on a continental shelf.

Due to the long distances that a stable tsunami wave train may propagate over, the tsunami is affected by the Earth's rotation (*Coriolis Effect*). The rotating Earth causes an apparent deflection of the tsunami waves towards the left in the Southern Hemisphere, and the right in the Northern Hemisphere. This effect causes a focusing of wave energy as the tsunami crosses the Equator, contributing to the tsunami hazard in Hawaii. The curvature of the Earth also becomes important; as the tsunami waves leave the source they tend to spread out due to the curvature of the Earth. However if they travel far enough they begin to converge again, so that if the Earth was entirely ocean they would all arrive together at the antipodal location from the source. The curvature effect increased the magnitude of the 1960 Chilean Tsunami when it reached Japan.

As a tsunami enters shallow water it undergoes further shoaling transformations similar to those affecting swell and surf. These can alter its characteristics considerably. In particular each wave can break up into a series of solitary waves, just like swell waves decompose in the surf zone at the beach. The individual waves of tsunami can behave in two ways when they reach the shore:

- as a non-breaking wave that behaves like a large rapidly rising and falling tide. The maximum runup is equal to the height of the wave when it reaches dry land, although it can be higher if the coast is very steep;
- as a breaking wave or bore. This is the type of wave many people associate with a tsunami. The maximum runup is normally less than the height of the bore when it reaches dry land.

Most tsunami reaching the New Zealand coast have behaved as a non-breaking wave, although they have tended to form bores within estuaries.

The shoaling effects on a tsunami vary with the local bathymetry and the direction the waves are travelling. Similarly the propagation effects in deep water also vary with bathymetry and travel direction. Hence the behaviour of any given tsunami can vary considerably along the coast. The tsunami effects at any location are also likely to differ depending on the tsunami deep water approach direction.

However, all tsunami from the same source region will propagate in the same way, undergoing the same transformations. Further, numerical modelling indicates that the tsunami amplitudes will scale linearly with the initial tsunami height, so doubling the initial height will double the final height at all coastal locations. This feature has been utilised to facilitate real time warning systems, by pre-computing the coastal tsunami wave height distributions for all expected tsunami source regions using a reference initial tsunami wave height. When a tsunami occurs, the values are scaled by the actual initial tsunami wave height to give a prediction of the expected wave heights.

Tsunami hazards

A tsunami interacts with the coast to produce a variety of hazards. The hazards created are specific to any section of coast. For example the 1993 Hokkaido Nansei-Oki tsunami varied in height from 5-30.5 m over 500 m of the coast of Okushiri Island, Japan. This caused a large variation in the amount of damage sustained. The hazards will also vary between tsunami events since tsunami are rarely generated from exactly the same source in the same way. Therefore the tsunami will behave differently as they travel.

The potential tsunami hazard is normally evaluated by the maximum tsunami wave runup. This runup can be measured as either the vertical height that the wave reaches, or the horizontal distance the wave floods inland (inundation). The inundation distance depends on how high the wave is at the shore and the local topography. Therefore the vertical runup is used most often. The hazard increases with increasing runup, and any runup exceeding 1 m is considered to be potentially catastrophic.

Tsunami bores

The most destructive tsunami are those that form a breaking bore due to the transfer of momentum to the still water trapped in front of the bore. This may strange since a breaking wave is losing energy and should therefore have less energy at the shore than an equivalent wave that does not break. However the damage is mainly caused by the high horizontal and vertical turbulence in the wave, so the available energy is used more effectively to inflict damage. The vertical turbulence of tsunami bores can lift and carry quite large objects; bores associated with the 1960 Chilean tsunami transported 20 tonne pieces of a Japanese seawall up to 200 m inland.

Tsunami bores may also form within estuaries and the lower reaches of rivers and streams. These have been common features of tsunami in New Zealand, and they have caused most of the severe damage. It used to be common practice to build road and rail bridges at the upper limit of tidal influence on coastal rivers and streams. This provided some protection from floods coming down the channel. Unfortunately this is also the position where the tsunami bores are strongest. Quite a few coastal bridges and their approaches have been damaged or destroyed by tsunami in New Zealand. For example, one of the 1947 Gisborne tsunami formed a bore that carried the main beams and deck of the Pouawa River Bridge 1.5 km upriver.

Floating debris

During the 1993 Hokkaido-Nansei-Oki tsunami most deaths (71%) and injuries were due to the impact of floating debris. Studies have demonstrated that debris carried by tsunami can generate very high impulsive forces; floating wood pushed by tsunami bores may exert impulsive forces of more than 9 tonnes. This is higher than many structures can withstand.

Tsunami may also spread liquid contaminants such as oil. This is of particular concern in port areas, especially due to the high concentration of combustible contaminants such as fuel oils, diesel and lighter hydrocarbons, and other hazardous chemical compounds. In New Zealand, many refuelling facilities in ports and marinas do not take any measures to protect fuel supply pipes from the effects of tsunami.

The most hazardous floating debris appears to be small boats in marinas or fishing ports. In Japan small fishing vessels swept inland by tsunami waves have been a major cause of fires associated with tsunami. The fires are often caused by overturned gas cooking appliances, and the extent of the problem is dependent on the number of persons living on board small vessels.

Combustible materials carried by tsunami may also be ignited by sparks from electrical equipment as they are inundated by tsunami waves. The 1960 Chilean Tsunami caused several electrical failures at Lyttelton, but no major fires resulted.

Inundation and return flow

The current velocities generated by flooding and receding tsunami waves can be high due to seem extreme variations in water level. The 1993 Hokkaido Nansei-Oki tsunami produced flows with velocities of 10-18 $m.s^{-1}$. Most drownings associated with tsunami have been of persons swept into deep water by the return flow. The return flow may also carry floating debris with the same potential for injury and damage as an advancing tsunami bore. The high current velocities make tsunami very erosive. The velocities are difficult to predict since erosion changes channel characteristics. In confined bays and regions with islands, the interaction of refracted and reflected waves can produce vary complex patterns of currents and waves.

Responses of harbours, estuaries and rivers

Tsunami waves may force oscillations within semi-enclosed basins such as estuaries, harbours and the lower reaches of rivers to produce seiches. Wellington Harbour has several natural modes of oscillation, and can resonate in response to tsunami excitation. This can increase the height of the tsunami at some locations around the harbour, particularly Evans Bay. Similarly Pegasus Bay near Christchurch has a natural mode of oscillation of 3.4 h, which has amplified historic tsunami. The waves recorded following the July 1998 Saundaun Tsunami increased from a zero-to-peak amplitude of ~10 mm at Kaikoura, to 100 mm at Sumner Head. Further amplification occurs in Lyttelton Harbour, which increased the zero-to-peak amplitude to almost 150 mm for these waves. This amplification partially explains why the 1960 Chilean Tsunami had a maximum peak-to-peak amplitude of 5.5 m at Lyttelton, while most of the New Zealand east coast experienced 1-2 m waves.

A tsunami causes fairly rapid changes in water level within estuaries. Even if the change in istance the wave floods inland (inundation). The inundation distance depends on how high the wave is at the shore and the local topography. Therefore the vertical runup is used most often. The hazard increases with increasing runup, and any runup exceeding 1 m is considered to be potentially catastrophic.

Tsunami bores

The most destructive tsunami are those that form a breaking bore due to the transfer of momentum to the still water trapped in front of the bore. This may seem strange since a breaking wave is losing energy and should therefore have less energy at the shore than an equivalent wave that does not break. However the damage is mainly caused by the high horizontal and vertical turbulence in the wave, so the available energy is used more effectively to inflict damage. The vertical turbulence of tsunami bores can lift and carry quite large objects; bores associated with the 1960 Chilean tsunami transported 20 tonne pieces of a Japanese seawall up to 200 m inland.

Tsunami bores may also form within estuaries and the lower reaches of rivers and streams. These have been common features of tsunami in New Zealand, and they have caused most of the severe damage. It used to be common practice to build road and rail bridges at the upper limit of tidal influence on coastal rivers and streams. This provided some protection from floods coming down the channel. Unfortunately this is also the position where the tsunami bores are strongest. Quite a few coastal bridges and their approaches have been damaged or destroyed by tsunami in New Zealand. For example, one of the 1947 Gisborne tsunami formed a bore that carried the main beams and deck of the Pouawa River Bridge 1.5 km upriver.

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A tsunami causes fairly rapid changes in water level within estuaries. Even if the change in the early hours of the morning, with minimal effort from emergency services. This system still requires 10 minutes warning of the tsunami arrival.

Therefore the main priority for tsunami mitigation in most countries is the development of an effective warning system. Most countries experiencing tsunami have a dedicated national organisation handling tsunami warnings. There are usually two operational systems: an international warning system such as the Pacific Tsunami Warning Centre (PTWC); and local warning systems, such as that operated by the Japan Meteorological Agency (JMA).

The concept of PTWC was developed following the Aleutian Tsunami in 1946 that caused severe damage in Hawaii, and expanded rapidly following the 1960 Chilean tsunami. The present PTWC was established in 1965 and is designed to handle teletsunami affecting the whole Pacific Ocean, as well as dealing with locally generated tsunami affecting Hawaii, and tsunami affecting
the territories and possessions of the USA.

There are several stages involved in the handling of teletsunami warnings:

- A network of seismometers is linked to a central processing facility. When large earthquakes occur, their epicentre, focal depth and magnitude are calculated. This presently takes 15-30 minutes, and it is hoped to reduce the time significantly. If the earthquake is considered capable of generating a tsunami, the PTWC notifies the national organisations in countries likely to be affected. This may take 30-60 minutes.

• The PTWC collects tide gauge data from the regions surrounding the earthquake. These provide the necessary evidence of the presence or absence of a tsunami. The PTWC also runs computer simulations to determine the likely travel times and wave heights given the earthquake location and size. These are continually up-dated and the results passed on to the affected countries.

• The national organisations evaluate the available information and determine which areas will be affected and require evacuation.

The PTWC communicates to the national organisations by way of 4 types of messages:

• *Tsunami Dummy* messages used to test the communication networks at regular intervals.

• *Tsunami Information Bulletin* messages that convey information about major earthquakes that are not considered to be capable of generating, or have not generated, a tsunami. Normally it includes seismic data on the earthquake.

• *Tsunami Watch Bulletin* messages provide information about earthquakes that are likely to generate, or have generated, a tsunami. They include expected arrival times for the first tsunami wave for pre-defined locations in all affected countries. A watch status is applied to all locations more than 3 hours travel time from the expected position of the first tsunami wave at the time of transmission. Normally this means that after an earthquake that may generate a tsunami, all locations 4-6 hours travel time from the epicentre are put on watch status.

• *Tsunami Warning Bulletin* messages advise all locations within 3 hours travel time of the expected position of the first tsunami wave that a tsunami is likely.

Over time the tsunami warning and watch zones expand, and the PTWC sends update messages every hour. The update messages include additional arrival time data, and include any available data on the characteristics of any tsunami waves detected. To assist with this aspect, the PTWC is in satellite communication with a network of tide gauges around the Pacific. Most of these are in the USA EEZ, and there are very few in the Southwest Pacific. One station is maintained on the Chatham Islands for the PTWC by the Ministry for Emergency Management. This station is the main gauge that can provide New Zealand with inform-ation about hazardous tsunami from South America.

This system works well for areas that are sufficiently far from the source to allow for the delays. It is almost ineffectual for local tsunami. Many countries around the Pacific Rim also maintain a national tsunami warning system to handle local tsunami. New Zealand does not.

M_0 (dyn-cm)	M_w	M_m	Risk
$<10^{27}$	<7.3	<7	No tsunami risk
10^{27}-10^{28}	7.3-8	7-8	Tsunami unlikely to be directly generated but secondary processes may produce a large localised tsunami
10^{28}-$5x10^{28}$	8-8.4	8-8.7	A tsunami will probably be generated but should only be locally catastrophc
$5x10^{28}$-$2x10^{29}$	8.4-8.8	8.7-9.3	Probable generation of a destructive tsunami with catastrophic effects distant from the source
$\geq 2x10^{29}$	≥8.8	≥9.3	Very likely generation of a destructive tsunami

Table 1. Tsunami risk levels for the South Pacific Ocean, based on the methodology employed by the Centre Polynésian de Prévention des Tsunamis. A catastrophic tsunami is defined to have peak-to-peak amplitudes exceeding 1 m. M_0 is the seismic moment and represents the total energy released by the earthquake. M_w and M_m are the moment and mantle magnitude, which are used as more convenient measures of the earthquake size than the energy released.

Huge Submarine Avalanches: is there a risk of giant waves and, if so where?

by Keith Lewis*, Jean-Yves Collot† and Derek Goring*

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Submarine avalanches can be enormous – orders of magnitude bigger than anything on land. They produce giant tsunamis capable of devastating coastal area to hundreds of metres above sea-level. Evidence of massive and not quite so massive submarine avalanches has recently been found around New Zealand. What do we know about them? Are they really a risk? If so, where?

Big onshore slides and avalanches can involve the catastrophic collapse of hundreds of millions of cubic metres of soil and rock. Huge ones may even involve many cubic kilometres and include hill-sized blocks. What may be the world's largest involves about 26km^3 and occurred in Fiordland. (Fig. 1)

In contrast, submarine "landslides" can involve many thousands of cubic kilometres of seabed. Obviously, submarine slope failures would be disastrous for any man-made structure, such as oil installations, pipes and cables, sited on the failed block. Loss of life and property is potentially much worse if failure propagates onshore into a harbour or town. It is also likely to be the end for anything in the run-out path below the landslide, although in deep water this may mean only submarine telecommunication cables. Potentially far more devastating than the direct effects of a submarine landslide are the large tsunamis they generate.

Perhaps the most dramatic evidence of a tsunami generated by a submarine landslide is from Hawaii. Some 100 000 years ago, the western flank of the 5-6 km high submarine volcano that protrudes from the sea as the island of Hawaii (Fig.2) collapsed. It generated a wave that washed white coral boulders to a height of nearly 200m up the black lava slopes of a neighbouring island and removed the thick, red, tropical soils to a height of over 370m above sea level. It has been suggested that the same wave devastated coastal areas to a height of at least 15m more than 7000 km away in Australia, although this is debated.

Nevertheless, it is indeed fortunate that Hawaii, and indeed much of the Pacific region, were uninhabited at the time. The huge run-up that tsunamis reach close to their sources was demonstrated in 1958 by a landslide into a mercifully unpopulated Alaskan fjord. Although the initial wave it generated was perhaps 60-100m high, it washed over a nearby spur to a height of 524m as it travelled down the fjord at 160-200km/hour.

Over the last two decades, abundant evidence has emerged of very large submarine landslides at many places around the world. Are they capable of generating large tsunamis? Do they occur around New Zealand? What is the risk?

One reason for prodigious submarine slope failures is that there are some prodigious submarine slopes. With two preferred levels of the earth's surface, the continentals within a few hundred metres of sea level, and the abyssal ocean basin at around 5km deep, there is a steep scarp, the continental slope, between the two. Beneath the sea, porous rocks and seabed sediments are naturally saturated with water, which tends to lower their strength compared with rocks and soils on land. At many places, this pore water includes small amounts of gas, which lowers the strength of slope sediments still further. Particularly where susceptible layers are undercut by submarine canyons and faults, or compressed at plate boundaries, continental slopes may become critically over-steepened. They become a disaster waiting to happen.

Knowledge of large submarine "landslides" is surprisingly new. The phenomenon has been inferred for almost a century from masses of jumbled blocks recognised in ancient marine strata high in the European Alps. It is only with the recent advent of "swath" techniques to rapidly map vast areas of seabed that we have realised the enormous size of slope failure on the modern seabed. Because of this rapid increase in understanding, even the terms for submarine "landslides", a somewhat inappropriate, general term for a wide variety of offshore slope failure, is still in a state of flux. In most recent articles, the term slump refers to largely intact back-tilted blocks that move and rotate on a curved glide plain that can be as much as 10 km beneath the seabed. Most slumps are thought to creep slowly rather than collapse catastrophically and they probably don't generate large sea-waves. What may be the world's largest submarine slope failures off South Africa, where individual failures reputedly exceed 20 000km^3, are probably of this type. Similarly, the term slides is commonly confined to the slow creep of largely intact sheets of soft surface sediment on low slopes. They occur on slopes of less than half a degree off Hawkes Bay, where 720km^3 slid seawards. In contrast, debris avalanches are catastrophic events, with broken, disaggregated blocks of slope rocks, some may be kilometres across, that plunge down steep submarine slopes. The

Figure1. - Bush cover marks blocky slide debris (SD) that collapsed from a head scarp (HS) along the mountains in the distance, in what may be the biggest onshore debris avalanche on earth, the Green Lake (GL) Landslide of Fiordland. Enormous by onshore standards , it involved the catastrophic collapse of about 26km^3 of mountain side about 13 000 years ago. The bush covered ridge in the foreground includes debris "bulldozed" in front of the main slide-blocks. (from Hancox, G; McSaveney, M; 1999: Land Stability http://www.gns.cri.nz)

largest avalanche off Hawaii took away the northern side of the island of Oahu and most of the 5 km high volcanic slope to a depth of 2 km below the original seabed (Fig.2). The avalanche moved fast enough to extend out 160 km from the toe of the slope, despite an uphill climb of 300 m. Its total volume is estimated to be about 5,000 km^3. The largest block in the avalanche is 1800 m high, and 30 x 17 km across, with a volume of over 900 km^3., making it larger than Mount Ruapehu. Nearby there are at least 10 other blocks 10 km or more across. Like the much later avalanche that threw the coral blocks high up onto Lanai, south of Maui, this enormous event would have generated a tsunami capable of inundating not only much of the Hawaiian Islands but perhaps many low islands and coastal lowlands throughout the Pacific. Big avalanche deposits occur off ocean islands in the Indian and Atlantic oceans as well as off the continental margins of Norway, West Africa and South America. Catastrophic failure can occur on many scales. Smaller events with much smaller blocks and fluid sediment may be referred to as a debris flow, and these can occur on their own or in front of an avalanches. Debris flows are common on slopes with high sedimentation rates, such as deltas, and one may have contributed to the Papua New Guinea tsunami. The more fluid ones may absorb water into their head and some metamorphose into a turbidity current. These mud and sand charged currents can travel at high speed over near flat submarine plains, for vast distances, cutting submarine cables as they go.

So, is there any evidence of big submarine debris avalanches around New Zealand? The answer is very definitely, yes! Recent joint programmes between French and New Zealand marine scientists, using state of the art "swath" mapping equipment, has revolutionised our view of the seabed in key, crustal-plate boundary areas in the Hikurangi Trough, off eastern North Island and northeastern South Island, and also southwest of the South Island (Fig. 3). In Hikurangi Trough particularly, slope failures have occurred on a vast scale.

The big one is off Ruatoria. (Fig.4) There, over 3 600 cubic kilometres of continental slope have collapsed as a blocky debris avalanche into the 3 500 m deep Hikurangi Trough. The avalanche travelled about 45 km out across its flat trough floor and a more fluid debris flow continued out in front of the avalanche to more than 100 km from the toe of the slope (Fig. 5). Within the avalanche deposit, a dozen or so blocks are more than 500 m high and more than 5 km across. The largest, Ruatoria Knoll, is over 1 000 m high, 17 x 13 km across, making it the size of Mount Ngaurahoe, and it has travelled over

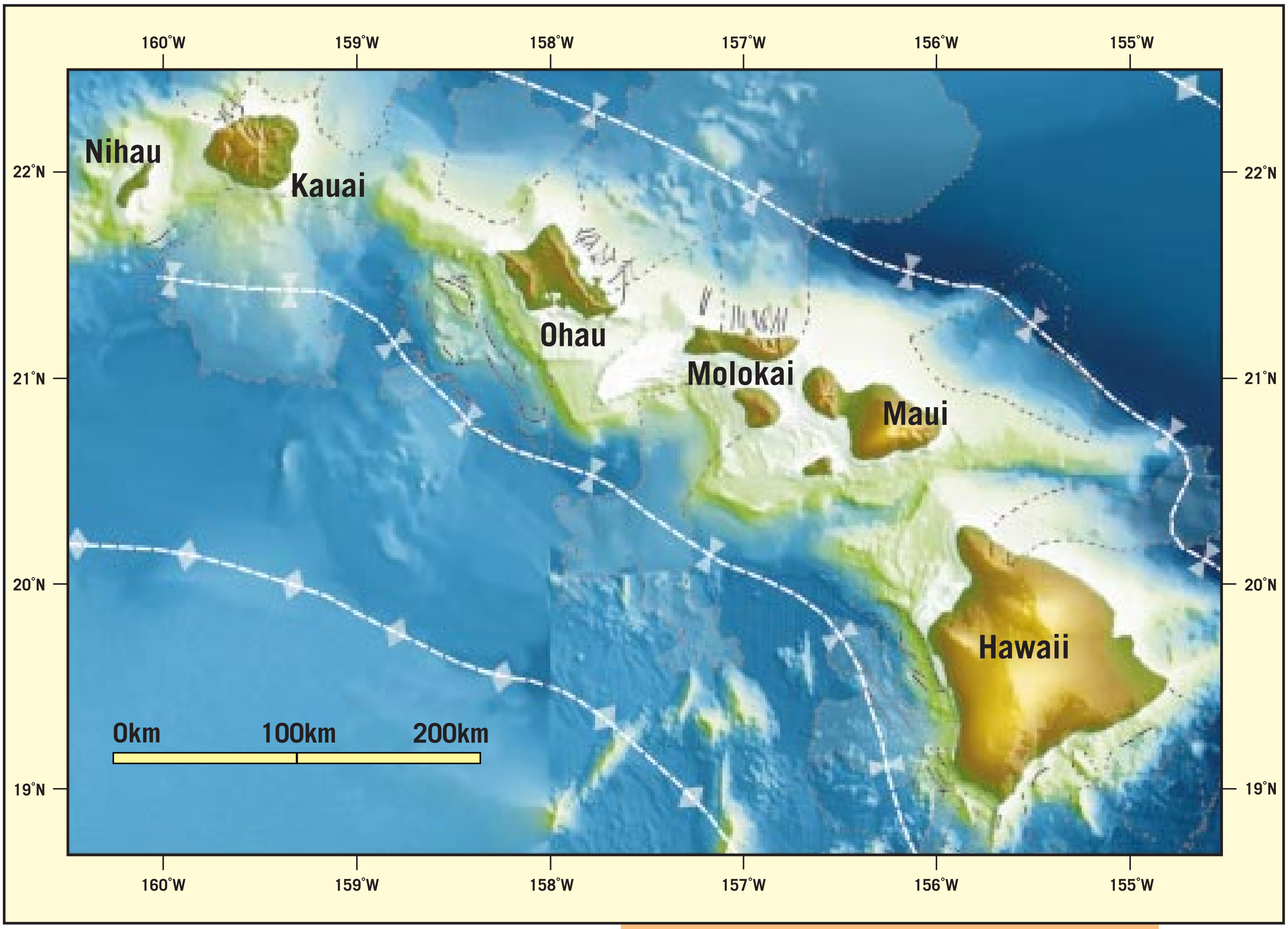
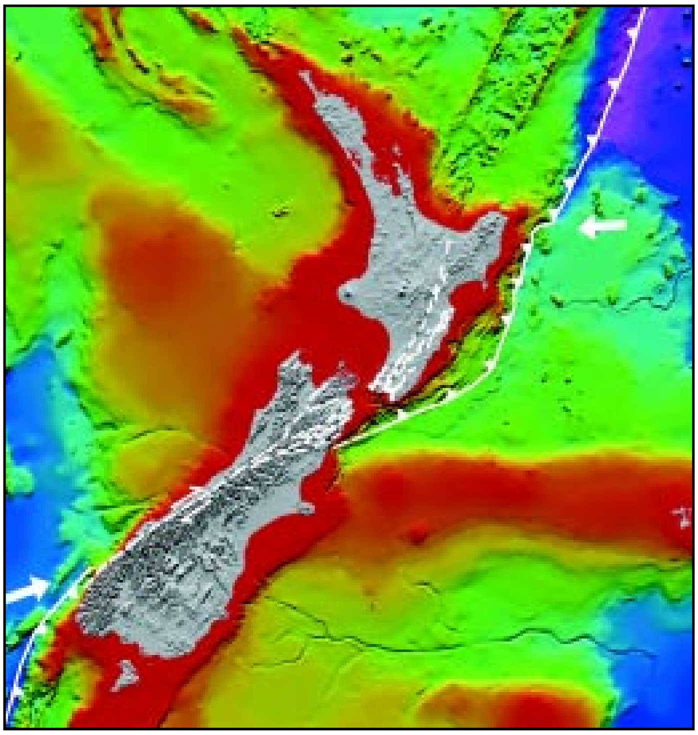


