

5.0 DEFINING TSUNAMI SOURCES

This section summarises known information about the possible sources of tsunami that could cause damage in New Zealand. For the purposes of emergency management and the time needed to respond and act on warnings, tsunami may be categorised as distant, regional or local source, depending on the shortest travel time of the tsunami from its source to the closest part of the New Zealand coastline. The travel time is generally consistent with the location of the sources, in that distant sources for New Zealand are mainly around the Pacific rim, while local sources relate to the New Zealand ‘continent’. The categories that we adopt for this report are:

- Distant source – more than 3 hours travel time from New Zealand
- Regional source – 1–3 hours travel time from New Zealand
- Local source – 0–60 minutes travel time to the nearest New Zealand coast (most sources are <30 minutes travel time)

It should be noted that a local source tsunami, which may strike the nearest shore within 60 minutes, may take more than sixty minutes to travel to other New Zealand locations. This affects the time available for Emergency Management to issue a warning and so needs to be kept in mind when warning systems are being considered.

The definition of a regional source in terms of travel time sometimes causes confusion when discussing tsunami sources occurring in the southwest Pacific ‘region’, as a tsunami from for example, the Solomon Islands, may take more than three hours to reach New Zealand. To minimise confusion we recommend explicitly using the words ‘SW Pacific region’ if classifying tsunami sources according to their location rather than their travel time.

5.1 DISTANT SOURCES

5.1.1 Earthquakes

Large to Great ($M > 8$) earthquakes are the most frequently-occurring source of damaging tsunami worldwide, and 80% of these earthquakes occur around the margins of the Pacific Ocean. At many of the plate boundaries in the circum-Pacific one tectonic plate is diving down beneath another tectonic plate, in a process called subduction (Figure 5.1). Often, the subducting and overriding plates can become stuck together due to friction on the boundary between the plates. Eventually the stored energy due to this “locking” process overcomes the strength of the plate boundary, and the two plates suddenly slip past each other in a large earthquake. The 2004 Indian Ocean tsunami was generated by this process in the Indian Ocean, where the Australian plate is subducted beneath the Eurasian plate along the Sumatran subduction zone. The Tohoku (Japan) 2011 tsunami was similarly generated on a plate boundary where the Pacific plate is subducted beneath northern Japan.

The potential of subduction zones to produce tsunami at the New Zealand coast has been assessed from the available data, including historical occurrences, numerical modelling and literature on earthquake recurrence and magnitude. The evaluation revealed that only sources in the circum-Pacific region (including New Zealand’s subduction zones and some offshore faults) are likely to generate tsunami with heights of > 2 m (Figure 5.1). Tsunami have been recorded along the New Zealand coast from other sources (for example the 2004 Indian Ocean tsunami from the Indian Ocean), but these are not expected to exceed 2 m in

the maximum 2500 year return period considered in this study. The characteristics of the tsunami source areas used in the probabilistic model in Chapter 6 have been based on up-to-date review literature of the Pacific Rim regions, and include input from Global Earthquake Model (GEM) workshops and consultation with international experts. A complete compilation of the literature is beyond the scope of this review.

The 2011 Tohoku earthquake demonstrated that there may be very long intervals of time between the largest earthquakes on a subduction zone (the most recent earthquake of similar size in Tohoku was in 869 AD). Consequently, the ability to infer a maximum potential earthquake size from historical data is very limited in the Pacific where, apart from Japan, records of historical events are typically only a few hundred years long. Attempts to infer maximum earthquake sizes from geophysical properties of the converging plates have also been hampered by the short span of records, and several factors that have previously been proposed as controlling maximum earthquake size have been contradicted by recent events. From a civil defence standpoint it should be assumed that any subduction zone could produce a magnitude 9 earthquake, unless there are strongly convincing counter-arguments. For hazard and risk assessments, the uncertainty in estimates of maximum magnitude should be reflected in the hazard modelling inputs (i.e., logic-tree weights, or epistemic uncertainty distributions) for the source regions.

5.1.1.1 South America

The west coast of South America (Figure 5.1) is one of the most frequent sources of tsunami in the Pacific, resulting from great earthquakes on the boundary where the Nazca, Cocos, and Antarctic plates subduct beneath the South American tectonic plate. Earthquakes along this coastline produce tsunami that are often directed towards New Zealand, due to both the orientation of the plate boundary on which the earthquakes occur, and to a lesser degree from refraction of tsunami waves by undersea ridges and other bathymetric features between South America and New Zealand. There are also very few island chains between New Zealand and South America to block or scatter the tsunami waves.