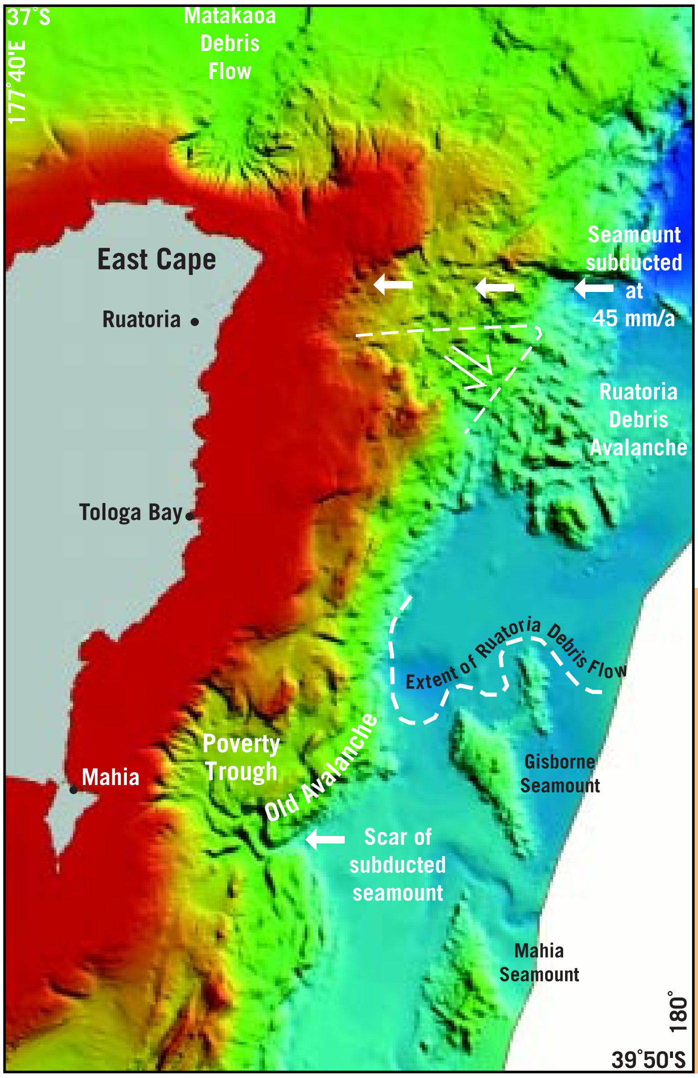
Figure 2. - Dotted lines delimit some of the many enormous submarine avalanche deposits around the Hawaiian Islands. Each avalanche infilled the axis of the 5km deep Hawaiian Deep (broken line with inward pointing flags) and extended for up to 100km up the flank of the 4.5 km deep Hawaiian Arch (broken line with diamonds). The largest avalanche, northeast of Oahu covers an area of 23 000km^2, has a volume of 5 000km^3, and has individual blocks larger than Mount Ruapehu. One of the younger avalanches at the southeastern end of the chain is believed to have generated an enormous tsunami that devastated neighbouring islands to a height of 370m above sea level. (limits of avalanches from Moore et. al Journal of Geophysical Research, 1989. Bathymetry from SOEST, Hawaii).

40 km from the toe of the slope. The Ruatoria Avalanche, together with the scar above it, covers an area of 4 000^2 km, making it similar in size to the Coromandel Peninsula. The debris flow covers another 8 500 km^2, some of it disturbed by the push of the avalanche behind it. If we imagine, for a moment, the Coromandel Peninsula suddenly cascading down a three kilometre high slope, we get some idea of magnitude of the event and of the amount of overlying water that might be displaced to form a tsunami.

This is indeed a horrifying prospect, but what are the chances of such a huge event happening in our lifetime? To make any realistic assessment of risk, we need first to understand what causes massive avalanches and how they generate tsunamis. Although large avalanches can occur on many types of margin, the Ruatoria Avalanche overlies the boundary between two of the earth's major crustal plates (Fig.3). The landward edge of the Hikurangi Trough marks the site where deep, "oceanic" conveyer belt Pacific plate is diving beneath "continental" crust at the edge of the Indo-Australian plate and "subducting" back into the earth's interior. It is disappearing under the continental slope margin off Ruatoria at about 45 mm per year – roughly the rate at which toe-nails grow. However, the subducting Pacific plate is not smooth. It has long-extinct volcanic cones protruding from it. Some of these form seamounts that rise more than 1 000 m above the adjacent abyssal seabed (Figs 3,5). When these reach the margin, they are not scraped off (at least initially), like conveyor-belt luggage, but plough into and under the steep margin as though it was made of crumbly Feta cheese. The process has been eloquently illustrated by the scars left by three small seamounts that are at various stages of impacting the Costa Rica margin. (Fig 6)

As a seamount first disappears into the margin it leaves a V-shaped notch in the toe of the slope, and it raises and crumbles the margin above and in front





of itself. As it penetrates further, like a slow-motion bullet, the seamount does not leave a bullet hole because the Feta-like rocks of the margins collapse back into the hole. The result is that the seamount leaves a trough in its wake to mark its passage. Similar troughs, often with small slumps and avalanches, occur behind subducting seamounts off New Caledonia, Japan and South America. Off Costa Rica, the direction of impact and the trough are perpendicular to the margin. Off the East Coast of New Zealand, they are oblique (Fig.3,5). This is probably the reason for the huge avalanche off Ruatoria. We think that a big seamount ploughed into the continental shelf off East Cape, first cutting a steep sided trough obliquely across the slope and leaving an unstable triangle of rock on its southern side, perched high above the Hikurangi Trough (Fig. 5). It was mainly collapse of this badly fractured triangle that is the reason for the huge size of the avalanche off Ruatoria.

Was it a single event or is it likely to happen again, and if so, how often? From here the news gets better. Judging by the thickness of sediment burying the debris flow, the Ruatoria Avalanche may be more than one hundred thousand years old - although the scientists involved are still arguing about its age. If indeed, subducting seamounts are the destabilising feature of the east coast margin, how often are giant avalanches likely to occur? Unlike the margins of most continents where there is no plate boundary and the record of old landslides are preserved for eons, such forms are transient on an active plate boundary margin. Faint oblique scarps on the East Coast continental slope may be the ghostly scars of earlier seamount impacts (Fig.5). Dating of these features is very uncertain but we surmise that

Figure 3. - A major crustal plate boundary (flagged line) is responsible for destabilizing much of the continental slope off the eastern North Island as the 'conveyor belt" Pacific plate "subducts" beneath the continental edge of the Australian plate at around 45mm per annum. This type of convergent plate boundary links to a mirror image one off Fiordland via the Alpine Fault.

Figure 5. - The East Coast margin shows the effects of many seamount impacts and margin collapses. Off Ruatoria, a debris avalanche deposit extends 45 km across a flat plain and a now buried fluid debris flow in front of it extends more than twice as far. They were derived from a triangle of rock at the southern edge of a trough that formed in the wake of a seamount passing obliquely beneath the margin. Several transverse scars and troughs on the margin may be the ghosts of old seamount-impacts and a blocky lower slope off Mahia may be an old avalanche that has been carried back to the margin by westward motion of the Pacific plate. In the south, a slope-toe indentation may indicate where a seamount has recently impacted the margin, there being, as yet, no collapse in its wake. On the Pacific Plate to the east, there are several seamounts, Gisborne Seamount and Mahia Seamount, that have yet to impact the margin.

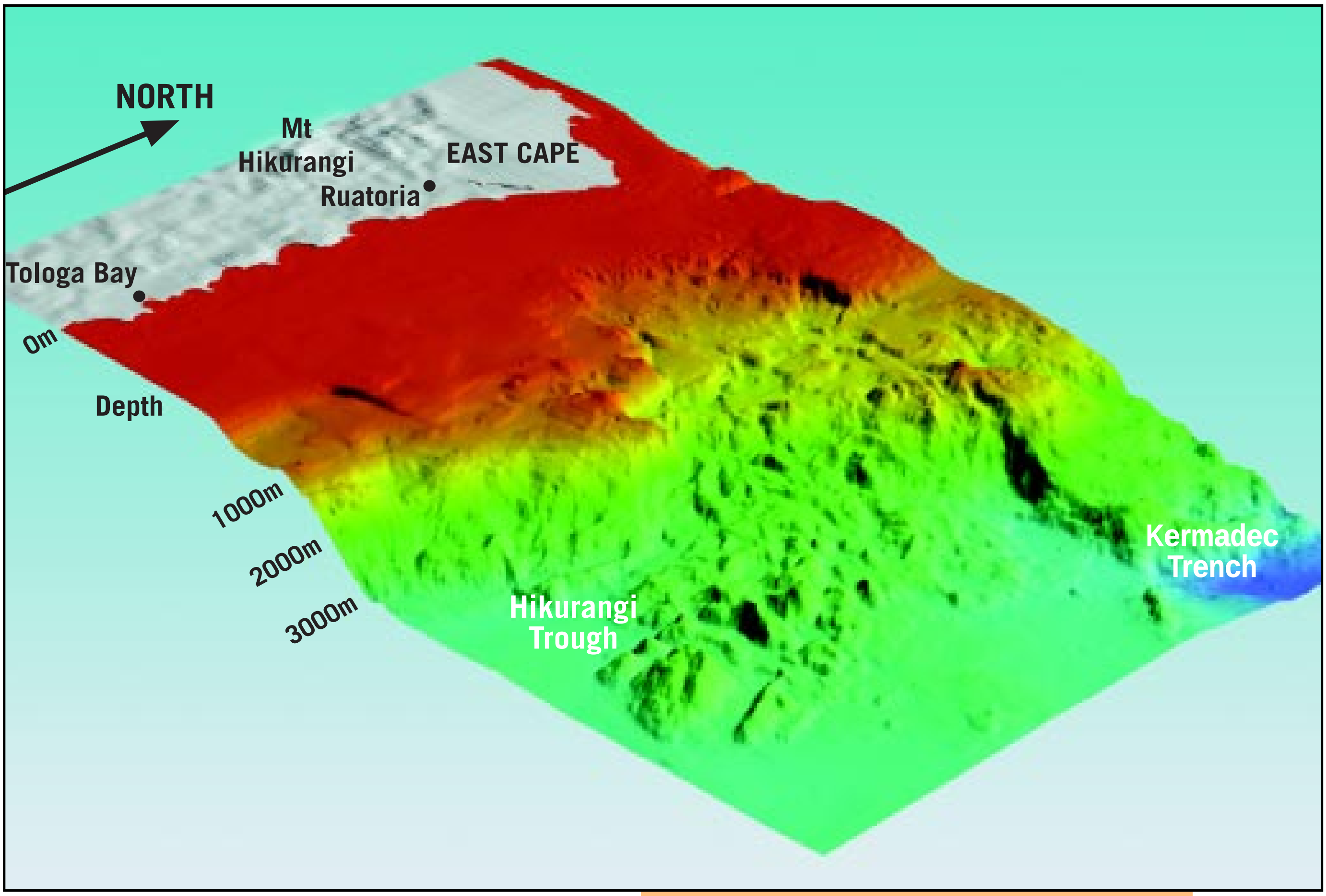


Figure 4. - A debris avalanche deposit off Ruatoria is enormous. An area as big as Coromandel Peninsula collapsed into the 3.5km deep Hikurangi Trough. Blocks the size of Mt. Ngaurahoe swept 40km out across the flat plain. The age of the avalanche is still unknown but it may be more than 100 000 years old. Collapse probably occurred after a large seamount on the Pacific plate was dragged into and under the margin causing the straight scar off East Cape and fractured rocks to the south. It probably generated a tsunami. We have no evidence of how big but suspect that it was very large.

large impact structures, and their resultant avalanches, occur at intervals of several hundred thousand years (Fig.7). Future impacts are implied by seamounts approaching the margin on the Pacific plate to the east (Figs. 3,5).

Their scatter gives an indication of the frequency of large impacts and slope failure, which again is in the order of several hundred thousand years. Although avalanches are clearly a rare event in a human time scale, we should perhaps mention that the margin of Mahia has not yet collapsed - not from the most recent impact there anyway. Also there are likely to be many smaller impacts and smaller wake avalanches than big ones.

Not far away, on the north-facing slope off East Cape, there is another large margin indentation, the Matakaoa indentation with a 500km^3 debris flow, the Matakaoa debris flow, downslope from it (Figs. 5,8). With its northerly orientation, this margin collapse can not be a direct result of seamount collision as seamounts do not impact from the north. However, it may be an indirect effect as it is in line with the axis of most rapid uplift of the onshore ranges, which are thought to be pushed up by material, including seamounts, that are scraped off the downgoing plate beneath the ranges. This uplift may have increased the inclination of the northern slope and reduced its stability. Judging by the ages of ashes in the thin blanket of mud overlying it, the debris flow occurred about 50 000 years ago.

From what we know at present, the risk from big submarine landslides is greatest from Hawkes Bay north to Bay of Plenty, and perhaps also on the steep active margin off Fiordland. We suspect that the giant tsunamis they might generate would devastate nearby coasts and low lying areas to north and south. Although the risk of such large local events is very low, similar waves can be generated around the Pacific and tsunamis can travel enormous distances across the open oceans, albeit with loss of energy, as they radiate outwards. For instance, even the relatively tiny 1998 Papua New Guinea tsunami was felt in New Zealand. Despite this, the risk of a giant landslide-generated tsunami affecting New Zealand coasts in our lifetime is still small. Nevertheless, submarine "landslides" come in many sizes and some of the smaller ones,

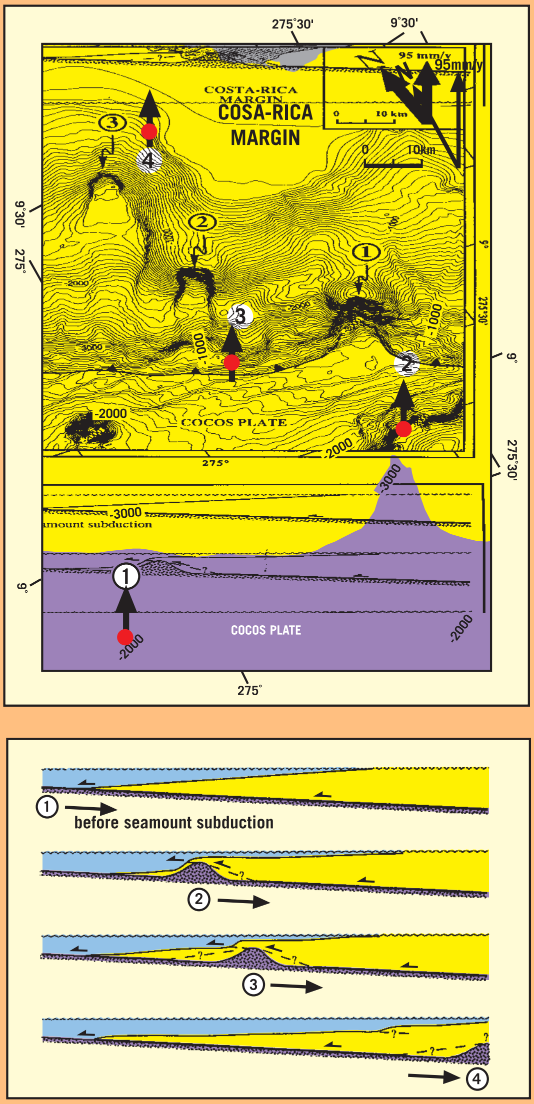


Figure 6. - Impact-scar troughs gouged by seamounts. This classic example shows three small seamounts at different stages of subduction (1,2 and 3) beneath the continental slope off Costa Rica. Below are sections through the margin for the three seamounts, There is minor slope failure behind the seamount at stage 1, and development of wake trough across the margin at stages 2 and 3. Impact is straight into the margin at 95mm per year.
(from Lallemand et al. Journal of Geophysical Research, 1994)

which are still large by onshore standards, may be very much more frequent and more of a danger.

Recent studies of the Kaikoura Canyon show that its head cuts to within a few hundred metres of the beach (Fig. 9). There it traps a nearshore "river of sand" which moves northwards along the north Canterbury shelf under the influence of large swells and ocean currents. The canyon head intercepts about 1.5 million m^3 of sand, mud and gravel each year. These build up as an unstable "rubbish tip" in the steep canyon head. Probably when shaken violently by large earthquakes associated with movement of nearby plate boundary faults, this fill collapses and sweeps down the canyon in a fluid debris flow. Each debris flow simultaneously incorporates water into its leading edge and dumps gravel behind, to become a very fluid, turbulent, high velocity, and possibly self perpetuating, turbidity current. A radiocarbon dated twig extracted from the most recently deposited gravel layer in the lower canyon suggests that the last catastrophic collapse occurred in about AD1830, which was before the earliest historical records of the area. This date coincides with the last major rupture of the nearby Hope Fault, as estimated from the growth of lichens on mountain landslides along the fault. The next oldest gravel deposit has a twig dated at AD1700, and again there is evidence of onshore fault rupture at about this time. If indeed, collapse in the canyon head coincides with large local earthquakes, which are estimated to occur every 100 – 200 years, then the volume of sediment that fails each time is probably of the order of 150 – 300 million m^3, which is only about a quarter of a cubic kilometre. This is tiny compared with the 3 600 km^3 off Ruatoria. We suspected that failure of such volumes in shallow water would have generated a tsunami big enough to have impacted local communities. Unfortunately, the 1830s were a time of great social upheaval for the inhabitants of the nearby coast and few stories of those days survive.

We may be living at a fortunate time in geological history, at least as far as the danger of large submarine slope failure is concerned. We live only about 6 000 years after the end of a sea-level rise of about 120 m that resulted from melting of glaciers after the last ice age. Since then, much of the gravel, sand and mud that reaches the sea in rivers, has remained on the inner parts of wide, wave-planed continental shelves that extend offshore to about 120 m deep. Canyons, like the one at Kaikoura, that incise the inner shelf and trap sediment moving along the shelf are rare. In contrast, during periods of glacially lowered sea level, rivers carried big loads from a devegetated landscape across a wide, exposed, wind-swept continental shelf to dump their load of sand, mud and gravel into the many canyons that incise only the edges of continental shelves. In those times, Kaikoura-type collapse of sediment in canyons-heads would have been much more common.

There may also have been times during the glacial changes of sea level when gas, which drastically reduces sediment strength, was more common in slope sediments. Ice-like methane clathrate (also called gas hydrate), consisting of methane molecules trapped in a water lattice, presently infills the pores of many slope sediments, so that the rock is impermeable to free gas generated beneath it. The contrast between clathrate and gas infilled pore spaces produces a strong reflector that cuts across reflectors from geological strata (Fig.10). The rather strange clathrate lattice is stable only in a narrow range of temperature and pressure conditions and, hence, depth. Thus the reflector from the gas/clathrate interface remains at a fairly constant depth beneath the seabed. When sea level falls at the start of an ice age, the water column above slope clathrates, and hence the pressure on them, is reduced. The clathrates become unstable releasing enormous volumes of gas, both from their own disintegration and from free gas trapped beneath them. The effect is analogous to a period of "earth flatulence". The release of so much gas has been blamed for destabilising submarine slopes around many continents. The tell tale bottom-simulating reflector indicating gas clathrate occurs widely on the East Coast (Fig. 10) and we suspect that the big failures north and east of East Cape may date from periods of falling sea-level when the clathrate became unstable. At present, the evidence is inconclusive.

How big is a tsunami wave generated by submarine avalanches, particularly very big ones? Frankly we don't really know. The evidence is largely geological and anecdotal. Computer models have been used to estimate the sizes of tsunamis generated by earthquakes, particularly those that involve rupture of faults at the seabed (Fig.11). If the seabed instantaneously drops by several metres, then it may be supposed that the sea surface suddenly drops by a similar height, at least on the continental shelf. This generates a wave that propagates outwards in all directions but is perhaps concentrated in certain directions by the alignment of the fault and seabed topography. Models are also used to predict what happens to tsunami waves when they reach the coast. There, the decreasing speed of wave propagation and topographic effects

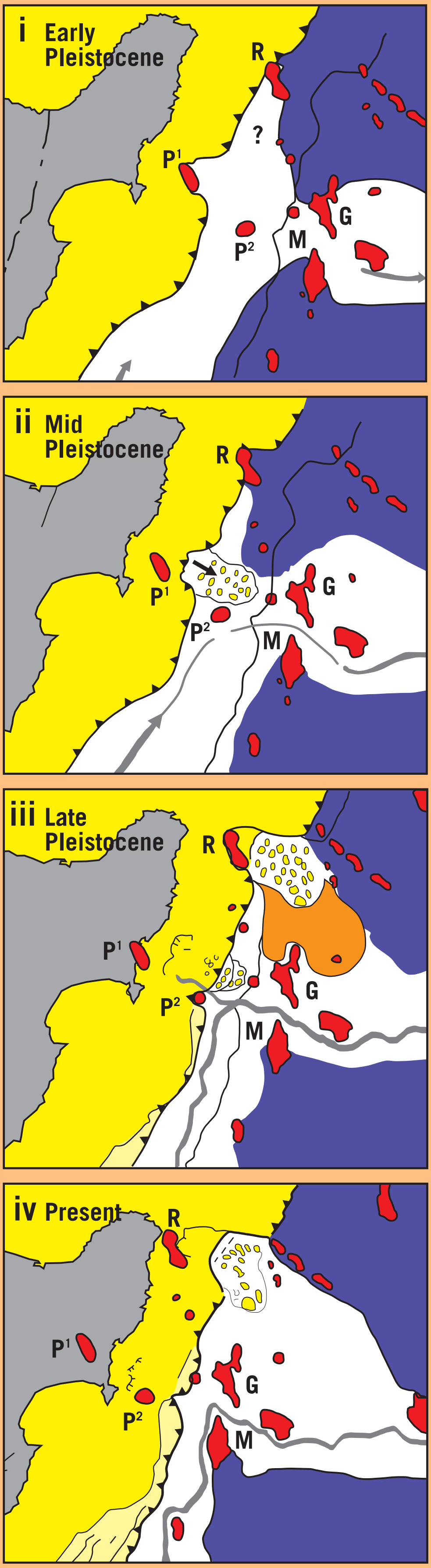


Figure 7. - Beginning at the top, over the last 1.5 - 2 million years, a succession of volcanic seamount (red) on the Pacific plate (blue, plus white where thick sediment cover) have impacted on the East Coast margin (yellow). Fracturing of the margin by the passage of these seamounts has triggered avalanches (yellow patches) and debris flows (fawn) in their wake. Seamounts and their avalanches and debris flows have dammed the trough to the passage of dense turbidity currents that overflow a meandering channel to form a wide plain(white) on the Pacific plate. Note the progression of the seamounts R (at Ruatoria) and P^1 (at Poverty Bay) under the margin, and the troughs and the margin collapses that develop in their wakes. The line between P^2 and M in figures i-iii, marks the Pacific plate yet to be subducted under the margin. Note the scatter of approaching seamounts on the Pacific plate to the east.

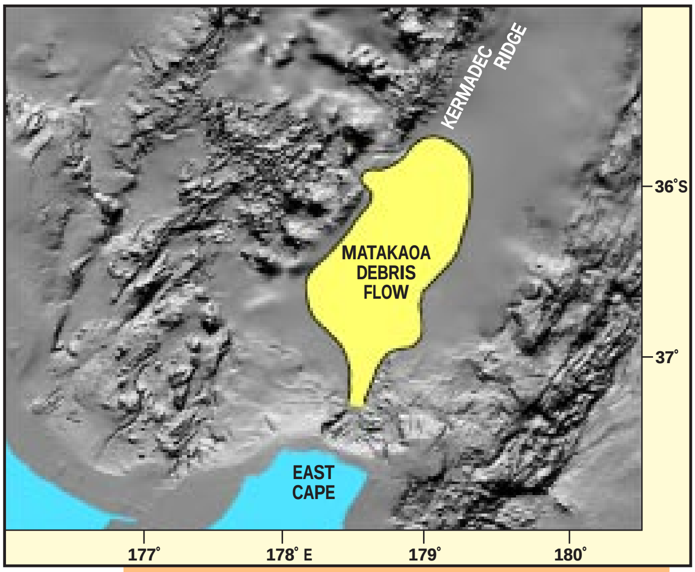
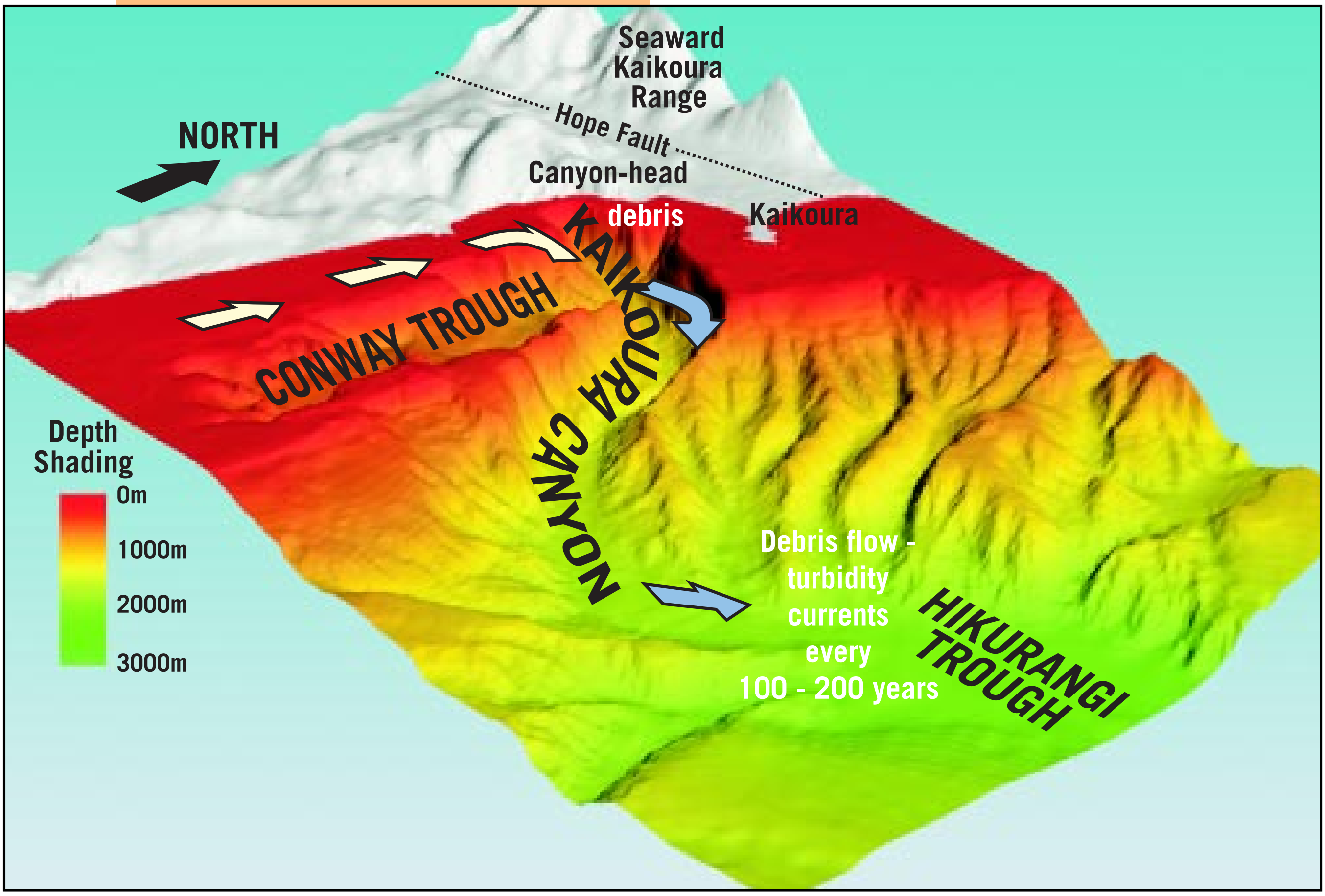


Figure 8 - Matakaoa margin indentation north of East Cape and Matakaoa debris flow on the flat plateau further north. These could not have been caused directly by seamount subduction but may be an indirect effect of plate boundary processes. (courtesy of Lionel Carter, NIWA).

Figure 9 -Sand, together with mud and some gravel, carried northwards along the shelf by waves and currents, pours into Kaikoura Canyon-head where it accumulates for perhaps 100 - 200 years before being shaken loose by large earthquakes associated with the rupture of nearby faults. The collapsing mass of sediment absorbs water, dumps its gravel load in the lower canyon, and continues as a high velocity, turbidity current for many hundreds of kilometres.

can cause run-up to many times the height of the tsunami wave in open water.

Modelling the nature of tsunamis generated by submarine avalanches is very much more complicated. Models must take into account many more factors than those for simple fault displacements. Certainly, a drop in the seabed can be expected to produce a corresponding drop in the sea-surface but the dynamics of water movement are complicated by a rapidly offshore-moving bulge in the seabed as the avalanches plunges downslope. The corresponding bulge in the sea-surface might be expected to be less obvious as the avalanche reaches deep water, where proportionally more water is displaced to the sides than vertically overhead. In addition, waves produced by the downslope moving bulge will interfere with those propagating out from the sea surface depression formed produced at the initial failure. The problem is complex enough for Kaikoura-sized avalanches, but even more difficult for the collapse of 3 600km² debris flow, of seabed such as occurred off Ruatoria. At present, all we can do is guess about many of the unknowns and, therefore, about the scale of landslide generated tsunamis, based mainly on analogy with Hawaiian and North Sea reports. We hope that, over the next



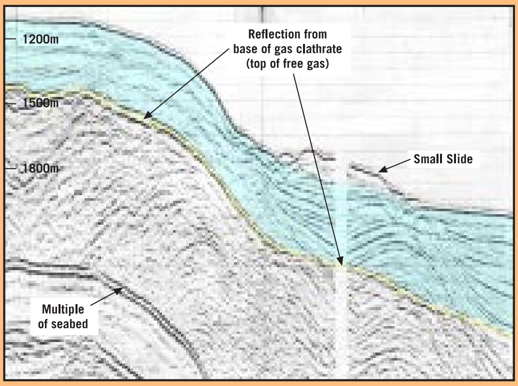


Figure 10 - A strong reflector that remains 400-500m beneath the seabed and cuts across geological strata marks free gas trapped beneath rocks whose pore spaces are filled with ice-like methane clathrates. The clathrate becomes unstable when the height of the overlying water column, and hence the pressure, is reduced by falling sea-level at the start of ice ages. This can release enormous quantities of gas into slope sediments making them highly unstable and susceptible to catastrophic avalanching.

few years, realistic models can be developed both for the generation of tsunamis from large submarine avalanches, and for their effects on New Zealand's vulnerable shores and harbours.

In summary:

From our new but still limited knowledge, we suspect that potentially catastrophic tsunamis can be generated by submarine avalanches of thousands of cubic kilometres of rock, on and near the active margins of New Zealand.

Such huge avalanches occur so infrequently, on a scale of hundreds of thousands of years, that the risk can probably be regarded as negligible in a human time-scale.

The risk of large "far-field" tsunamis, from big avalanches anywhere around the Pacific, is still largely unknown as many margins have not yet been mapped in appropriate detail.

Smaller avalanches may that may generate tsunamis many metres high (we have not yet done any calculations) might occur during earthquake-generated collapse of soft sediment in canyon heads. Those at Kaikoura might be expected perhaps every century or two and the last one was 170 years ago. We have no estimates for other canyon heads around New Zealand.

In the near future, we plan to further assess the nature of offshore landslides around New Zealand and to develop feasible models of tsunami generation and propagation to nearby and distant coasts. We are still some way from realistic assessments of risk at vulnerable shores, ports and coastal towns but work is progressing.

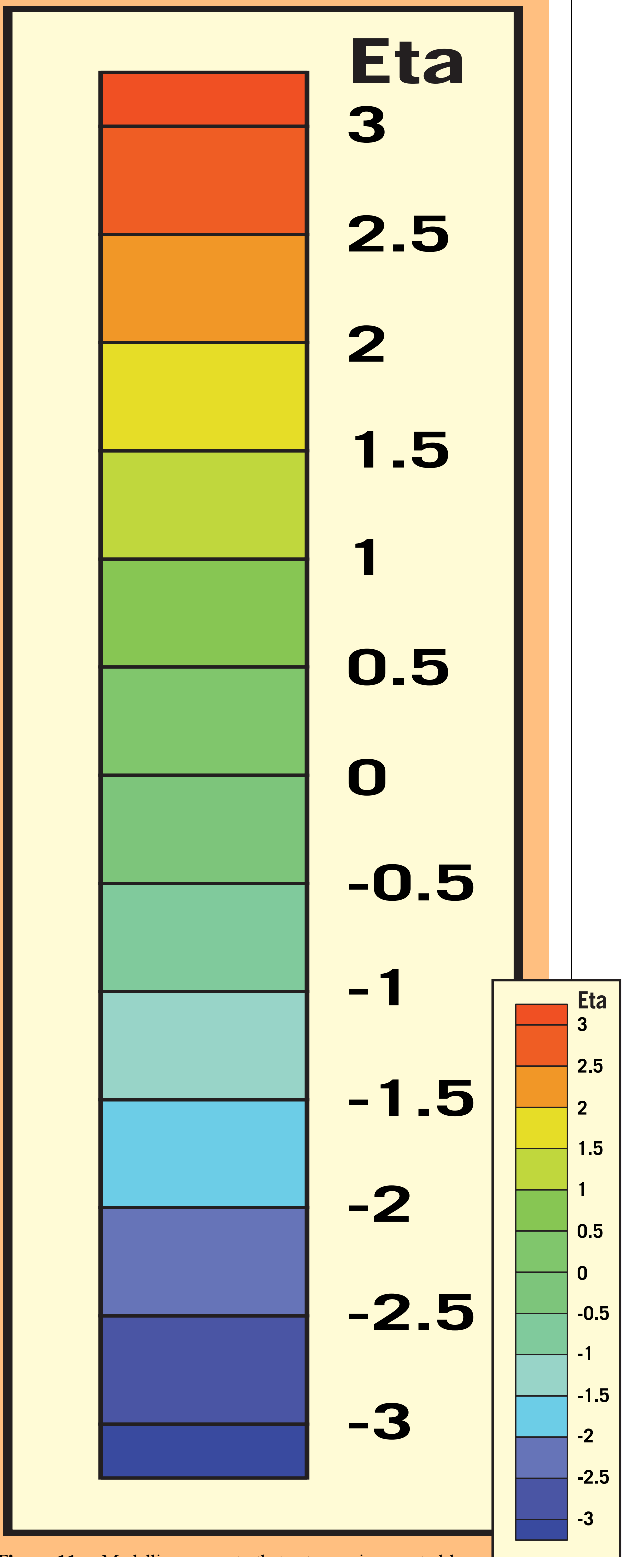


Figure 11. - Modelling suggests that a tsunami generated by displacement of an active fault on the outer shelf off Hawkes Bay will surge around the bay causing significant increases in height at some places and little effect in others. Immediately after the fault moved, a 3m high wave hits Wairoa. At 100 minutes after the fault moved, a 2.5 m high wave reaches sites in southern Hawkes bay and the sea drops by 3 m at Wairoa. After that the wave reverberates around the bay for hours. At present, we do not have models to accurately predict more complex effects of different types of slope failure.

Volcanoes and tsunami hazard – implications for New Zealand

by Willem de Lange and Gegar Prasetya
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Tsunami generated by volcanic eruptions are much less frequent than tsunami produced by submarine earthquakes, but they account for a large proportion of the deaths caused by tsunami. A review of volcanic tsunami with particular reference to the Krakatau event 1883 identified 10 main mechanisms that may account for tsunami generation (Table 1). Four of these - caldera collapse, debris avalanches, submarine explosions and pyroclastic flows- have been suggested as the mechanisms that produce the largest tsunami.

Table 1. Tsunamigenic processes associated with volcanism and the characteristics of the tsunami produced. The characteristics indicate the range of values reported in the literature for historical events. Avalanches are considered to include landslides and other mass movement phenomena. This table does not include the largest tsunami generated by the Krakatau 1883 eruption as the source mechanisms(s) are still subject to debate.

	Tsunami characteristics			
Mechanism	Source volume (km^3)	Wave height (m)	Period (min)	Travel distance (km)
Volcanic earthquakes	1-10	up to 17	10-40	< 500
Submarine explosions	< 1	1-6	1-10	< 50
Pyroclastic flows (nuées ardentes)	1-100	up to 25	1-40	< 250
Caldera collapse and subsidence	1-10	up to 15	short	< 50
Avalanches of cold rock	< 1	1-10	short	< 50
Basal surges and shock waves	< 1	up to 5	aperiodic	< 10
Avalanches of hot rock	< 1	small	short	< 10
Lahars	< 1	small	short	< 50
Atmospheric phase coupling	?	small	15-40	> 1000
Lava entering the sea	< 1	very small	short	< 10

Physical (scale) and numerical modelling approaches provided the methodology to understand the physical processes of volcanogenic tsunami. The results can be used to develop predictive tools for volcanogenic tsunami hazard assessment.

Previous studies have suggested several alternative mechanisms to account for the largest tsunami formed during the 1883 Krakatau eruption (caldera collapse, debris avalanches, submarine explosions and pyroclastic flows) Physical and numerical modelling have provided useful insights into the actual mechanisms and the characteristics of the tsunami waves produced. The findings support the hypothesis developed from geological evidence that the largest Krakatau tsunami were generated by pyroclastic flows.

The findings on tsunami generation processes made by this study are as follows:

1. Submarine explosion mechanism

A single explosion cannot produce a high wave. As the size of the explosion increased, the efficiency decreased markedly once the radius of the explosion bubble exceeded the water depth. The maximum wave height produced was less than that observed at Krakatau in 1883. The efficiency of the submarine explosion mechanism is increased by using a sequence of smaller explosions, instead of one large explosion. However the timing between explosions is critical; if the explosion are too close together or too far apart, the efficiency decreases. Based on the numerical modelling it is considered that the optimal timing will vary with water depth and explosive yield.

2. Pyroclastic flow

Waves can easily be generated by gravity flows entering the water, regardless of the slope. The wave properties depend on the relative densities of the flow and the receiving body, and the velocity of the flow. The angle of entry of the flow into the water determines the deposition pattern of sediment: at low angles the nexus of deposition is at the base of the slope; and this moves further away from the shore as the slope angle increases. This behaviour accounts for both the moat evident around the base of the former Krakatau volcano, and the shallow Steers and Calmeyer shoals. In the physical tests, the less dense material from the pyroclastic flow propagates near the water surface for long distance. This is consistent with evidence from Krakatau. Although the tests were not performed, it is probable that multiple smaller pyroclastic flows will produce larger waves than a single large one. Numerical modelling of single pyroclastic flows demonstrates that the resulting tsunami is highly directional, with the largest wave heights along the axis of the flow. Physical modelling of column collapse using a free-falling circular flat plate produced a radially uniform 'flow'. The distribution of ignimbrite produced by Krakatau in 1883 is fairly uniformly distributed around the source, although there is some indication of thicker flows being directed towards the northeast and west. This suggests multiple pyroclastic flows flowing radially from the eruption column, with possibly large late phase flows (occuring just before caldera collapse) being preferentially directed towards the northeast and west. These late flows were responsible for forming the largest tsunami waves.

3. Caldera collapse

The efficiency of this mechanism to generate a wave is dependent on the 'collapsing time' and the volume of the caldera. A very sudden collapse with appropriate caldera dimensions could produce significant waves. This mechanism starts with the dropping of sea level around the area, producing an initial negative wave displacement.

From these findings, if a super violent explosion did occur during the 1883 Krakatau eruption as suggested by reports in the historical record, then the water waves (tsunami) that caused the devastating effect on the surrounding island were not caused by the direct transfer of explosive forces. Instead a sequence of one or more pyroclastic flows, resulting from a collapsing eruption column, in and around the Krakatau complex are the most likely mechanism causing the largest tsunami. The formation of the Steers and Calmeyer shoals by the 1883 Krakatau event was reproduced by the pyroclastic flow experiments using coarse sand and mud with steep entry angle (> 60°). This simulation also left the moat evident in the present bathymetry.

The numerical modelling of volcanogenic tsunami from the Auckland Volcanic Field (Table 4) showed that volcanic tsunami are not a major threat to Auckland. However under suitable conditions a volcanic eruption could produce moderately large tsunami that generate strong currents (Figures 6 and 7). The maximum impact occurs along shores close to the source of the tsunami. The modelling presented in the preceding chapters represent the worst possible scenarios. Therefore it is unlikely that any coastal region will experience wave heights much larger than 1 m.

Of interest was the consistent development of trapped waves (edge waves) along the North Shore beaches, and the seiching within Tamaki Estuary. Tamaki Estuary has shown an amplified response to historic teletsunami, so this may be a fundamental behaviour of the estuary.

Implications for Civil Defence

The numerical modelling of volcanogenic tsunami for the Auckland Volcanic Field has some implications for the local Civil Defence:

Phreatomagmatic eruptions are the most common eruptive style. These typically form maars by a series of explosions as evidenced by the layering in the surrounding tuff rings. The modelling shows that a series of explosions is the most efficient explosion mechanism. If the timing between eruptions is optimal then quite large tsunami can be generated (up to 20 m at source) that can have a significant impact on nearby shores.

Modelling for pyroclastic flows from Browns Island shows that an eruption in shallow water will have a bigger impact than a deeper water site, such as Rangitoto Channel.

The tsunami produced will reach the shore within minutes of the eruption, but can persist for up to an hour. The tsunami travel further along the deeper water open coasts, than within the shallow Waitemata Harbour.

Tamaki Estuary was affected by all the sources considered. The Estuary seiched readily, which amplified the tsunami waves at antinodal points.

Therefore the Tamaki Estuary is likely to show a significant response to any tsunami.

Overall the hazard associated with volcanogenic tsunami in Auckland is low.

4. Pyroclastic Flow Tsunamigenesis

Pyroclastic flows are hot, variably fluidised, gas-rich, high particle concentration mass flows containing pyroclastic debris. The deposit produced by such flows is an ignimbrite. Pyroclastic flows are most commonly produced by the collapse of an eruption column, but they may also result from the collapse of volcanic domes as demonstrated at Mount St Helens in 1980 .

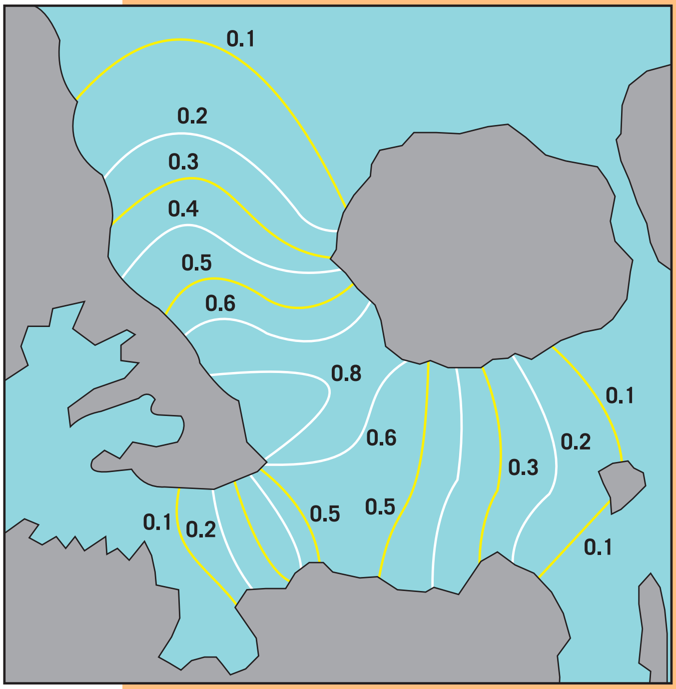
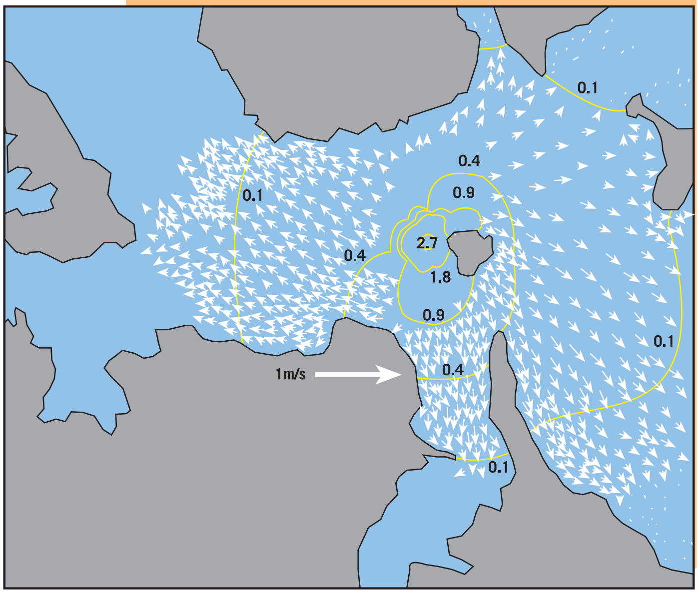


Figure 6 - The patterns of wave propagation 5 minutes after a pair of phreatomagmatic explosions in the Rangitoto Channel. The explosions were 2 minutes apart.

Figure 7 - Wave and current conditions 5 minutes after a pyroclastic flow from Motukorea (Browns) Island. The waves entering the Tamaki Estuary set up a seiche that amplified the waves near State Highway One and the Otahuhu power station.



A small eruption column tends to produce single flows, but as the magma discharge rate increases it is possible to produce multiple flows simultaneously. High discharge rates are also associated with high flow velocities.

There is a growing body of evidence that pyroclastic flows can enter water and produce water supported mass-flows that can generate tsunami. A review of the documented examples of subaqueous pyroclastic flows and ignimbrite deposits identified a range of different types of interactions between flows and water bodies. The important parameters controlling the interactions appear to be the bulk density of the flow, the velocity and discharge rate of the flow, and the angle of incidence between the flow and the water surface. Table 2 summarises the effect of these parameters.

The 1883 Krakatau eruption involved low bulk density flows with high velocities. The resultant ignimbrite shows clear evidence of high temperature emplacement. This suggests that the deposit was mainly the product of a low density flow with a high angle of incidence, i.e. column collapse directly into the sea. Physical modelling of pyroclastic flows indicates that these conditions also replicate the moat observed at Krakatau, and produce a hummocky ignimbrite with large mounds at the distal ends of the ignimbrite, that are consistent with the temporary islands formed by the 1883 eruption.

Table 2 indicates that a large number of different interactions may be possible when a pyroclastic flow interacts with a water body. However, this can be simplified if the main result of interest is the formation of a tsunami. There appear to be four main mechanisms by which a sufficient volume of water could be displaced to form a tsunami:

1) Deposition at the shoreline causing a lateral displacement as the zone of deposition moves offshore.

2) Upward and lateral displacement of water caused by the propagation of a water supported mass-flow.

3) Downward and lateral displacement of water caused by the sinking of debris from a segregated flow travelling over the water surface.

4) Upward displacement of a large volume

of water due to the deposition of a caldera-infill ignimbrite or pyroclastic flow deposit.

The final mechanism is only important if sufficient time is available between the formation of the caldera and the eruption to permit the caldera to be flooded by sea water. This is usually not the case. A common feature of the remaining three mechanisms that distinguishes them from most earthquake tsunami generation mechanisms, is the lateral displacement of water. This should be simulated in any model used to consider tsunami generation by pyroclastic flows. However, numerical models in particular tend to model the result of pyroclastic flows as a single upward or downward displacement applied as an initial deformation of the water surface.

One way to include both lateral and vertical displacements in a numerical model is to treat the pyroclastic flow as a horizontal piston represented by a sequence of wedge-shaped displacements of the seabed. At each time step, the sea bed moves up or down as necessary to replicate the passage of the flow (Figure 1). The movement is defined by three parameters: the time when movement starts; the vertical velocity; and the time when movement stops. The resulting sea bed deformation is then assumed to be transmitted to the sea surface.

The model was implemented by modifying an existing finite element numerical model which generated a tsunami by defining an initial sea bed displacement and assuming a matching sea surface deformation. In the original model the displacement occurs instantaneously at time zero. The model was changed to allow the specification of the additional parameters discussed above. It is still assumed that the sea bed displacements are transmitted directly to the sea surface. This assumption is not strictly valid. However the physics of pyroclastic flow interaction with seawater are not well defined, and the model does well at replicating the features observed in the two-dimensional physical model tests.

This methodology was intended to simulate flows that cause an initial upward displacement (mechanisms 1, 2 and 3 above) as initial upward displacements tend to produce larger waves than initial downward displacements. This method produces a dynamic water surface displacement that contrasts with the static water surface displacement commonly used as an initial condition.

5. Potential Bay of Plenty Volcanic Tsunami Sources

Within the Bay of Plenty region (Figure 2), four main volcanic tsunami source regions can be

	Shallow submarine vent		Low angle of incidence		
Bulk density	Low discharge rate	High discharge rate	Low flow velocity	High flow velocity	High angle of incidence
Low	• hot, water supported mass-flow • upward and lateral displacement	• large upward displacement • pyroclastic pond deposits	• explosive disintegration or • deposition at shoreline • lateral displacement	• segregated flow over water • flow may sink and form warm, water supported mass-flow • downward and lateral displacement	• hot, water supported mass-flow • upward and lateral displacement
High	• cold, water supported mass-flow • upward and lateral displacement		• explosive disintegration or • deposition at shoreline • lateral displacement	• segregated flow over water • downward and lateral displacement	• hot, water supported mass-flow • upward and lateral displacement

Table 2. Summary of the results of the interactions between pyroclastic flows and water bodies. This table does not consider any interactions involving the post-eruption redistribution of pyroclastic debris.

	Mount St. Helens	Krakatau	Taupo
Volume (km^3)	1.0	10.0	100.0
Mean thickness (m)	5.0	20.0	80.0
Area (km^2)	200.0	500.0	1250.0
Semi-major axis (km)	10.3	16.3	25.8
Semi-minor axis (km)	6.2	9.8	15.5
Flow velocity ($m.s^{-1}$)	100.0	150.0	200.0
Flow duration (s)	206.0	217.2	257.5

Table 3. Summary of pyroclastic flows characteristics used to simulate volcanic tsunami in the Bay of Plenty. The flows were assumed to cover an elliptical region defined by semi-major and semi-minor axes given.

Rangitoto Channel	
Vent location:	Offshore in Rangitoto Channel
Underlying country rock:	Seafloor sediments over Waitemata Group Flysch (Sandstones and mudstones)
Magma type:	Basalt
Water depth:	< 20 m
Erupted volumes:	Tehpra - 107 m3; lava -108 m3
Generation mechanism:	Submarine explosion, maar formation and pyroclastic flow.
Tamaki Estuary	
Vent location:	Tamaki Estuary
Magma Type:	Basalt
Underlying country rock:	Micaceous sand of tidal mudflats, over Waitemata Group Flysch (sandstones and mudstones)
Erupted Volume:	Tephra fall - 107 m3, base surge deposits - 5x107 m3
Water depth:	< 5m
Generation mechanism:	Submarine explosion and maar formation
Motukorea	
Vent location:	Motukorea Island
Magma type:	Basalt
Underlying country rock:	Waitemata Group Flysch (sandstone and mudstone)
Generation mechanism:	Pyroclastic flow
Rangitoto Volcano	
Vent location:	Rangitoto Island
Magma Type:	Basalt
Underlying country rock:	Waitemata Group Flysch (sandstone and mudstone)
Generation mechanism:	Pyroclastic flow

Table 4 Scenarios used to generate volcanic tsunami in the Auckland region.

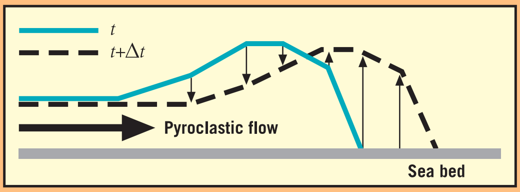
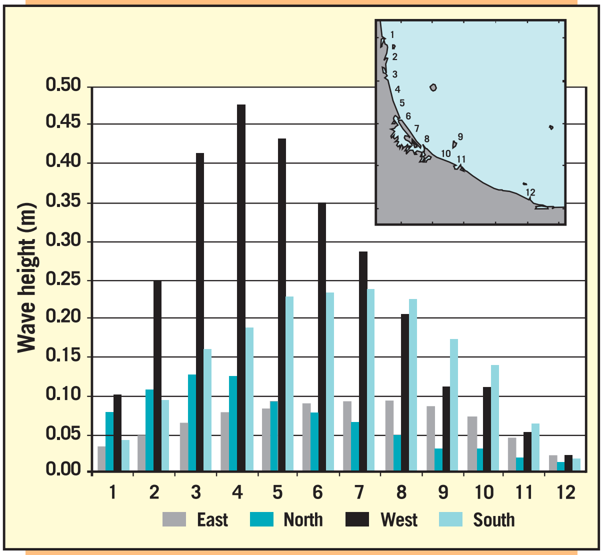
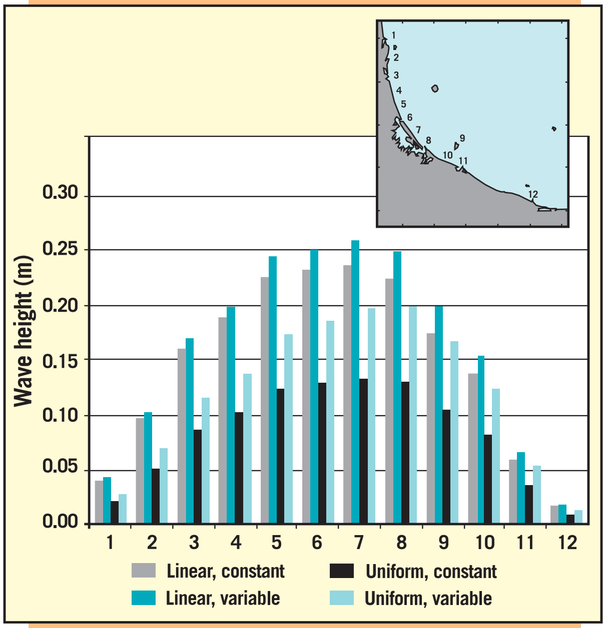
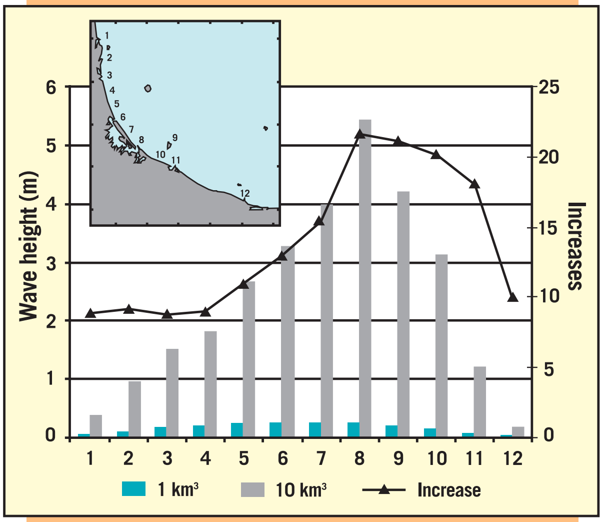


Figure 1 - Schematic diagram of the vertical displacements at the sea bed associated with the propagation of a subaqueous pyroclastic flow. The combination of vertical rise, followed by a relaxation causes







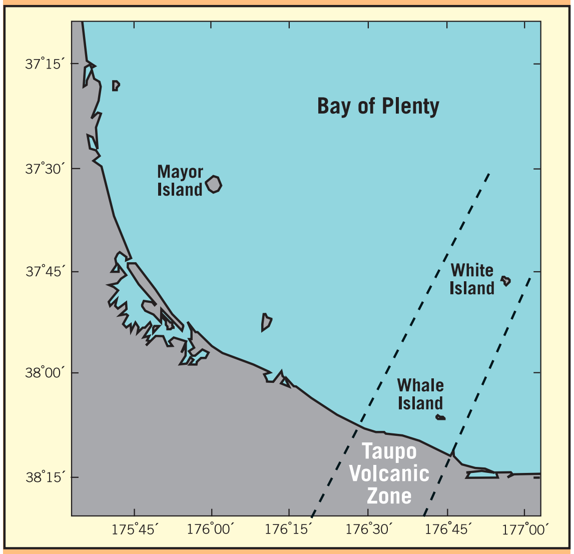


Figure 2 - Potential volcanic tsunami sources in the Bay of Plenty, New Zealand.

Figure 3 - Tsunami maximum wave height distribution around the Bay of Plenty coast resulting from a Mount St Helens scale pyroclastic flow travelling in different directions from Mayor Island. A uniform thickness distribution and constant vertical velocity have been applied.

Figure 4 - Tsunami maximum wave height distribution around the Bay of Plenty coast resulting from a Mount St Helens scale pyroclastic flow travelling south from Mayor Island. Various combinations of deposit thickness and vertical velocity have been applied.

Figure 5 - Comparison of the maximum wave heights produced around the Bay of Plenty coast by a Mount St Helens (1 km3) and Krakatau (10 km3) scale pyroclastic flow travelling south from Mayor Island. A uniform thickness distribution and constant vertical velocity have been applied. The increase is the ratio of the Krakatau height distribution to the Mount St Helens.

identified:

1) White Island is an andesitic volcano that is currently active and represents the northern-most extension of the Taupo Volcanic Zone. The main crater of the island is breached to the south-south-east, and the crater wall is only a slight distance above sea level, with parts of the crater floor lying below sea level. The volcano has produced small pyroclastic flows and debris avalanches in the past, and could potentially generate a tsunami.

2) A zone of hydrothermal vents extends south-south-west towards the shore between White Island and Whale Island. A slight risk of phreatic activity is associated with this zone.

3) Mayor Island is a peralkaline volcano with a previous history of pyroclastic activity and caldera collapse. Although the volcano is not currently active, the most recent activity occurred within the last 2000 years.

4) The Taupo Volcanic Zone is the main

zone of volcanic activity in the North Island of New Zealand and strictly includes White Island. However, the activity at the offshore northern end is predominantly andesitic, whereas the inland section in the Bay of Plenty is rhyolitic in character. The Okataina Volcanic Centre is the largest volcanic complex close to the coast and within the last 400,000 years has experienced 5 or 6 large eruptions involving ~500 km^3 of material

At least one of these eruptions produced pyroclastic flows that entered the sea and left ignimbrite deposits 50-100 m thick.

6. Mayor Island tsunami

Numerical modelling indicated that the main volcanic tsunami risk involves pyroclastic flows from Mayor Island. The modelling considered single pyroclastic flows with volumes of 1 km^3, 10 km^3 and 100 km^3 (Table 3). This assumption may be unrealistic for the larger volume events that would involve multiple flows. However, the available evidence indicates that the flows would be sufficiently close together that the displacement of water should be similar to that produced by a single flow.

Besides the variation in volume displaced, the modelling examined the effect of varying the following parameters:

1)The direction of the flow. Single pyroclastic flows are highly directional which should affect the distribution of energy in any tsunami generated. With larger volumes, multiple flows may reduce the significance of direction if there is no preferred orientation of the flows.
The distribution of the 1883 Krakatau ignimbrite deposit suggests this may be the case for the main tsunami event.

2)The thickness of the flow. The flow was treated as being either of uniform thickness, or having a maximum thickness along the semi-major axis with the thickness decreasing linearly to the margins of the flow. The thickness of real ignimbrites tends to be controlled by the flow characteristics and the pre-existing topography. This may be too complex to predict.

3)The methodology used to simulate generation of a tsunami by a pyroclastic flow is controlled by the velocities of the vertical displacements (Figure 1). Two main approaches were used to determine these velocities:

a)A constant uplift velocity was applied once the flow reaches any given location, and it continues until the required thickness is achieved.

b)A variable uplift velocity is applied so that the flow reaches the final thickness simultaneously across the whole region affected. This produces the highest velocities at the distal margins of the flow, and the lowest velocities close to the vent.

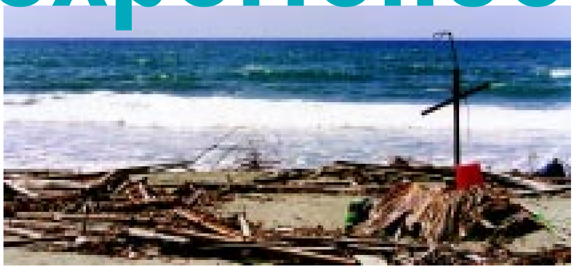
Figure 3 shows the effect of pyroclastic flow direction on the coastal maximum tsunami wave height distribution. The directionality of the tsunami is evident, although part of the variation can also be explained by changes in travel distance, and hence dissipation.

The depth distribution has a significant effect on the wave height distribution, particularly when a variable vertical velocity is applied (Figure 4). This occurs due to the larger displacements at the margins of the flow with a uniform depth distribution, and the correspondingly faster uplift velocities produced by the variable option. The vertical velocity method used has a larger effect than the depth distribution, with the largest waves being produced by the variable option.

Numerical modelling of tsunami generation by earthquakes (initial static water surface displacement) usually produces a linear relationship between the volume of water displaced and the wave height distribution. This is not the case for the pyroclastic flow model (dynamic water surface displacement) applied to the Bay of Plenty (Figure 5). Increasing the displacement volume from 1 km^3 to 10 km^3 increases the wave heights by a factor of 10 for an initial static water surface displacement.
However the same volume increase with a dynamic water surface displacement produces larger increases (Figure 5). This effect is most pronounced along the semi-major axis of the flow. If large volume eruptions are treated as multiple flows with a radial distribution around the vent, the directional effect is reduced but the scale factors are still larger than predicted by displacement volume ratios.

Clearly the correct definition of vertical velocity is necessary to produce reliable predictions of tsunami characteristics. This could not be done for the Bay of Plenty due to the lack of data for calibration. Therefore the model is being applied to the Krakatau 1883 eruption to develop a suitable strategy for defining the parameters controlling the dynamic displacement. Physical pyroclastic flow models have also been used to provide calibration data. The physical models indicate that a dynamic displacement model is required to simulate the interaction of pyroclastic flows and water bodies.

Tsunami - the experience



by Mauri McSaveney

Institute of Geological & Nuclear Sciences Ltd

In the company of friends, we are enjoying a quiet moment at the end of a busy day. It is Friday evening. We are on the beach chatting after watching the last rays of the setting sun. Further along the beach, our children, happy to be home for a four-day holiday, are playing touch rugby. The earth shakes - another earthquake. In fact, quite another earthquake! It has been a while since we experienced such shaking. Someone remarks that the tide is going way out. We argue about when it last was low tide. It must be some mistake, it should not be low tide now. A distant murmur grows to a rumble, then a roar, like approaching jet planes, or is it a fleet of helicopters? Then the tide is coming in with a vengeance. A huge wall of water the colour of Milo, glowing red near the crest, is crashing in from the sea, then another, and another, and we are fighting for survival as the sea sweeps us inland. All is a turmoil of swirling water, sand, trees, friends and our disintegrating homes.

As the swirling water returns to the sea, some are pulled with it, a few still fighting desperately for survival amongst the floating debris that was once a thriving community. Those of us who are fit young men are likely to survive the experience. The sun has set. It is dark. Our clothes have been torn from us. We feel as if all our skin has been sandblasted off. We have taken a severe battering in the surf and may have several broken bones. It will not be light for another 11 hours. We shout for our family, but mostly it is other men that answer back. Calling frequently, those of us able to move find our way towards each other in the dark. Screams of injured people become fewer and fewer through the night.

Another scene: It is morning. We are expecting an early call from a distant friend. The call hasn't come, and we can't get through when we try to raise them. Our Kiwi pilot friend from Palmerston North is scheduled to fly over the area later in the morning, so we ask him to see what's up. As his plane reaches the area, he can hardly believe his eyes - an unrecognisable scene of devastation - all the houses are gone. So little of the community structure has survived the night of terror that no one has been able to raise an alarm.

This is what a tsunami disaster can be like. The good people of Arop and Warapu on the northern coast of Papua New Guinea experienced this first hand on the night of Friday the 17th of July, 1998. Along 19 kilometres of coastline, a wall of water over ten metres high (locally up to 15 metres) swept over the beach and into a tropical island paradise, destroying every building in its path. The true death toll is impossible to determine, but it is known that more than 2189 people died. In Arop and Warapu the death rate may have exceeded 70%. Within 24 hours of the outside world learning of the tragedy international medical teams of trauma specialists were arriving in the area and aid was beginning to pour in. But for the first twelve hours these people were on their own, almost completely

Above - Putting new meaning to finding Paradise on a tropical island beach: - Some of the PNG tsunami's victims, sucked out into the ocean by the powerful backflow out of Sissano Lagoon, were buried in shallow graves along the beach.

stripped of the means to help themselves, their homes and community infrastructure totally destroyed.

Tsunami do not just affect those exposed to their brute force: there is the anguish of grieving relatives from neighbouring communities as over the following days they wander through an unfamiliar scene of carnage and destruction, finding and burying body parts, and hoping not to recognise friends and relatives among them. The following extract conveys how the PNG tsunami affected some:

It was Thursday 16th July 1998 and the school term break had just commenced for nine year old Florentyna who said to her mother, “Mummy, I want to go home to the village with grandma and grandpa” Upon hearing this, the younger sister, Sharon, said, “I want to go with them too, please mummy.” Florentyna Numara was the oldest daughter of Canisius and Gertrude Numara .

Gertrude recalls: “On the day of the disaster, 17 July 1998, when I felt the earthquake, I grabbed my two boys, ages four and one and ran outside of the house. Then, like a lot of people, I heard a sound like a jet aircraft taking off coming from the northwest direction of the sea.

With the two boys in my hand, I ran out onto the road and looked out to the sea to see what this noise was all about. There I saw a great big wave already crashed and as white as anything rolling towards the shore. In the next few seconds, it crashed on to the Aitape wharf. It was both a spectacular and frightening sight.

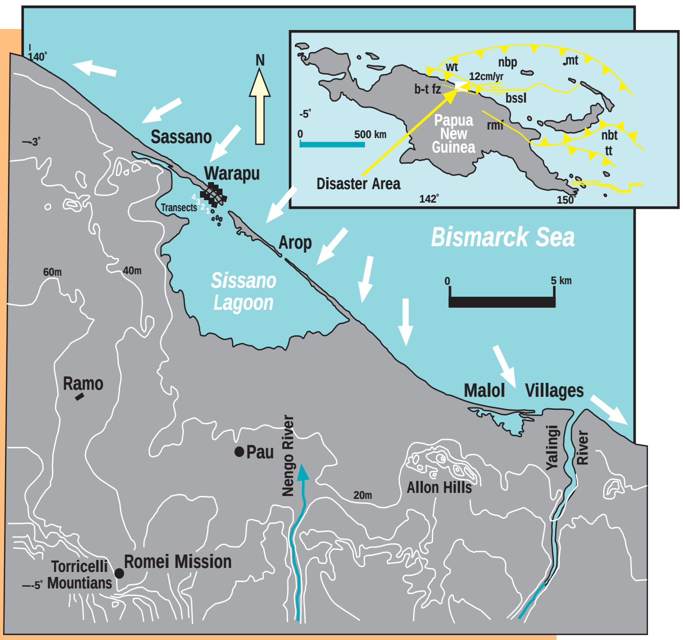
Still shaking, I went back to the house with the two boys and as soon as I sat down with Canisius the first words which came to my mouth were: “The two girls”...

I fed the boys and put them to sleep. It was 8:30 pm. I had lost both my appetite and sleep and sat up throughout the night ’til the following morning. At the crack of dawn, I went down to the beach and saw the damage caused by the wave to the coastline of Aitape. I did not like it. I had this sudden urge to get information, any information, on the possible damage by the wave to the villages on the west coast of Aitape.

At 7.30 am on the day after, the urge for me to get information grew more intense. The first person I met who I thought might have some information on the west-coast villages was Br Paias Teke, OFM. All he could tell was, “Radio long Sissano i no kamap” (the radio at Sissano is not on air). I could not rest and went all over town until I ran into another person who had first hand information from the MAF Pilot who had just flown into Aitape from Vanimo. True to my fear and shock, I was told that my once beautiful village was completely wiped out. That news only increased my anxiety and fear about the safety of my two girls, my parents, my relatives and the Barupu people generally.

I rushed to my house, grabbed hold of a few essential items and told Canisius that I was going to the village to find out for myself about the safety of my daughters. He insisted that I do not do so and that he alone should go. This he did....

Canisius recalls that as soon as he arrived into Warapu by boat, and having seen Arop on the way to Warapu, he was so shocked, horrified, bitter, dumbfounded and incredulous. A quick search of the completely devastated village and the accessible lagoon resulted in no sight of either of the two girls or their bodies. He did not know what to believe! Someone suggested that he go to the other side of the lagoon at Aroporo where the survivors were... This he did and found my father who told him that Florentyna was not with them - only Sharon. Apparently Florentyna had run away from the house during the earthquake and was still separated from



Sissano Lagoon lies on the northern coast of Papua New Guinea. Arrows pointing to the coast indicate the tsunami direction as it hit the coastline. The area is geologically very active because it lies near a number of fast-moving crustal plates around the Bismarck Sea (mt - Manus Trench; nbp - North Bismarck Plate; wt - Wewak Trench; bssl - Bismarck Sea Seismic Lineation; b-t fz - Bewani-Torricelli Fault Zone; rmf - Ramu-Markham Fault; nbt - New Britain Trench; tt - Trobriand Trench).

my father and mother during the subsequent onslaught by the tsunami.

Canisius arrived back into Aitape that same evening and brought me the sad tidings. I had never been as distraught as when Canisius brought home to me the news. There have been many troughs in my life - but nothing will equal this as long as I live. "Why did she have to go home?" I lamented vainly, but still keeping some hope that she was swept ashore into the mangrove swamps and would still be found alive.

Canisius left again on Sunday in another attempt to find her. Upon his arrival at the village there were quite a number of survivors, government workers and Church workers moving about, either burying the dead, rescuing survivors or survivors picking up what little pieces were left of their personal effects. Something told him that he should look along the beach before he ventured into the lagoon. To his horror and shock, there on the beach, was "the prettiest girl I have ever seen in my whole life, Florentyna, with only her t-shirt and pants on, lying as peacefully as can be with her face towards the sea, as if she was admiring the now calm and benign sea and ocean."

After he got over his shock and disbelief, Florentyna's father removed his shirt and, as respectfully as he could, wrapped it around her body which was by now beginning to decompose and, like everyone else, he dug a grave in the sand with his hand and bade her farewell.

Gertrude Numara's recollections, in the Lutheran Diocese of Aitape Rehabilitation Committee Newsletter of February 1999.

Many of the coastal villages of northern Papua New Guinea operate on a near-subsistence, 20th -century economy: no paved roads, no mains power, no sewer and no high rise buildings. The people have little need of these things but they do have aluminium runabouts with outboard motors, radios, cassette decks, photo albums, bicycles, sewing machines, video cameras, and high-school diplomas - much the sorts of things one expects to find in a remote New Zealand coastal community.

What might we expect were a similar-sized tsunami to hit a New Zealand city? In 1960, a tsunami up to 10 metres high hit the town of Hilo on the south coast of the island of Hawaii, killing 61 people. The totality of the destruction bears a striking similarity to the experience in Papua New Guinea. The following story describes the aftermath:

At dawn my Grandpa, Dad and Uncle Harold went by car to a spot about where Burger King is today. You could not drive any closer to the Bay because there was a 20-foot pile of rubbish all along Hilo Bay. They climbed up the pile and looked over and saw only a vast open space where formerly the houses and businesses of downtown Hilo had been....

They walked to the location of the family business and all that was left was a flat cement slab that looked freshly poured. The wave had sandblasted off every speck of oil and grease. There was not any sign of the buildings or their contents. Everything had been destroyed. They found the store's 2,000-pound safe on the bayfront by where the Ironworks is today. Uncle Harold helped my Great Grandpa clean out the safe. Dad says Uncle Harold remembers his job was to iron the money to dry and save it...

from "My Dad and the 1960 Tsunami: the story of Tom Goya" as interviewed by Isaac Goya, an essay submitted to the Pacific Tsunami Museum 1998 essay contest.

Around the Pacific, hundreds of people were killed and thousands of structures were washed away by the same tsunami but it passed New Zealand at low tide and did little damage.

The recent Papua New Guinea tragedy was a very local disaster caused by a local tsunami triggered by a magnitude (Mw) 7.1 earthquake just offshore from the disaster area. The earthquake certainly was big enough to register on the world seismograph network but the network is not well placed to locate earthquakes around this area of northern New Guinea. Even today, when all the records are in, there is considerable uncertainty as to precisely where the first earthquake occurred. Before this event, scientists would have considered an earthquake of Mw 7.1 too small to cause a large, damaging tsunami. Since the disaster, the scientific community has been seeking an explanation for why the tsunami was so large. Some think that there was a large vertical movement of a fault just offshore, others favour the occurrence of a very large landslide triggered by the earthquake.
An earthquake-triggered escape of a natural gas reservoir from beneath the sea floor has been suggested because one witness who had watched the sun set 12 minutes earlier reported seeing the sea rise above the horizon and then spray vertically perhaps 30 metres just after the main shock. A group of us favour a more conventional explanation: movement of a gently sloping fault reaching to the sea floor in the deepest portion of the Bismarck Sea some 40 km out from the northern New Guinea coast. The seismic signals radiated by the fault

rupture do not help to differentiate the mechanism of faulting because they are consistent with movement on either a steeply dipping or gently dipping fault approximately parallel to the coastline.

When the initial medical emergency was over, teams of scientists from around the Pacific rushed to the area while the evidence of the tsunami was still fresh. Australian and Japanese seismologists joined with their counterparts in Papua New Guinea to deploy portable seismographs to better determine where aftershocks were occurring. A truly International Tsunami Response Team used a boat offshore as their base while they measured the runup heights of the tsunami and surveyed damage along the coast. Together with scientists from the University of Papua New Guinea they interviewed survivors. The New Zealand Society for Earthquake Engineering decided that there were lessons to be learned for New Zealand and sent a New Zealand team of which I was grateful to be selected as a member. We too linked with scientists from the University of Papua New Guinea, whose local knowledge and ability to tap into the PNG wantok system proved to be invaluable. Other research teams came after us, including a Japanese oceanographic research ship which undertook detailed measurement of the sea floor. Doubtless there will be more teams to follow because this tragic tsunami has been very instructive. A consensus on the final verdict has still to be reached. Our New Zealand team has presented one version: time will tell if we are right.

Within 50 kilometres offshore from the northern New Guinea coast lies the Wewak trench, over 5000 metres deep. It is where the tectonic plate which forms the ocean floor off northern New Guinea dives obliquely under the leading edge of

These fit young Papua New Guineans surfed through the coastal forest on a 1.5 metre high tsunami and lived to tell the tale; but one lost his wife and children.
(Photo: Peter Goldsmith)

This tsunami survivor indicates the height the water reached beneath his family home at Malol Village. Ninety five died at Malol, on the eastern edge of the disaster zone - mostly when their homes were swept away.
(Photo: Peter Goldsmith)

Even the strongest components of this child's bicycle were twisted and bent by the enormous power of the tsunami, which left debris in trees up to 17.5 metres above the sea in this area.

Immense trees were plucked from the sandy spit fronting Sissano Lagoon and transported more than a kilometre into the lagoon.

Every human structure was destroyed in the villages of Warapu and Arop on the sand spit fronting Sissano Lagoon. What debris was not caught in the trees, was washed into the lagoon.
(Photo: Peter Goldsmith)

the Australian tectonic plate at a very fast rate of 120 mm per year. It is one of the most active tectonic-plate margins on the planet. Interpreting the records from their array of portable seismographs, the Australian seismologists found a broadly dispersed pattern of aftershocks consistent with rupture of a very gently dipping fault somewhere just above the plate interface. This would suggest that the earthquake was similar to a type known as a subduction-zone earthquake (when one tectonic plate is drawn [subducted] under another), although this one probably was not directly on the plate interface. Large subduction-zone earthquakes are the leading cause of tsunami around the Pacific Ocean. But still, this earthquake was much smaller than those usually associated with tsunami.

Our New Zealand team found evidence that the earthquake had caused the area of Sissano Lagoon to sink a little - a very little - only some 30-50 centimetres, but enough to drown trees by submerging their bases along the inner margin of the spit fronting Sissano Lagoon. It was not the first earthquake to cause the area to subside. Nearby, we saw rotted stumps of trees drowned long ago in the lagoon. In 1907, the lagoon subsided an enormous 1.8 to 3.6 metres. We believe that July 1998 was the third time this century that the Sissano area has subsided in an earthquake but others still dispute our claims because they missed seeing the evidence for themselves.

The oceanographic survey ship found a very rugged sea floor out from Sissano Lagoon, including evidence which the international scientific crew interpreted to be a truly enormous submarine landslide, but the landslide was covered with a coat of post-landslide sediment and was cut by what appeared to be the trace of a fault. Clearly, this landslide could not be the cause of the latest tsunami.

The detailed offshore bathymetry was the last piece of evidence that our New Zealand team needed to confirm our ideas on what caused the tsunami. Briefly summarised, we believe that the earthquake was caused by a local rupture near the plate interface. The area of rupture extended from deep beneath Sissano, rising gently away from the coast almost to the base of the trench in almost 5 kilometres depth of water. An area of some 60 000 hectares of the earth's crust above the shallowly dipping fault plane moved about 2.2 metres towards the trench. This caused the spit and local sea floor at Sissano Lagoon to subside a little, about 0.4 metre. Out in the deep water the sea floor rose a little, perhaps 0.6 metre. Because the sediment on the ocean floor in the deep trench is soft, the fault would not have ruptured as a plane but would have curved upwards as it neared the sea floor, so the 0.6 metre is likely to be an underestimate of the true maximum rise in the sea floor. The fault probably did not rupture all the way to the sea floor, but instead the soft sediments folded. The vertical movement of the sea floor would have tilted the ocean surface by a metre or so over a distance of 40 kilometres, with the water surface sloping towards the land. But this was not the only effect of the fault movement: although the coastline stayed put, the steeply sloping sea floor moved horizontally by as much as about 2 metres with the amount increasing away from the coast. It had the effect of driving a wedge under the ocean mass, adding another 0.6 metre or so of uplift to the water out in the deep trench. About 40 kilometres out from Sissano Lagoon, the ocean surface then was 2 metres higher than at the coast and this particular section of coast lay in a hollow on the sea surface.

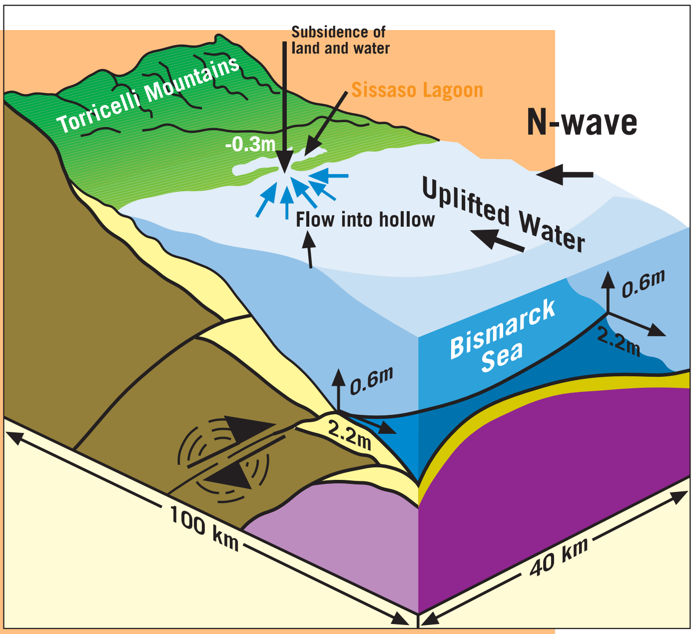
But probably what tipped the balance to create this extraordinary tsunami was that the trench margin out from Sissano Lagoon was not straight: it bulged out into the trench, so that the deformation of the sea floor did not produce a linear wave but a curved wave, concave towards the coast. As the mass of water rushed landward into rapidly shallowing water, it slowed and the enormous amount of energy contained in the deep column of water went into building the amplitude of the wave. Slowing more on the crest of the shallowing bulge, the wave progressively became more concave as it approached the shore. The concavity also had its effect: the wave was converging, focussing the wave energy in towards the centre of curvature. We can not identify which parts of the sea floor in the deep trench pushed up the curved wave but we can identify the 19 kilometres of coastline that the wave energy was focussed on.

The plate-tectonic setting of the northern coast of New Guinea is very similar to that of the east coast of North Island, although the relative plate motions there are three times faster than those in New Zealand. Could a similar disaster happen in New Zealand?

It not only could happen, it has happened and it will happen again. It will happen more than once but it will not happen very often. In the 1820s, several hundred Maori were killed at Orepuki when a local tsunami hit the Southland coast. In 1855, a tsunami at least 9.1 metres high washed the southern North Island coast of the Wairarapa. Other tsunami from the 1855 Wairarapa earthquake washed over the then-unoccupied isthmus where the Wellington suburbs of Kilbirnie, Lyall Bay and Rongotai now stand, and a tsunami washed into shops along Lambton Quay in Wellington despite the coastline having been lifted 1.2 metres in the earthquake. In 1868, a tsunami originating from a large earthquake in Chile destroyed a village on the Chatham Islands. A local tsunami in 1947 left seaweed in telephone lines over ten metres above the sea (and fish in the local Bar!) near Turihaura Point 10 km north of Gisborne. In 1960, a tsunami generated by another huge earthquake in Chile crossed the Pacific and caused a 6 metre-high tsunami at Gisborne, but fortunately its arrival at the New Zealand coast near low tide blunted its potential for damage.

Communities can learn to coexist with tsunami. Japan probably is the most advanced nation in learning to deal with tsunami. They have had much experience. On July 12th 1993, waves ranging from 5 to 10 metres high crashed ashore in Aonae, a small fishing village on Okushiri's southern peninsula within five minutes of a magnitude 7.8 earthquake centred perhaps 15 to 30 kilometres offshore in the Sea of Japan. Water washed over seawalls erected after past tsunami disasters. Currents swept up buildings, vehicles, docked vessels and material in coastal storage areas, transforming them into battering rams that obliterated all in their path. Collisions sparked electrical and gas fires but access by fire engines was blocked by debris. Despite the devastation, both warning technology and community education greatly reduced the number of casualties to only 239 dead. The Japan Meteorological Agency issued timely and accurate warnings and many residents saved themselves by fleeing to high ground after the main shock - even before the warning. Okushiri clearly demonstrated that the impact of tsunami can be reduced.

Our historical record is too short to give an accurate picture of the likely future frequency of disastrous tsunami in New Zealand. For tsunami from sources far outside New Zealand - tele-tsunami - we can look to the historical records of older Pacific nations in South America and Japan. At least half of the larger tsunami seen here in historical time, however, have been local ones - caused by major New Zealand earthquakes, and were not noted as significant tsunami elsewhere. How are we to learn of the true long-term



The "New Zealand" model of why the tsunami was so high at Sissano Lagoon uses a wave-focussing mechanism. We suggest that horizontal and vertical movement of a bulge on the steep sea floor raised a curved wave in the sea surface, where the curvature aimed and focussed the tsunami on the lagoon. Witnesses saw the sea drop below normal low tide before the large breaking wave struck. A wave of this shape, with a leading depression in front of it, is called an N wave.

frequency of large tsunami around the New Zealand coastline? Do we have combinations of offshore geological structures and sea-floor topography as found off northern Papua New Guinea that might cause focussed tsunami? These are some of the questions for which New Zealand tsunami scientists must seek answers to help communities plan and implement effective tsunami-hazard mitigation.

Warning centres now exist to detect distant tsunami generated elsewhere in the Pacific. They can warn that an earthquake capable of generating a tsunami has occurred and can predict when a tsunami might arrive if one is generated, but they cannot predict how big it might be. The global tsunami warning network is now deploying ocean-bottom sensors to detect passing tsunami to provide verification that a tsunami has been generated. Warning centres can alert authorities to tsunami arriving from far away with plenty of lead time to allow people in danger to move to safety. The only way to escape one generated locally is to know and respond very quickly to the warning signs: coastal earthquakes, however slight; receding or rising waters; strange or loud noises coming from the ocean. If you experience any of these, get off the beach quickly, onto the highest ground around or about a kilometre inland.

17 July 1998

Saundaun Tsunami

by Willem de Lange

Department of Earth Sciences, The University of Waikato

On 17 July 1998 at 08:49 GMT, a moderate earthquake occurred in the Saundaun province of north-western Papua New Guinea (PNG). The earthquake event was recorded by the Pacific Tsunami Warning Centre and a bulletin issued warning of the possibility of small local sea level oscillations. Sea level data from Micronesia indicated that the magnitude of any tsunami generated was small, and the event was considered insignificant. However, more than a day after the event, it became apparent that the earthquake was followed by a locally catastrophic tsunami, now known as the Saundaun Tsunami.

This event captured media attention; significantly more so than of any of the other 9 major tsunami events of the past decade, undoubtedly because of the extent of the human tragedy. Early reports described particularly gruesome scenes with human remains dangling from trees, while others were mauled by dogs and crocodiles. Estimates of the death toll kept rising daily, and the risk of epidemics kept the story in the front pages of national and international newspapers for as long as a week after the event.

Table 1 - Casualty figures for the Saundaun tsunami as at 2 August 1998

Location	Dead	Injured	Survivors
Warupu	1,071	369	1,460
Arop	863	0	1,404
Malol	126	0	3,616

As of 6 August 1998, the official death toll had reached 2134 (Table 1), approaching the 3000 deaths reported for the deadliest tsunami of this century, the 1933 event off the coast of Sanriku, Japan. The magnitude of the devastation from the Saundaun tsunami is made exceptionally intriguing by the moderate size of the earthquake and the extreme geographic concentration of the affected area. This type of event has been reasonably common in the historic record of New Zealand tsunami events, particularly along the west coast of the South Island and the east coast of the North Island.

The earthquake appears to have been generated near the triple junction of the Australian and Pacific (Caroline) and North Bismarck Plates. Specifically, the North Bismarck and Pacific (Caroline) plates collide obliquely in the region north of the Sepik River, with the Pacific Plate underthrusting the margin, or subducting beneath it. It has been argued that the thick crust (20 km) of the Earipik - New Guinea rise may be inhibiting the subduction process. Alternatively, intermediate depth seismicity beneath the Sepik Province and central Irian Jaya has been associated with Pacific Plate subduction. The juncture may be better described as an arc-continent collision. The intense shallow seismicity north of the Sepik river reflects the deformation of the overlying plate margin as it is dragged west-south-west by the underthrusting Pacific Plate. Finally, the section of the Caroline Plate involved may constitute an independently deforming sliver, characterised by diffuse seismicity in the region of the Admiralty Archipelago, and limited to the north by the West Melanesian Trench, a feature probably aseismic west of 149°E.

This tectonic setting has some similarities to New Zealand. The main features in common are a narrow continental shelf next to a deep trough produced by subduction, and high sedimentation rates on the continental shelf due to river discharge from adjacent highlands. This occurs along the west coast of the South Island and the east coast of the North Island. In the case of North Island, the presence of large sea-mounts on the descending Pacific Plate appear to be affecting the subduction process (see Lewis *et al.* in this issue).

From the seismological standpoint, the Saundaun earthquake was only of moderate size. Estimates of its conventional magnitudes are m_b = 5.9 and M_s =7.0. The seismic moment was determined by the Quick CMT algorithm at Harvard to be M_0=5.2 x10^{26} dyne-cm (M_w = 7.1). Mantle magnitude estimates computed at Papeete and North-western University (Chicago) were M_m =6.8. The slight discrepancy between body- and surface-wave magnitudes is upheld by the calculation of the estimated energy in the body waves, and of the slowness parameter (ø =-5.5). This indicates that the earthquake source was somewhat deficient in high frequencies, but it did not exhibit the strong character of slowness found in tsunami earthquakes, such as Nicaragua (1992; ø=-6.30) or East Java (1994; ø =-6.01). Further, the mantle magnitude of the earthquake was stable with frequency, and did not grow with period, as is the case for the tsunami earthquakes. The Saundaun earthquake was followed 20 minutes later by an aftershock of m_b = 5.6. Careful study shows that the aftershock was itself preceded by a smaller event, 30 seconds earlier, with magnitude m_b = 5.3. The aftershock has a mantle magnitude M_m =5.75 and ø =-4.80, indicating that it was not a slow event.

The preliminary epicentre of the Saundaun earthquake was given by the National Earthquake Information Centre (NEIC) at 3.10°S; 141.80°E, a location significantly inland, while the PTWC determined an offshore location. The epicentre location has been revised several times giving a provisional location of 2.932°S; 141.797°E, which is practically on the coastline, 7 km to the west of the Serai. The main aftershock location was at sea (2.916°S; 142.081°E), 7 km due north of Sissano Lagoon. The preliminary characteristics of the source of the earthquake, obtained by Japanese seismologists, suggested a fault area of 30 by 15 km, with a slip of about 2 m. This geometry would be in general agreement with a simple model in which the hypocenter of the main shock would be at the western end of rupture, and the aftershock would mark the position of the eastern end. The Harvard CMT mechanism can be interpreted either as shallow angle oblique subduction of the Caroline plate under the Sepik province, or as nearly pure dip-slip on a fault dipping steeply 79° NNE.

ITST deployment and procedures

To investigate this event, map the inundation and to determine whether the preliminary media reports of extreme inundation flows were indeed limited over a fairly small area of about 10 km as had been reported in the press, an initial International Tsunami Survey Team (ITST) with thirteen scientists, a medical specialist and two film crews from Australia, Japan, New Zealand and the USA was deployed. In different earlier incarnations the ITST has performed inundation surveys for the 9 major tsunami catastrophes of the past decade, in Indonesia, Japan, Mexico, Nicaragua, the Philippines, Peru and Russia. The ITST is normally an official party sponsored by UNESCO through the Intergovernmental Oceanographic Commission (IOC). However at present the necessary protocols have not been established to permit the rapid deployment of a UNESCO/IOC sponsored survey party. Therefore the official status of the ITST is usually established retrospectively.

Inundation surveys involve measuring tsunami inundation height and inland penetration distances, whenever watermarks and other indicators can be found. These data are highly ephemeral and can be easily lost due to storms, or recovery operations in the affected regions. Whenever possible, the team also measures aftershock distributions to better determine the rupture area. The ITST uses standard surveying gear, GPS receivers for locating the inundation marks consistently on maps, corers for sediment sampling, and portable seismometers. Standard procedures as set out in the IOC Post-tsunami Survey Field Guide were followed. For the Saundaun Tsunami was necessary to dispense with the written questionnaires due to literacy and language difficulties. However a list of questions was agreed on that were used in oral interviews with the aid of local translators.

The Tsunami Survey

The initial ITST team met in Port Moresby on July 31, although some problems were experienced due to the team being split between two different Travelodge Hotels located some distance apart. The Japanese and Australian members visited the PNG Seismological Observatory and the University of PNG to obtain any relevant information that could be provided. There were also informal discussions with the New Zealand High Commission staff, and RAAF and RNZAF personnel who had been involved in the relief work in the affected area. The PNG Commissioner responsible for the disaster recovery phase requested that the ITST proceed immediately to Aitape.

The ITST flew to Wewak, the closest access point to the affected area on the following day. On the evening of the 1st the survey plan was laid out. The ITST split into two teams; one part of the team included all the Australian, New Zealand and US members, and will be henceforth referred to as the US team for brevity. This team proceeded west by boat to Aitape the disaster control center, stopping at the offshore islands of Kairiru and Walis for measurements. The Japanese team travelled by four-wheel drive overland, stopping for measurements along the way, while some members of the same team would fly to Lumi and Vanimo to install seismometers for aftershock Table measurements. The ITST also met with representatives of the Australian Geological Survey Organisation (AGSO) and the PNG Seismological Observatory who were installing seismometers. The US team transported some of the AGSO seismic equipment into the field area.

The weather on the 2nd was stormy and both teams had significant difficulties taking measurements and reaching Aitape. The road from Wewak to Aitape involves a number of unbridged river-crossings, which were blocked by swollen rivers following heavy rain on the morning of the 2nd. Helicopter travel was limited as most of the available craft were being used to distribute relief supplies, and Aitape had run out of suitable fuel. However, it was possible to use helicopters to fly small groups in and out of the area at the start and end of the day, and when they were released from relief work. Members of both the US and Japanese teams briefed the local disaster relief agencies in Aitape on the morning of the 3rd and asked for permission to enter the affected area which had been closed. The ITST was warmly welcomed and obtained permission to work inside the closed area. Table 1 summarises the casualty figures available at the briefing.

Local authorities reported that the tsunami had brought out many superstitious beliefs among the populace, and had been blamed by some on impiety. Further, following a press release from SOPAC about the possibility of closely spaced pairs of tsunami events, there were rumours that a second tsunami was imminent. Hence the local authorities solicited the team's help in explaining to the survivors the causes of tsunami, this being the first tsunami to hit the area in recorded times. This was agreed to, and arrangements made for a series of public presentations and discussion sessions to be held later in the week.

The ITST revised the survey plans with the US team sailing to Sissano Lagoon immediately. Meanwhile one part of the Japanese team would attempt to drive as close to Sissano by four wheel drive vehicle as possible (there was no road from Aitape to Sissano), and the other would fly in by helicopter and meet the US team there.

Sissano Lagoon is fronted on the ocean side by a two narrow sand spits with a fairly narrow mouth and limited ebb-tide delta close to the western end of the lagoon. The lagoon is almost semicircular with the back shore about 4 km from the mouth (Figure 1). The villages of Arop were located on a sand spit at the eastern end of the lagoon, and the Sissano villages lay about 1 km west of the lagoon entrance. Profiles measured by the ITST indicated that the sand spits had a maximum elevation <3 m above mean sea level, and were normally less than 100 m wide. Apart from *Casuarina* sp., and a tree known locally as laulau, most vegetation on the eastern spit was severely to totally destroyed. Therefore there was no suitable route for evacuation of the spit.

There was no evidence of any of houses or their remains anywhere along the sand spits, other than a few inclined or tilted rows of foundation poles. Some were snapped off near ground level. The foundation poles were roughly smoothed poles 0.1-0.2 m in diameter. These were normally vertical, in two parallel rows, and supported bearers that carried the floor and walls of the houses. Most of the coconut palms, banana and sago plants that surrounded the houses were gone. The remaining coconut palms and not were bent close to the roots or had been uprooted. The average flow depth over the spit near the lagoon mouth was found to be

10 m, while near Arop it was 15 m. It was not possible to determine maximum runup, since the most landward penetration point was over 4 km away in a swamp where crocodiles had been reported.

The team inferred from the damage and sedimentary structures left behind that the water current induced by the tsunami over the sand spits was at least 10 $m.s^{-1}$, and most probably it peaked to twice this value. These estimates are consistent with eyewitness reports whose descriptions allowed local scientists to infer 15-20 $m.s^{-1}$ current velocities. It should be noted that the force of the tsunami current on an object is 1000 times greater than the force on the same object by a wind of the same speed. The team also observed sediment deposition of up to 15 cm over most of the sand spits, and evidence of scour and sediment splays.

Three team members sailed with a smaller boat inside the lagoon in an attempt to determine how far the wave had penetrated. Navigation was difficult because of the number of entire trees, tree-stumps and building debris that were scattered everywhere, and it was not possible to reach to the back of the lagoon without significant further danger to team members. However, they did notice that the debris from the community on the sand bar reached to the back of the lagoon which is swampy and fronted by mangroves. The height of the wave there was probably less than 1m, as there were several mostly intact houses on stilts that were taken to have been unaffected by the tsunami.

The teams flew helicopter sorties when equipment could be spared from the relief effort to determine inland penetration in the immediate vicinity of the lagoon for areas that were inaccessible on foot due to the very adverse conditions. The teams also interviewed numerous eyewitnesses in relief camps who helped put together the sequence of events. In total, 80 inundation data points were measured and 30 different topographic transects, covering densely an area of more than 40 km, with several points measured at Vanimo about 100 km west of Sissano and Wewak, 180 km east. Generally, as seen in Figure 1, the inundation heights diminished rapidly about 10 km east and west from the worst hit area which extended between Arop and Sissano. At the village of Serai, about 15 km from the lagoon mouth the inundation height was about 4 m, and there was no damage in the village. The coastal topography changed suddenly at about 7 km west of Serai, where measurements in a lumber mill suggested a 1.5 m runup height. There were no reports or observations of damage beyond that point, yet the tsunami was observed by eyewitnesses in Wutung, Vanimo and in Manus Island and recorded by tidal gage stations in Japan and Hawaii. In Wutung, on the border with Indonesia the wave height was reported to have ranged from 2-3 m. This value has yet to be confirmed by measurement.

It is clear that most of the tsunami was fairly narrowly focused onto a 40 km strip of coastline between the Rainbaum (Arnold) River and Aitape, and diminished rapidly to either side. This narrow focus of the tsunami energy is surprising for a seismic tsunami. From the measurements of flow directions the tsunami appears to have approached from the east near the Rainbaum River. In the worst affected area by Sissano Lagoon it propagated practically perpendicular to the shore. Closer to Aitape the waves clearly approached from the west. These data suggest that the tsunami spread radially from a source almost directly off Sissano. and then dissipated rapidly. This source does not correspond with any of the epicentres determined for the main shock.

The first wave arrived within five to ten minutes from the mainshock and was reported uniformly as a leading depression N-wave (LDN). This caused a noticeable recession of the water, and led some people to move towards the sea. The elevation wave following was reported as a wall of water, making thunderous noise resembling a jet aircraft. One person described the wave as C-shaped when observed from an angle, suggesting a plunging breaker. However, at the time of the tsunami the sun had just set and it would have been difficult to properly observe the wave. The first wave was followed shortly after by another two waves, the third of which was clearly smaller than the first two. It appears that all waves were closely spaced in time, suggestive of a highly dispersive wave train, rather than individual waves generated by strong aftershocks or sequential rupture. Most eyewitnesses indicated that the 3 waves occurred over a time span of 15-20 minutes, suggesting they were around 5 minutes apart.

Post survey debriefing

The ITST reassembled at Aitape on August 6 and briefed the authorities about its' findings. The relative location of the mainshock and aftershock is consistent with the reports that the aftershock was felt stronger than the main shock, neither of which produced ground motions stronger than Modified

Mercalli Intensity V to VI. Without suitable offshore measurements of the bathymetry and subsurface structure, it was not possible to confirm the likely cause of the tsunami. Most team members felt that a secondary mechanism such as a submarine landslide was the probable cause.

The team presented the following preliminary recommendations:

• There should be no relocation of people to locales which are fronted by water and backed by rivers or lagoons. Memorials should be built at the worst stricken locales to remind future inhabitants of this disaster and thus discourage future habitation of high risk locales. These memorials could be as simple as large signs.

• Schools, churches and other critical facilities should never be located closer to 400 m from the coastline, and preferably 800 m in at-risk areas.

• The local *Casuarina* tree species withstood the tsunami wave attack significantly better than coconut palms, and *Casuarina* forests should be planted in front of coastal communities, whenever possible.

• There should be evacuation drills annually on the anniversary of this disaster so that all people in at-risk areas know that if they feel the ground moving they should run as far from the beach as possible.

• Every family in an at-risk area should have a designated *Casuarina* tree with a ladder or carved steps to allow vertical evacuation of the able, when there is no other option.

• The residents in non-affected areas should return to their homes, after being briefed about what to do in the event that they feel a ground motion or if they see unusual water movements. (As in most other 1992-98 tsunami, this tsunami was preceded by an LDN)

• The local fishermen should be allowed to resume fishing in the open ocean only. The local authorities should collect samples of the lagoon water (using procedures described to them) and have them tested monthly to quantify the evolution of the water quality in the lagoon to determine when it would be safe again for fishing and habitation.

As promised in the first meeting with the authorities, public presentations were given at the local hospital and two local schools. In these presentations, the ITST explained that tsunami are natural phenomena, and that all communities bordering the Pacific Ocean are at risk. They also described the mechanism of tsunami generation and gave physical demonstrations of the difference between tsunami and swell waves. They discussed some simple warning signs that a tsunami may be imminent and stressed that ground motion is not always a precursor: anybody who lives close to the coastline should be on the lookout for unusual water motions and they should know what to do.

The teams also tried to explain some of the unusual phenomenon reported to them by eyewitnesses. Reports of the sea bubbling and of foul smelling gas and warm water stinging the eyes were attributed to the tsunami stirring up the stagnant bottom waters of Sissano Lagoon. The lagoon is normally calm, and quite possibly a layer of vegetable matter would have accumulated on the bottom, building up an oxygen-poor environment where noxious gases may have developed in the sediments. Some eyewitnesses had described the tsunami as “a water-fire infernal mountain of water with fire sparkles flying”. This was taken by locals as an explanation of the severe burns observed among the dead and some survivors. The team explained that most likely the tsunami had triggered bioluminescence, a phenomenon also known as “sea-fire” where *dinoflagellates* and other marine organisms emit light when stirred. The team had observed dramatic examples of bioluminescence in the wake of their boat over several nights while onboard, and speculated that this may have created the appearance of sparks flying as the wave approached. The burns reported were not from heat but from friction, which probably caused significant skin loss as the victims were dragged over hundreds of meters among debris and trees. Other victims had the skin flayed from exposed portions of their bodies, giving the appearance of being sand-blasted. This was attributed to the sediment carried by the tsunami waves.

Further investigations

The initial ITST investigation has been followed by several other survey teams, including a second International Tsunami Survey team, a group from the New Zealand National Society for Earthquake Engineering, and a cruise by the R/V Kairei. These later teams undertook more specialised investigations into the characteristics of the tsunami deposits and the geologic structures offshore of Sissano. The PNG Seismological Observatory also undertook a review of historical tsunami data for the Saundaun Province.

The historical data indicate that the Sissano area had experienced severe tsunami in the last 200 years, as there have been on other parts of the PNG coast. Here and elsewhere on the north coast there is a mythology of tsunami that doubtless stems from the pre-historic experiences. However, in historic time, which dates from the first mission settlements in the 1880s, tsunami appear to have been few and far between, and until the Saundaun Tsunami none had caused serious damage.

Possibly the largest recent tsunami near Sissano occurred in 1873. However, no quantitative data are available for this tsunami. A major earthquake occurred on 15 December 1907 triggering coastal subsidence forming the tidal Sissano Lagoon. Prior to the earthquake there was a small lagoon, called Warupu Lagoon, with two small islands inhabited by the Warupu clan. With the main earthquake about 100 km^2 of the coast subsided forming a large lagoon. Present day Sissano Lagoon is about 26 km west of Aitape with water depths estimated to be 1.8-4 m. A catalogue of PNG tsunami compiled by the Australian Geological Survey reports that this earthquake was accompanied by a tsunami, but the source of this information is not given. Contemporary written records describe the events of 1907 in some detail and make no mention of a tsunami. The PNG Seismological Observatory concluded that there was no tsunami.

There was major onshore earthquake near Sissano in 1935 that did generate a tsunami, but reports by local authorities refer to the tsunami as a moderate event, and do not record any deaths or significant damage. Probably the 1935 tsunami was small, less than 2 m high, and not particularly destructive. For example, the big church at Sissano that was built in 1926 appears to have survived this tsunami unscathed, but was totally destroyed in 1998. In 1926 the church was much closer to the water's edge, and the coastal strip has since built seawards by at least 60 m (indicative of the high sedimentation rates in this area).

Similarly the tsunami reported in 1951 was either a non-event (some residents have no recollection of it, and there is no official record of it other than a hand-written note in the Observatory files) or, if it happened, was too small to cause death or damage. Hence it was concluded that, at least in historic times, the Sissano coast has been no more at risk from tsunami than other parts of the coast of PNG, Indonesia, Solomon Islands and Vanuatu that are in zones of earthquake activity.

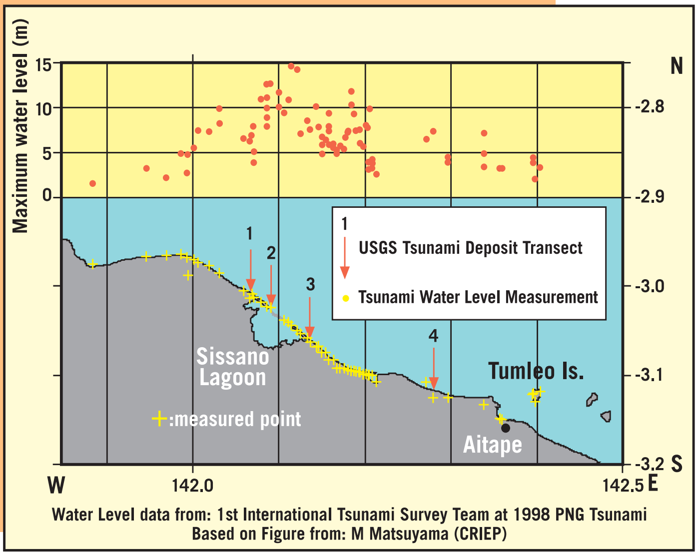
Soon after the initial ITST returned, a second group of international scientists was organised to retrieve the seismographs, collect more water-level and velocity data, assess damage to buildings and structures, and to examine the sediments left behind by the tsunami. The 2nd ITST arrived in Aitape, Papua New Guinea on September 29, 1998 and included representatives from Japan, the United States, Korea, and Papua New Guinea.

Of particular interest to New Zealand was the examination of the sedimentary deposits left by the tsunami. When sediment is deposited by a tsunami and preserved, a geologic record of that tsunami is created. By looking at the sedimentary sequence in an area, it may be possible to identify such deposits and infer the occurrence of prehistoric tsunami. The recognition of deposits from past tsunami can extend the relatively short or non-existent historical record of tsunami in an area. Because it is not yet possible to predict when a tsunami will occur, obtaining a record of prehistoric events may be one of the only means to assess future risk.

Therefore a primary goal of the 2nd ITST was to determine whether or not the Saundaun Tsunami produced a recognisable sediment deposit, and if it did, what were its' characteristics. The ability to interpret the height, power, and extent of a tsunami from its deposits is not only valuable for understanding the Saundaun event, but also for identifying and deciphering tsunami deposits, both ancient and modern, world-wide. Another goal of the survey was to determine whether the sedimentary record in the Sissano area contains information about past tsunami.

The ITST measured land elevation, flow depth, flow direction, and tsunami deposit thickness and character along cross-shore transects at four sites:

Figure 1 - Main impact area for the Saundaun tsunami of 17 July 1998. The crosses indicate locations surveyed by the ITST, and the dots represent the measured tsunami maximum water levels at each survey location



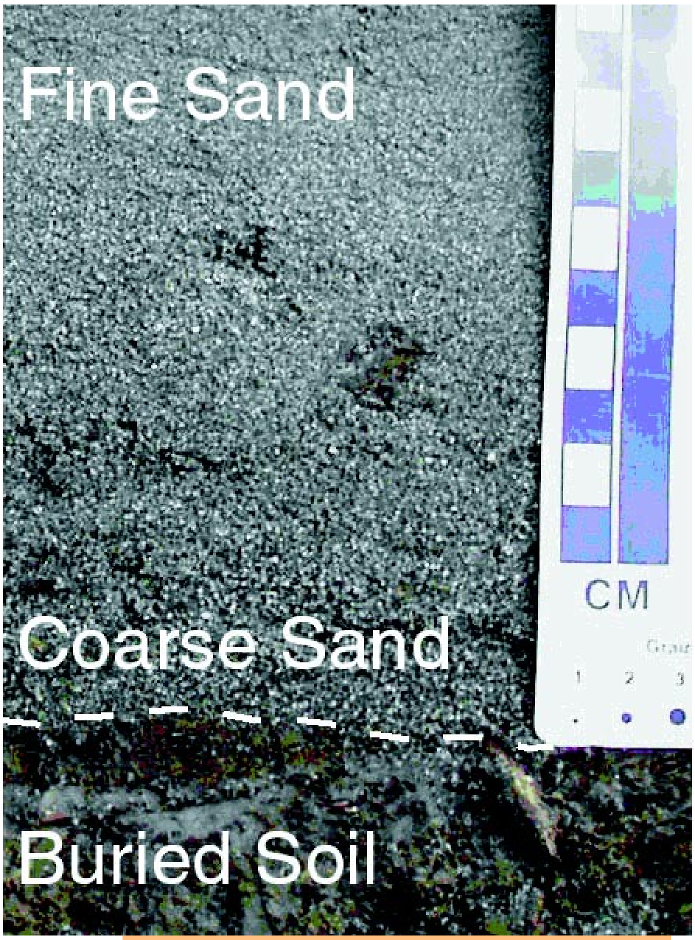


Figure 2 - Typical tsunami deposit from Arop transect. The deposited a normally graded, grey-coloured sand, here about 10cm thick, on a brown soil containing roots.

Sissano Village, East Sissano Spit near the entrance to the lagoon, Arop School, and Waipo Village, midway between Arop and Aitape (Figure 1). Results at the four sites were similar, and preliminary results for the Arop School transect were released via the World Wide Web. The Arop School transect was near the village of Arop, which was totally destroyed by the tsunami. Small pits were dug along this transect to examine the deposits left by the tsunami. The deposits were measured and described in the field and samples were taken for laboratory analyses.

Tsunami deposits were common and were identified as grey-coloured sand typically overlying a brown soil containing many roots (Figure 2). In places, plants were found bent over and buried by the sand. In other places, plants and roots were removed by the tsunami, leaving an erosive base to the deposit. The lower part of the tsunami deposit sometimes included rip-up clasts of the underlying muddy soil. The recently deposited tsunami sand is believed to have come from offshore of the beach as numerous sand dollars were found near the surface of the deposit. Another common characteristic of recently deposited tsunami sand was normal grading (a decrease in the size of the sand grains from the bottom to the top in the deposit).

Little internal structure was found in the tsunami deposits, although in a few places some faint horizontal stratification was observed at the top of the deposit. Overall, the recently deposited tsunami sand was relatively uniform in thickness (5-10 cm), and extended from 60 to 675 m inland, pinching out 50 m from the maximum limit of runup. The deposit fined landward (near the shore the sand particles were larger than the sand farther inland). Local variations in the thickness of the sand deposit were associated with small local topographic variations.

Information from the initial ITST suggested that the maximum water level near the beach at the Arop Village transect site was approximately 10 m. Flow depth indicators (e.g., water marks on structures, debris wrapped around a tree or other obstruction) further inland, about 500 m from the shoreline, record maximum water depths of 1-3 m.

The Arop School transect was chosen as a good site to look down into the sedimentary record for evidence of past tsunami. Away from sandy river or ocean sources, the depositional environment at this site was probably that of a quiet water lagoon. At 135 m from the shoreline, several long push cores were taken and described. Several 2-m long cores were taken and described, and one 4-m long core was obtained. The top metre of each core was characterised by a 5-10 cm thick normally-graded sandy layer at the surface (deposited by the July 17 tsunami); this is underlain by 20-30 cm of a brown muddy soil. Beneath the soil is a uniformly gray muddy sediment. A thin coarse silt/fine sand layer is present approximately 120 cm below the surface. This layer is 3-4 cm thick and was found at a similar depth in each of the cores. This layer was likely deposited by a past tsunami. Material just below this fine-sand layer was obtained for possible dating to determine the approximate age of this thin layer.

The Japanese research vessel Kairei visited the Sissano region on a joint SOPAC/JAMSTEC cruise arranged by the South Pacific Geoscience Commission (SOPAC) and the Japan Marine Science and Technology Centre (JAMSTEC). The first leg of the cruise was during December 1998 and January 1999, undertook preliminary bathymetric observations of the offshore area beyond the 200 m depth contour. The bathymetric data acquired during the cruise indicated the presence of a potential fault

40 km long and a potential underwater landslide, both in the vicinity of the likely tsunami source. As the data starts about 5 km from the shore there may be other features such as gas/mud diapirs and fault scarps uncharted. However the pre-existing bathymetric data does suffice to explain significant tsunami focusing near the village of Arop (on the spit) for either an earthquake or landslide generated tsunami from the area surveyed.

In March 1999, the R/V Natsushima to examined the fault and landslide with a ROV in an attempt to date and characterise these potential tsunami sources. A nearshore survey and study of the lagoon are being proposed to SOPAC to check for potential sources closer to shore than examined by the R/V Kairei. Several days of seismic profiling will be conducted by the American R/V Ewing sometime in July 1999. These data are crucial for assessing whether this coastline is both tsunami prone and tsunami vulnerable.

The cause of the tsunami

At present it is uncertain whether the source of the tsunami was seafloor deformation, a coseismic submarine slump or both. Numerical models in both Japan and the US estimated a preliminary vertical deformation of 2m, which by itself can not explain the size of the tsunami. It is also interesting to examine Abe's tsunami magnitude. Using his formula $M_t = \log H + B$, where H= 0.13±0.03 m is the average maximum tsunami amplitude for this event as recorded in Japan, and B=8.4 is a constant which depends on pairs of source and observation regions, then $M_t = 7.51$, suggestive that this was a very efficient earthquake in tsunami generation. Further evidence for the generally small character of the earthquake source includes the maximum Mercalli intensity VI reported near the epicentral area, and failure to observe any permanent changes in sea level or other spectacular surface expressions of the shaking. However, liquefaction was reported on the beach at Arop.

All this would point to the possibility of a landslide. In addition the tsunami runup distribution suggests a highly directional tsunami, which is characteristic of landslide tsunami, but not normally associated with seismic tsunami . The roughly shore parallel orientation of the seismic rupture would also make it difficult to account for the distribution, unless there is some focusing by bathymetric features on the shelf. Finally there is the unresolved inconsistency between the earthquake location and the tsunami source. Simple methods are available to predict the approximate landslide dimensions and location required to generate tsunami. These indicate that the tsunami need involve a sediment thickness of 35-50 m, in initial water depths of 100-500 m. Given these constraints a local landslide moving a volume of 3-5 km^3 would generate a LDN wave that would runup up to 7m. These values are consistent with a relatively small submarine landslide, and do not require any special behaviour or characteristics for the landslide.

Nonetheless, some seismologists argue that high-angle reverse faulting events have occurred in the past near the epicenter of this event, and that an almost vertical fault plane with a the source located in deep water would explain the runup observed without a landslide. So far the available offshore data are equivocal with both a potential fault zone and landslide being identified in bathymetric data. A resolution of this puzzle will have to wait until the fault plane is mapped better when the measurements of aftershocks is completed, and the bathymetric, seismic and side-scan surveys undertaken to map the offshore bathymetry and structure have been fully analysed.

Further work will continue to model the source and determine the potential of future tsunami from the same subduction zone, and to find out whether the possibility for a transpacific tsunami exists, either from the seafloor displacement or from a coseismic slump. Understanding this event will hopefully lead to the production of inundation maps for the north coast of PNG. Having access to inundation maps helps the local authorities locate schools, hospitals, and other critical facilities.

Implications for New Zealand

Clearly the type of infrastructure present along the New Zealand coast is quite different to that along the least developed coastal area in PNG. Therefore it could be argued that New Zealand would not suffer the same number of casualties if a similar event were to occur here. Nonetheless there are a number of important aspects of this disaster that are relevant to New Zealand.

Firstly there are significant similarities between the physiography of the affected area and portions of the New Zealand coast. The east coast of the North Island and northern South Island has a narrow continental shelf that drops steeply into a submarine trench, similar to the Saundaun province of PNG. Only one definite landslide tsunami is known from the east coast of the North Island; a 15 m high surge caused by a landslide during the 1931 Napier Earthquake. However surveys of the offshore bathymetry and stratigraphy have shown the presence of very large prehistoric submarine landslides off Hawke Bay. These landslides involve volumes at least an order of magnitude larger (>50 km^3) than that postulated for the Saundaun Tsunami. The analysis of tsunami hazard undertaken for the region by the Hawkes Bay Regional Council was based on earthquake induced fault motions, and did not consider landslide effects.

The west coast of the South Island also has a narrow continental shelf that drops off to abyssal depths. Historical records of New Zealand tsunami events show that several landslide generated tsunami have occurred along the west of the South Island. There are also a couple of unexplained events that were probably caused by landslides. So far these events have mostly involved subaerial landslides that have produced smaller tsunami than the Saundaun event. The one exception is the event that killed an unknown number of Maori around 1820. This wave or waves appears to have been similar in size and extent, but the source mechanism is unknown.

Researchers in Australia have also identified deposits along the coasts of New South Wales, Queensland and Lord Howe Island that they attribute to tsunami with heights in excess of 15 m. These were attributed to large submarine landslides on the flanks of the Hawaiian volcanoes. However numerical modelling indicates that these would not have produced sufficiently large waves in Australia. The wave heights in New Zealand were 2-3 times the Australian heights, but still insignificant. The current theory is that they were produced by either bolide impacts or submarine landslides in the Tasman Sea. As yet there is no record of matching tsunami deposits in New Zealand.

Therefore, historical and other evidence indicate that New Zealand has experienced landslide generated tsunami in the past. These have been quite large, but localised in their extent. It is probable that New Zealand will experience landslide generated tsunami in the immediate future (1-30 years). It is likely that this event will be in the range 5-15 m, but in some areas it is possible that the waves can be much larger (>25 m).

Most of the New Zealand coast has a higher relief than the Saundaun province. Coastal dunes are typically 6-15 m above mean sea level. However there are regions where development has greatly reduced the height to provide views and access to the beach. These regions have a correspondingly greater hazard.

New Zealand construction standards are more stringent than most of those evident in the affected region. Some of the buildings such as the school and mission at Sissano seem to have been built to a similar or better standard than holiday homes along the New Zealand coast. The tsunami waves completely destroyed the buildings. The damage to the buildings was caused by three main processes:

- floating and subsequent collisions;
- the direct impact of the wave and the associated currents;
- impact by debris carried by the wave.

The buildings were mostly unaffected when the flow depths were less than 1 m, as in many cases the main bearers of the floors and walls were above this level. Many New Zealand houses are closer to the ground. The buildings suffered increasing damage as the flow depth increased from 1 m to 3 m, and at greater depths the buildings were totally destroyed. Given the height of most of the New Zealand coast, the flow depths produced by a similar event would mostly be less than 3 m.

A significant proportion of the casualties could have been avoided by timely evacuation. This was not possible for the population on the narrow sand spits, but it was an option for the rest of the affected region. Unfortunately lack of tsunami awareness meant that most stayed in the hazardous region to watch the water recede, and many were unable to outrun the following wave. It is probable, given historic behaviour in New Zealand, that a significant number of people would also fail to evacuate in time here. The standard Civil Defence publicity associates tsunami with strong earthquakes. In areas where submarine landslides are considered a realistic possibility, it is desirable to revise this to include all felt earthquakes and any unusual water motions.

Finally, although the actual cause of the Saundaun Tsunami has not yet been determined, it is evident that the triggering earthquake did not display any special characteristics that would have allowed prediction of the resulting tsunami. The PTWC assessed this event as only causing small localised sea level disturbances. A similar event here would produce the same assessment. The ITST could not determine any remote method that could have provided a suitable warning for the affected region. The two difficulties were the absence of any earthquake signature that the ITST could associate with a large tsunami, and the short travel time (15 minutes or less). The same would apply in New Zealand.

Tsunami hazard
And inundation modelling for the *Firth of Thames*

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The Firth of Thames lies at the southern end of the Hauraki Gulf, a semi-enclosed sea next to the largest population centre in New Zealand, the Auckland metropolitan region (Figure 1). The Hauraki Gulf is extensively used for recreation and has a significant amount of infrastructure located around its shores. Therefore the potential tsunami hazard is of concern to regional and local planners around the Hauraki Gulf. This region has recorded at least eleven tsunami and one meteorological tsunami (rissaga) since 1840 (Table 1). Most of these were small events. However 3 tsunamis (Chile 1868, Chile 1960 and Alaska 1964) and the rissaga (Krakatau 1883) were damaging events with wave heights up to 2 m.

Since the last significant tsunami in 1964, there has been considerable investment in coastal infrastructure including marinas, tourist attractions and upgraded port facilities. This has accelerated with the upcoming Americas Cup Regatta of 1999-2000. Both the Auckland and Waikato Regional Councils have initiated Lifelines Projects to identify the impacts of a variety of hazards, including tsunamis. This requires potential tsunami hazard to be assessed at a local scale.

The historical data are relatively scarce, particularly for the largest events in 1877 and 1883, at which time the Hauraki Gulf was relatively sparsely populated. Moreover, local sources may produce damaging tsunamis but none has occurred during recorded history. Therefore numerical modelling of potential tsunami events provides a powerful tool to obtain data for planning purposes.

Three main scenarios have been identified for numerical modelling:

1) A teletsunami event from an earthquake off the West Coast of South America. Historically this region has produced the largest teletsunamis in the Hauraki Gulf.

2) A tsunami generated by a local earthquake along the Kerepehi Fault. This fault bisects the Gulf, has been active during the last century at the southern inland end, and is overlain by a considerable thickness of soft sediment that may amplify the seismic waves.

3) A tsunami generated by a volcanic eruption within the Auckland Volcanic Field. This field has involved a series of mainly monogenetic basaltic eruptions over the last 140,000 years. Many of these eruptions have involved phreatomagmatic eruptions around the coastal margins, or within the shallow waters close to Auckland.

The last of these scenarios is discussed in a

Figure 1 - Map of the Hauraki Gulf - Firth of Thames region.

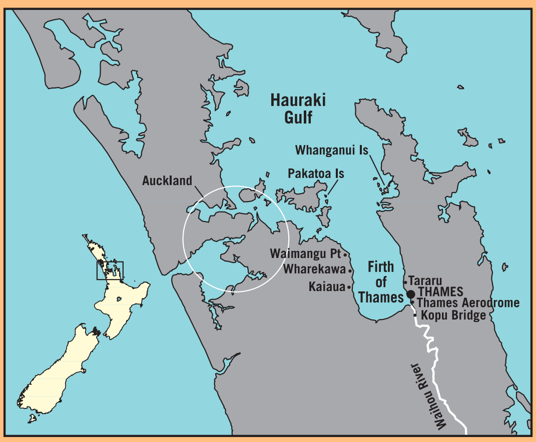


Table 1. Tsunamis recorded within the Hauraki Gulf between 1840 and 1994. The maximum runup is the maximum elevation above the expected tidal elevation.

Date of event	Source	Maximum runup
13 August 1868	Earthquake, northern Chile	2.0 m - Port Charles
10 May 1877	Earthquake, northern Chile	3.7 m - Port Charles
27 August 1883	Eruption, Krakatau, Indonesia	1.8 m - Auckland port
4 November 1952	Earthquake, East Kamchatka	0.1 m - Auckland port
22 May 1960	Earthquake, southern Chile	2.5 m - Little Omaha Bay
28 March 1964	Earthquake, Alaska	0.5 m - Auckland port
14 January 1976	Earthquake, Kermadec Islands	<0.1 m - Auckland port
22 June 1977	Earthquake, Tonga	<0.1 m - Auckland port
19 December 1982	Earthquake, Kermadec Islands	Unknown
20 October 1986	Earthquake, Kermadec Islands	<<0.1 m - Auckland port
18 June 1993	Earthquake, Kermadec Islands	Unknown

separate article in this issue. The remaining two will be considered in relation to the tsunami hazard in the Firth of Thames.

The Kerepehi Fault

The geology of offshore regions of Auckland and Northland has not been studied beyond the level of general reconnaissance. Moderate to poor quality seismic reflection and refraction profiles were collected in the late 1960s and early 1970s for petroleum exploration. Seismic line separations of 20-50 km provide low density data from which interpretation of the location and continuity of major geological structures can be made.

However, throughout the Hauraki Gulf geophysical data suitable for determining the activity, and hence tsunami generating potential, of faults is absent or of doubtful quality. There is poor offshore stratigraphic control for determining the age of any displaced sediments. Basement rock crops out on the seafloor, so that some faults have seafloor expression. There was no evidence to determine the age of faulting, and potential for future movement. Furthermore, because of the poor petroleum prospectivity it is unlikely that any additional oil exploration data will be collected soon.

Several faults can be recognised by scarps along the seafloor in the Hauraki Gulf, particularly near the continental shelf break. Faults have both northerly and north-westerly trends, although the north-westerly trend is probably late Miocene in age and offset by younger northerly trending faults. Previous studies have suggested that the faults recognised from the inland Hauraki Depression continue north-west into the Hauraki Gulf to near Whangarei. However, these faults were not identified by seafloor scarps, possibly due to high sedimentation rates, particularly in the Firth of Thames. These faults delineate a structural feature, the Hauraki Rift, that controls the form of the Hauraki Gulf. Due to relative youth of the Hauraki Rift (Pliocene) and the presence of late Pleistocene active fault traces on land, all faults in the Hauraki Gulf are considered active.

The best known active geological structure within the Hauraki Rift is the Kerepehi Fault which was inferred to trend NNW through the middle of the Firth of Thames. Using data determined for the southern, inland sections of the fault, Kerepehi fault is inferred to have an average vertical separation rate of ~0.5 $mm.y^{-1}$, recurrence interval of 4,500–9,000 years and Most Credible Earthquake (MCE) of M_w=6.9. This moment magnitude has been calculated assuming a surface rupture length of about 25 km, an average fault slip of 2.5 m, a focal depth of 10 km and a fault dip of 60°W.

Normally it is considered that the minimum magnitude required to generate a tsunami is M_w = 7.3. However, some tsunamis are generated by smaller magnitude earthquakes known as tsunami earthquakes. These involve earthquakes where a significant (>10%) proportion of the rupture occurs in soft sediments, or where the earthquake displacements occur within a confined water body. Both of these conditions may apply to an earthquake involving the Kerepehi Fault.

There have been no historic earthquakes involving the offshore Kerepehi Fault. Therefore the first stage of the investigation involved locating and characterising the offshore sections of the fault. Seismic reflection data collected by the University of Waikato, Defence Scientific Establishment, and the New Zealand Oceanographic Institute during the 1980s were collated and examined. A total of 135 km of shallow seismic sub-bottom profiles were found to cover the Firth of Thames.

The data show that the offshore Kerepehi Fault is similar to the onshore feature, trending NNW up the central Firth of Thames (Figure 2). The fault is subdivided into four segments evident between the Hauraki Plains coast and offshore of Waiheke Island by three WSW-ENE trending transverse faults. The full extent and the exact location of the most northern segment could not be determined and two locations were proposed (D1 and D2 in Figure 2). Examination of subsurface reflectors indicates that the average fault displacement per earthquake is 2.1 m for segments A, B, C, and D1. This is consistent with the reported displacements on land and the M.C.E. scenario given above. However the displacements for segment D2 appear greater with an average displacement of ~7.3 m.

Five local earthquake scenarios were developed (Table 2); one for each of the recognised fault segments. The fault segments identified are shorter than the 25 km rupture length identified for the on land MCE scenario. Therefore two simulations were undertaken for each scenario: a narrow rupture zone (1.25 km) constrained by the underlying geological structures and involving rupture of two adjacent segments; and a wider (5.8 km) rupture zone involving only one segment. The varying segment lengths, and hence rupture lengths account for the variations in MCE magnitude given in Table 2.

Tsunami generation and propagation

The potential impact the various MCE scenarios have on tsunami hazard around the Hauraki Gulf was investigated using a linear deep water generation and shallow water propagation finite element model, 'TSUNAMI'. The modelling determined that a linear relationship existed between the fault displacement along each segment and the maximum wave height predicted for any location in the Hauraki Gulf. Therefore the expected wave height can be easily assessed if better data become available to redefine the MCE characteristics. More importantly the maximum predicted wave heights are small for all scenarios except those involving segment D2 (Table 3). Segment D2 scenarios involved a much larger displacement, and the rupture zone included Pakatoa Island which experienced the largest wave heights.

The small magnitude of the generated tsunami (Segment D2 being the exception) could be due to the shallow water conditions, which should strictly be simulated by a non-linear model. Therefore empirical parametric equations were used to determine if the model predictions were sensible. The formulae estimate the height of regional tsunami at a particular site as a function of propagation distance and the source characteristics of the generating earthquake. They are given by:

$$\log H = M_w - \log R - 5.55 + C$$

where H is the tsunami height (m), M_w is the earthquake magnitude, and R is the distance from the source (km), and C is a constant that depends on the tectonic setting of the earthquake. Typically C = 0.0 for tsunami generated in fore-arc settings, and C = 0.2 for back-arc tsunami generation (such as the Hauraki Rift). This relationship tends to overpredict the wave height for locations close to the source, so it should only be applied for propagation distances that exceed a threshold (R_o in km) given by:

$$\log R_0 = 0.5 M_w - 2.25$$

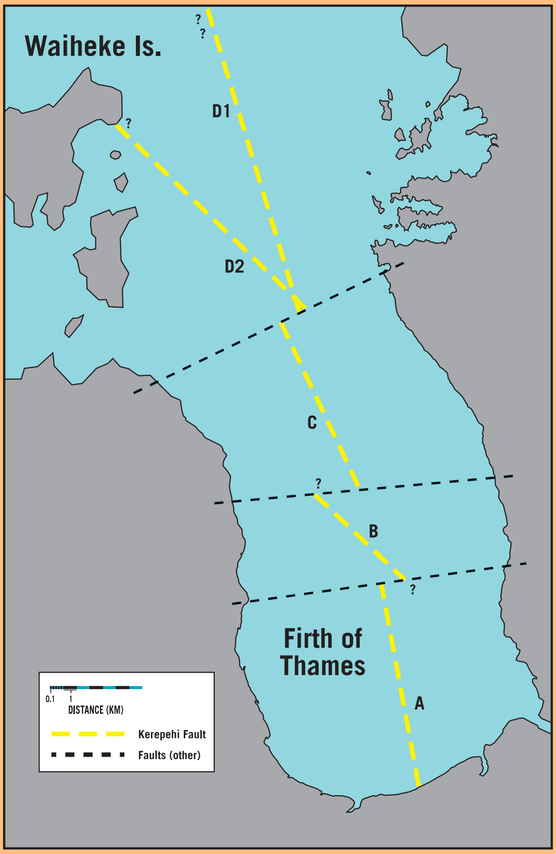


Figure 2 - Map of the trace of the offshore Kerepehi Fault in the Firth of Thames. The position of the northern segment is poorly defined by the available seismic data, so two possible locations were defined (D1 and D2).

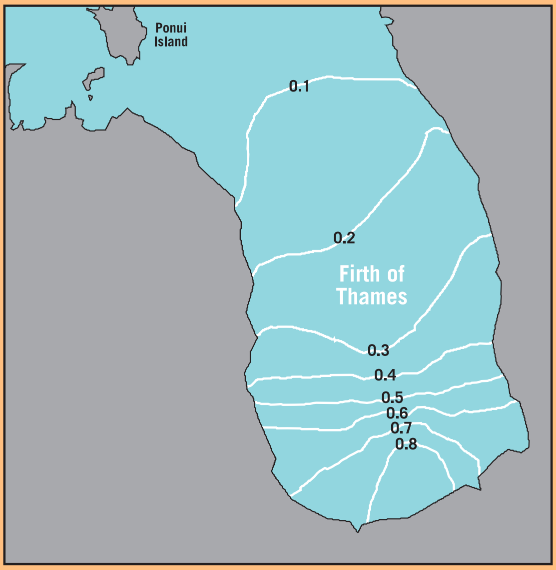


Figure 3 - An example of a tsunami simulation for an earthquake located on segment A of the Kerepehi Fault in the Firth of Thames. This contour map of tsunami wave heights 25 minutes after the earthquake shows the effect of water depth. The speed of a tsunami decreases with decreasing water depth. Therefore, although the waves were initially directed east-west due to the NNW orientation of the rupture zone, the tsunami energy is channeled northwards by deep water. At this stage the largest waves are at the southern end of the Firth of Thames near the source region.

For sites closer than this critical distance, the wave heights are given by an alternate relationship:

$$\log H = 0.5 M_w - 3.30 + C$$

Comparison of the two predictive methods showed that the TSUNAMI numerical model predicted wave heights that were about 25% of the magnitude of heights predicted by the empirical relationships. This difference is greater than the factor of uncertainty associated with the formula (1.5), suggesting that the linear model TSUNAMI under-predicted the maximum shoreline wave heights. A further check was undertaken using a non-linear hydrodynamic circulation finite difference model 3DD. This model could not simulate tsunami generation, so an initial surface deformation was calculated for the tsunami source region, and this was then allowed to propagate away.

Segment	Southern End Point Longitude (E)	Latitude (S)	Northern End Point Longitude (E)	Latitude (S)	Likely Vertical Displacement / Event	M.C.E. (M_w)
A	175°27′53″	37°12′32″	175°25′15″	37°04′41″	2.1	6.7
B	175°25′51″	37°04′31″	175°21′30″	37°01′19″	2.1	6.5
C	175°23′47″	37°01′09″	175°19′55″	36°54′32″	2.1	6.7
D1	175°20′35″	36°54′12″	175°15′38″	36°42′50″	2.1	6.8
D2	175°20′59″	36°54′03″	175°11′36″	36°46′48″	7.35	7.1

Table 2. Most Credible Earthquake (MCE) scenarios used to simulate local tsunami generation in the Firth of Thames by the Kerepehi Fault. Summary of the location, likely vertical displacement for a given displacement event and Most Credible Earthquake Magnitude presented as MW values.

Segment	1.25 km surface rupture width: Maximum wave height (m)	Location	5.8 km surface rupture width: Maximum wave height (m)	Location
A	0.64	Southern Firth of Thames	0.78	Southern Firth of Thames
B	0.05	Wharekawa	0.20	Kaiaua
C	0.08	Waimangu Pt.	0.33	Waimangu Pt.
D1	0.09	Whanganui Is.	0.41	Whanganui Is.
D2	2.18	Pakatoa Is.	3.19	Pakatoa Is.

Table 3. Summary of the uncorrected maximum tsunami wave heights predicted by the TSUNAMI linear generation and propagation model for 1.25 km and 5.8 km surface rupture widths. Allowing for non-linear effects the predicted wave heights should be doubled.

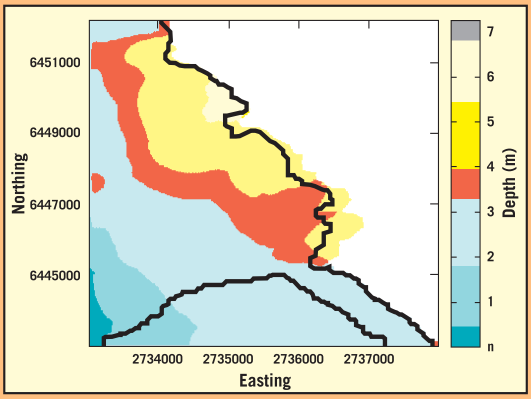


Figure 4 - Inundation map for the Thames region. Maximum water elevation is expressed in metres above mean sea level and represents inundation levels resulting from the input of a 3 m amplitude wave at the western grid boundary. Pink stars represent the coastline. The stop banks have been removed for this simulation, allowing the low lying land on the southern coast of the Firth of Thames to be flooded. The maximum runup occurs just north of Moanataiari

Comparison between the predictions of the non-linear 3DD and linear TSUNAMI models showed that the linear tsunami generation component of TSUNAMI was under-predicting the initial displacement over the fault zone by 50%. Correcting for this doubles the maximum wave heights at the shoreline. It was also found that the model time steps used during the generation phase in TSUNAMI were too short, causing unrealistic wave propagation velocities. Correcting the time step did not affect the predicted wave heights, but changed the wave arrival times, which vary from 15-25 minutes after the earthquake around the Firth of Thames coast.

The adjusted results were in good agreement with the empirical relationships for the deeper fault segments (D1 and D2), but were still about 50% lower for the shallower segments. In the future a non-linear tsunami generation model may be applied to the shallow fault segments. However, at present a suitable model is not available.

Even allowing for the non-linear corrections, none of the scenarios generated a detectable wave within metropolitan Auckland. The effects were confined to the Firth of Thames and nearby islands. This is due to high dissipation of tsunami energy in the shallow waters around the inner Hauraki Gulf Islands. It is still possible that an earthquake further offshore in the deeper waters of the outer Hauraki Gulf may produce a significant tsunami affecting Auckland. However insufficient data exist to define the source characteristics for such an event.

Teletsunami propagation

Teletsunami propagation into the Hauraki Gulf were simulated using both TSUNAMI and 3DD. Due to the lack of suitable data for the deep water characteristics of teletsunami, a standard wave amplitude of 0.5 m was used for all the simulations. A range of wave periods from 10 to 30 minutes were used for the simulations. Historical data indicate wave periods of 20-30 minutes in the Hauraki Gulf. However data from other locations give periods around 1 hour for the same event (Derek Todd's article in this issue). These variations suggest that teletsunami force oscillations of different periods on the continental shelf and within embayments (Derek Goring's article). Therefore simulation of teletsunami events can only be used to indicate the tsunami behaviour, until data are available to determine the offshore characteristics of teletsunami and their forced oscillations.

The teletsunami simulations for the Hauraki Gulf suggested that the maximum predicted rise above mean sea level at any location increases with increasing tsunami wave period. Also the confined nature and shallow waters of the Firth of Thames focused wave energy resulting in an increase in tsunami wave height, with the largest wave height above mean sea level observed at Tapu. The wave heights at Auckland were greatly attenuated, consistent with historical data.

Inundation modelling

The impact of a tsunami can vary greatly along the coast, even if the offshore wave height remains constant. Therefore one of the major areas of international tsunami research is the development of suitable models to predict tsunami inundation. New Zealand is participating in the Tsunami Inundation Modelling Exchange (TIME) project sponsored by the International Union of Geodesy and Geophysics (IUGG) and the Intergovernmental Oceanographic Commission (IOC). The TIME project has developed a variety of tsunami numerical models, including a non-linear finite-difference inundation model TSUNAMI-N2. This model was used to examine tsunami inundation around Thames.

Inundation modelling requires detailed bathymetric and topographic data. For this simulation a digital terrain model for Thames was constructed from topographic data provided by the Waikato Regional Council and hydrographic data from Royal New Zealand Navy charts. This model was used to generate elevations at 20 m intervals for the coast from the Kopu Bridge over the Waihou River to Tararu. Most of this coast has a stop-bank with a height of 3.5 m above chart datum, which is sufficiently high to hold back all of the waves predicted by the Kerepehi Fault simulations, and observed during historical teletsunami. Therefore a 3 m amplitude wave was chosen as a worst case scenario to test the software. This corresponds to a maximum elevation of 3 m above chart datum, which is comparable to the storm surges that damaged Thames in recent years.

TUNAMI-N2 indicates that the most severe run-up heights occur in the region extending from south of Tararu to northern Moanataiare/Kuranui Bay (Figure 4). If the stop banks were not present, the most severe horizontal inundation is predicted to occur on low lying land south of Opani Point where run-up heights in the order of 0.5-3.0 metres flood farmland and homes adjacent to Orongo Road. With the present stop banks, the most severe horizontal inundation should occur near the Thames Aerodrome and Rhodes Park, where maximum run-up heights are approximately 4 m and extend a maximum distance of approximately 450 m eastward of the coastline. The model also showed that water levels in the Waihou River would steadily increase as water was unable to completely drain out between successive waves.

The TUNAMI-N2 model also predicted overland flow velocities of 10-20 $m.s^{-1}$, which would cause considerable damage to structures. However the model does not take into account the presence of buildings and vegetation that might retard the flow. This requires suitable friction coefficients determined from historical tsunami inundation patterns, which are currently not available.

Implications for hazard management

The greatest hazard for the Firth of Thames, and Thames in particular, is associated with displacement along the southern offshore Kerepehi Fault segment, which produces wave heights in the order of 1 m. The largest tsunami wave heights (~7.4 m) were generated by displacement in deep water and had the most severe impact upon Pakatoa, Ponui, and Rotorua Islands. The maximum mainland wave height resulting from displacement along the Kerepehi Fault impacts at Deadmans Point and is in the order of 2 m. However the overall tsunami hazard associated with displacement along the Kerepehi Fault is low.

The Firth of Thames amplifies teletsunami waves by about 50 % of their amplitude in the outer Hauraki Gulf. Therefore they will be more hazardous in the Firth of Thames. However the wave heights associated with large historic teletsunami have been moderate, comparable to the normal tidal range, and existing coastal protection structures would prevent inundation under most circumstances. Therefore teletsunami represent a less significant hazard than tsunamis locally generated along the Kerepehi Fault.

Inundation modelling for Thames indicates that the maximum runup occurs between Tararu and Moanataiari, and land adjacent to the Thames aerodrome experiences the greatest horizontal inundation. This requires a combination of tides, storm surge and tsunami to increase water levels in the southern Firth of Thames to more than 3 m above chart datum. The tsunami amplitude would still need to exceed 2 m for significant inundation to occur.

REGIONAL *TSUNAMI* STUDIES: CANTERBURY & OTAGO

by Derek Todd
Tonkin & Taylor Ltd

Introduction

In recent years both the Canterbury and Otago Regional councils have undertaken scoping studies on the potential impacts of a major tsunami as part of wider Engineering lifelines Projects. The Canterbury study, undertaken in 1994, was concentrated on Christchurch City and Lyttelton, with the likely impacts extrapolated to other areas within the region. The Otago study, undertaken in 1997, attempted to identify potential problem areas on a site by site basis over the whole region.

While both studies involved developing scenarios on the magnitude of the maximum credible tsunami, there were several differences in the approach taken in each study. This paper sets the approach taken and results from each of the studies.

Christchurch Tsunami Study.

Scenario

Based on historical records, the most likely generating source for a significant tsunami effecting Christchurch is from a large seismic event centred on coastal South America (as occurred in 1868,1877 and 1960). The tsunami event adopted for the Engineering lifelines study was from this source. Local tsunamis from earthquakes or landslides on the Continental shelf were not considered to pose the same threat as a major far-field tsunami.

For a worst case scenario, the initial tsunami wave was set to coincide with a high spring tide when water levels in the Avon-Heathcote Estuary and Brooklands Lagoon (mouth of the Waimakariri River) would be at 1m above MSL.

The water levels adopted in the scenario involved a total water level variation at the open coast and in Lyttelton harbour of 10m inclusive of tide (tidal range of 2 m) For simplicity, the tsunami wave shape was assumed to be sinusoidal, with water levels reaching 5m above and below MSL. Based on the wave period of the 1960 tsunami at Lyttelton, a three hour period was used in the scenario, with minimum time from peak to trough of one hour for the first wave. Due to a falling tide the second wave would be 1m lower than the first, and the third wave coinciding with low tide being 2m lower than the first.

It was assumed that the maximum water levels would be reduced to 4m at the entrance to the Avon-Heathcote Estuary and the Waimakariri River due to dissipation of energy in the limited water depths on the ebb tide deltas. At the Sumner Esplanade from Scarborough to Cave Rock, the tsunami height was also assumed to be reduced to 4m due to shoaling in the shallow water at this location.

Sea conditions at the time of tsunami were assumed to be normal with swell heights of 1m, and a swell wave period of 8 seconds. River flow was also assumed to be average conditions.

The water levels in this scenario were considered a "best guess" estimate of the maximum credible tsunami. In comparison to other major tsunamis recorded at Lyttelton, the total water level variation in the scenario is 2.38 m and 4.26 m greater that in 1868 and 1960 respectively. Maximum water levels in the scenario are over 2 m higher than occurred in the recorded tsunamis, however since both of these events do not coincide with high tide, it is more meaningful to compare maximum water levels above predicted tide levels. This information is not available for the 1868 event, however, in 1960 the maximum was 2.9 m above predicted, compared with the scenario of 4 m above predicted.

Calculation of effects

The above tsunami water levels and swell conditions were used in wave breaker and run-up formula from CERC (1984) to calculate dune or beach structure heights which would be required to prevent inundation from the tsunami. The following results were obtained:

- 8m elevations above MSL for water level of 5m above MSL
- 7m elevations above MSL for water level of 4m above MSL
- 5m elevations above MSL for water level of 3m above MSL

From the Waimakariri River to the Estuary there are approximately 50 locations covering a total length of 2.5km where the dune or structure heights are below the 8m contour, hence where there is potential for sea water to overtop the beach during a tsunami event of the magnitude used in the scenario. The volume of water entering each of these locations was calculated based on the beach length and height of the opening, and the length of time run-up elevations were above the dune or structure elevation.

Water volumes entering the Waimakariri River and the Avon-Heathcote Estuary were calculated from the cross-sectional area of the mouth openings. It was calculated that during the initial 30 minute period, 8.9 mill m^3 could enter the Waimakariri mouth and 13.8 mill m^3 could enter the Avon Heathcote Estuary into already full water bodies due to tidal effects.

The water volumes at each breach were then mapped against the land topography behind the beach/estuary/lagoon to get potential inundation areas and depths. For the Avon and Heathcote

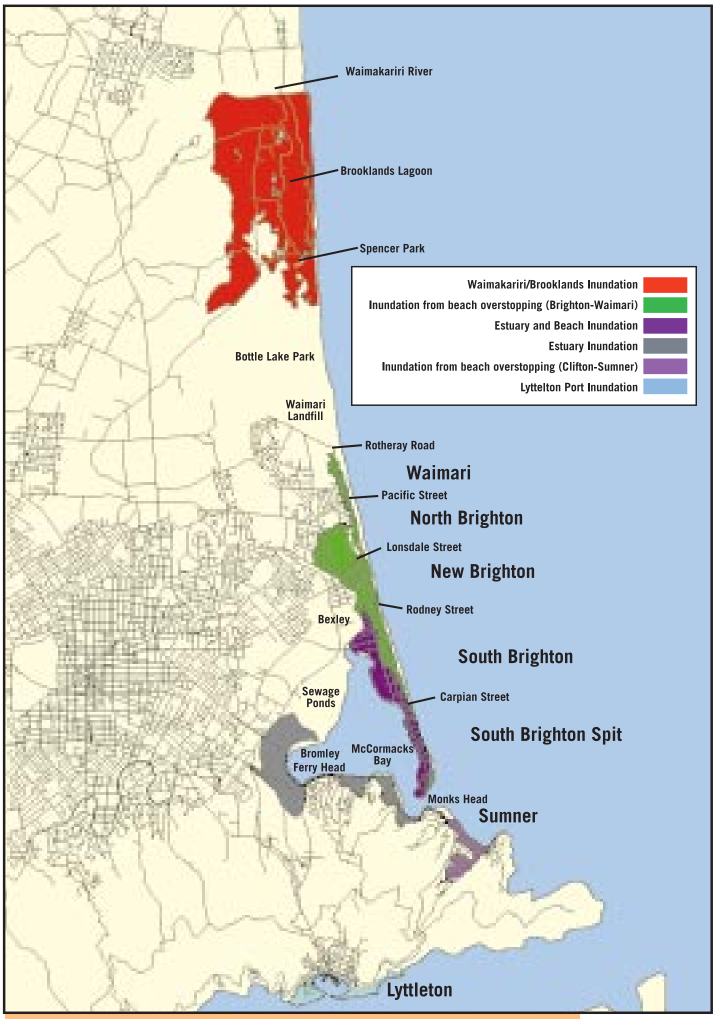


Figure 1: Potential inundation areas from Christchurch tsunami scenario

rivers, a MIKE 11 model was used to propagate the tsunami wave up the river channels.

Results

A total area of 1100 hectares could potentially be inundated, which could affect an estimated 11,000 residents. A map of these locations is shown in Figure 1, and breakdown of the effects in each area is as follows:

Waimakariri River mouth and Brookland Lagoon: Inundation 965 hectares land below 2 m contour. Water depth up 0.75 m.

Bottlelake Forest and Waimari Landfill: Isolated small blowouts, only limited inundation.

Waimari and North Brighton: Overtopping at the new subdivision, surf club and dune contouring areas. Inundation of 160 hectares to depth of 0.3 m

North Brighton: Inundation of 600 m of sea walls. Inundation of total area to max depth of 1 m. Water flowing to the Avon River.

South Brighton & Spit: Inundation of total area of 270 hectares to a maximum depth of 1 m.

Avon-Heathcote Estuary: Inundation around the Estuary at Moncks Bay, McCormicks Bay, Ferrymead, Bromley and South Brighton. Minor inundation at Bexley, but unlikely to affect the oxidation ponds.

Inundation possible on the Avon River up to 12 km upstream from the Estuary, and waves noticeable to the city centre.

Problems with scour of seawalls and bridges with strong velocities.

Sumner: Inundation of 70 hectares to depths of 0.7 m

Lyttelton: Inundation of 80 hectares of Port area below the 5 m contour to depths of 1-2 m. Not likely to effect Lyttelton township.

Otago Tsunami Study.

Scenario

As with Christchurch, historical records suggest that the most likely generating source for a significant tsunami affecting Otago is from a large seismic event centred on coastal South America. The tsunami event adopted for the Engineering lifelines study was from this source. The risk of a local tsunami from earthquakes on the Otago Continental shelf was assessed. The results showed that the frequency and magnitude of this type of tsunami was less than from a far-field source.

The water levels adopted in the scenario were based on a 20% increase in height from the highest recorded event (1868), adding 0.2 m for long term sea level rise and coinciding with a high spring tide. Conditions at the time of tsunami were assumed to be light seas and mean river flows.

The 'best guess estimate of the return period for this sized tsunami was calculated to be 1:350 years (25% probability of occurring in the next 100 years) based on the sequence of large South American earthquakes since 1513AD.

Calculation of Tsunami Water Levels

The historical review revealed that no previous event coincided with high tide, with the maximum recorded water level above predicted tide being 2.4 m at Oamaru in 1868. However, due to the sparsity of information from the 1868 event, it was necessary to establish a correlation of reported heights for the 1868 and 1960 events, and apply this to all sites of interest along the coast. The results revealed that heights in the 1868 were 75% higher that in 1960. Correlations were then established for sites with different orientations from the 1960 data. The results of this correlation showed that water level at south and east facing locations were 65% higher that north facing locations.

The analysis also showed that there was a general reduction in magnitude at river mouths, and a dampening of height as the tsunami wave travelled up Otago Harbour. The resulting water levels above mean high water spring were:

- 3.1 m for open coast orientated to the south and east
- 2.1 m for open coast orientated to the north
- 0.9 m for Dunedin

Levels at river mouths were scaled for orientation and mouth width relative to the Clutha River mouth.

Calculation of areas of potential impact

Areas along the Otago coast where significant hazards could potentially occur were identified, and the above tsunami water levels were applied to run-up formula from CERC (1984) for the beach profiles at these locations. The risk at each site was categorised as extreme, high, medium or low depending on the results of the run-up analysis. No attempt was made to map areas of potential inundation.

The results of this classification were as follows:

- Extreme Risk Areas (Direct inundation greater than 1 m deep) 3 sites:

Taieri river mouth (ground elevation only 2 m above msl, inundation depths up to 2 m), Tikoraki Point on the Moeraki Peninsula, and the road from Port Chalmers to Aramoana.

- High Risk Areas (Direct inundation less than 1 m) 15 sites

Includes Oamaru Harbour, Kakanui Beach, Moeraki wharf, Karitane Estuary & Beach, Purakanui Inlet, Aramoana, the Portobello Rd, Kaikorai Lagoon, Brighton Beach, both mouths of the Clutha River, and Kaka Point.

- Medium Risk Areas (Direct inundation possible, or estimated run-up greater than 1 m above ground elevation) 7 sites

Includes Portsmith Drive, Dunedin (due to effect of steep seawall slopes).

- Low Risk Areas (Estimated run-up less than 1 m above ground elevation) 4 sites.

Notable sites not considered at risk included Oamaru outside the harbour, Waikouaiti, Warrington and Dunedin beaches from Tomahawk to St. Clair beaches.

THE TSUNAMI WARNING SYSTEM IN THE PACIFIC -NEW ZEALAND'S PLACE

by Tom Finnimore
Ministry for Emergency Management

"Tsunami" is the Japanese term meaning wave in harbour. As such it is most descriptive of the observed phenomenon frequently referred to as tidal wave or seismic sea wave, with both of these terms having misleading connotations with respect to the mechanism of generation. In South America, the term "maremoto", or moving sea, is frequently used. However, the use of the word "tsunami" is most commonly accepted by scientists and by most of the lay public in the Pacific basin countries.

The most destructive Pacific-wide tsunami was generated along the coast of Chile on 22 May 1960. No accurate assessment of the damage and deaths attributable to this tsunami along the coast of Chile can be given. However, all coastal towns between the 36th and 44th parallels either were destroyed or heavily damaged. The combined tsunami and earthquake toll included 2 000 killed, 3 000 injured, 2 000 000 homeless and US$550 million damage. Waves in one coastal part were estimated to be 20 metres high. The tsunami caused 61 deaths in Hawaii, 20 in the Philippines and 100 or more in Japan. Estimated damages were US$50 million in Japan, US$24 million in Hawaii and several millions along the west coast of the United States and Canada. Wave heights varied from slight oscillations in some areas to 12 metres at Pitcairn Island; 10 metres in Hawaii, and 6 metres at various places in Japan. Along the east coast of New Zealand, wave heights varied from 1 to 5.5 metres damaging small boats and fishing vessels, sea walls and jettys, houses, footbridges, a hotel, as well as farm buildings.

In 1966, the Intergovernmental Oceanographic Commission (IOC) of UNESCO established the International Co-ordinating Group for the Tsunami Warning System in the Pacific (ITSU) whose main purpose is to recommend and co-ordinate programmes beneficial to those Pacific countries[1] whose coastal areas are potentially threatened by tsunamis. For this purpose, member states are encouraged to designate appropriate warning systems as well as stations to the International Tsunami Warning System in the Pacific, and to execute an educational programme on tsunamis and associated dangers.

The objective of the Tsunami Warning System is to provide accurate and reliable tsunami warning services within the shortest possible time frame of tsunamigenesis in order to protect life and property within the tsunami risk zones of the Pacific basin. A secondary objective is to adequately inform and educate all persons living within the tsunami risk zones so that they are properly prepared to respond to warnings when issued.

In order to provide a permament service for the Pacific, the IOC approved the USA's offer to establish the Pacific Tsunami Warning Centre (PTWC) at Honolulu, Hawaii. Its main responsibility being to issue warnings to all participants having designated civil defence organisations within 60 minutes of a tsunamigenic earthquake. The Centre is appropriately equipped with communications and computer facilities with scientific staff available on a 24 hour basis monitoring and, when required, issuing tsunami warnings to Pacific states.

PTWC collects and evaluates data provided by participating countries and issues appropriate bulletins to all Pacific nations, states or dependencies regarding the occurrence of a major earthquake and possible or confirmed tsunami. There are three categories of bulletins.

A tsunami warning/watch bulletin is issued based on earthquake location and magnitude, generally exceeding 7.5 on the Richter scale. The area within three hours tsunami travel time of the epicentre is placed on a tsunami warning status, with the area within a 3-6 hour travel time zone placed on watch status. The estimated arrival time of the first wave is disseminated for the tide stations within the warning/watch areas. Subsequent bulletins are issued on an hourly basis until cancelled when the PTWC determines that a threat no longer exists.

A tsunami information bulletin is issued by the PTWC when a major earthquake occurs in a coastal or near-coastal location, or within the Pacific basin with a magnitude of 6.5 to 7.5 on the Richter scale with the evaluation that a tsunami was not generated. If the PTWC's evaluation that tsunami generation is possible, the nearest tide stations will be monitored.

There are two communications links from the PTWC to the Ministry for Emergency Management. The prime one is via the Aeronautical Fixed Telecommunications Network through NZ Airways Corp to the Ministry (24 hours) and the backup, via the Global Telecommunications System through the NZ MetService to Ministry (24 hours).

For New Zealand, procedures dealing with tsunami warnings issued from the PTWC are well documented and are based on Part Three of the National Civil Defence Plan. Basically the procedure is described below.

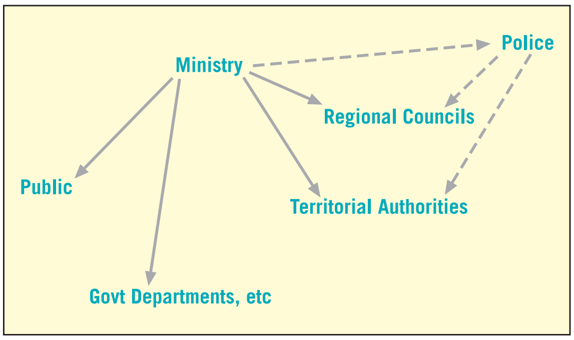
A warning is followed up by a phone call from PTWC to the Ministry for Emergency Management.

• the Ministry's response on receipt of watch/ warning is to issue a warning direct to:

• regional councils and territorial authorities
• the general public
• Minister of Civil Defence
• Department of Prime Minister and Cabinet
• New Zealand Police
• New Zealand Defence Force
• New Zealand Fire Service
• key departments and agencies
• the media - radio, press and television

During the event, the Director Emergency Management will confer with the appropriate accredited scientific advisors. The National Emergency Operations Centre maintains a link with the PTWC and New Zealand local authorities.

Processing a watch/warning from the PTWC is illustrated as follows:



• Whereas Police National Headquarters is advised as a Government Department, the Police Districts check that local authority Civil Defence organisations have received the message.

• Ministry staff in Auckland and Christchurch check with regions to ensure that they have received the message, if not the warning message is conveyed orally.

• Regions check that territorial authorities have received the message.

•Information on the expected arrival time of the initial wave is contained in the watch and warning message as well as information on wave height if known.

Local Responsibilities

The National Civil Defence Plan asks territorial authorities and regional councils to incorporate warning procedures in their respective civil defence plans. This also includes the task of maintaining contact lists.

Testing the Warning System

Tests of the New Zealand National Warning System are conducted quarterly. Results overall are satisfactory with the message getting to 90% of over 100 recipients within 15 minutes and to all within 45-60 minutes.

Communication system tests are conducted periodically by the PTWC. This includes conducting communication checks with tide recorder stations both at Lyttelton and Marsden Point. The automatic data recorder at the Chatham Islands is constantly monitored and can be interrogated from Hawaii.

[1] Member states are: Canada, USA, Mexico, Guatemala, Nicaragua, Costa Rica, Columbia, Ecuador, Peru, Chile, Cook Islands, Tahiti (French Polynesia), New Zealand, New Caledonia, Fiji, Western Samoa, Australia, Indonesia, Singapore, Thailand, Philippines, People's Republic of China, Republic of Korea, Democratic People's Republic of Korea, Japan and the Russian Federation.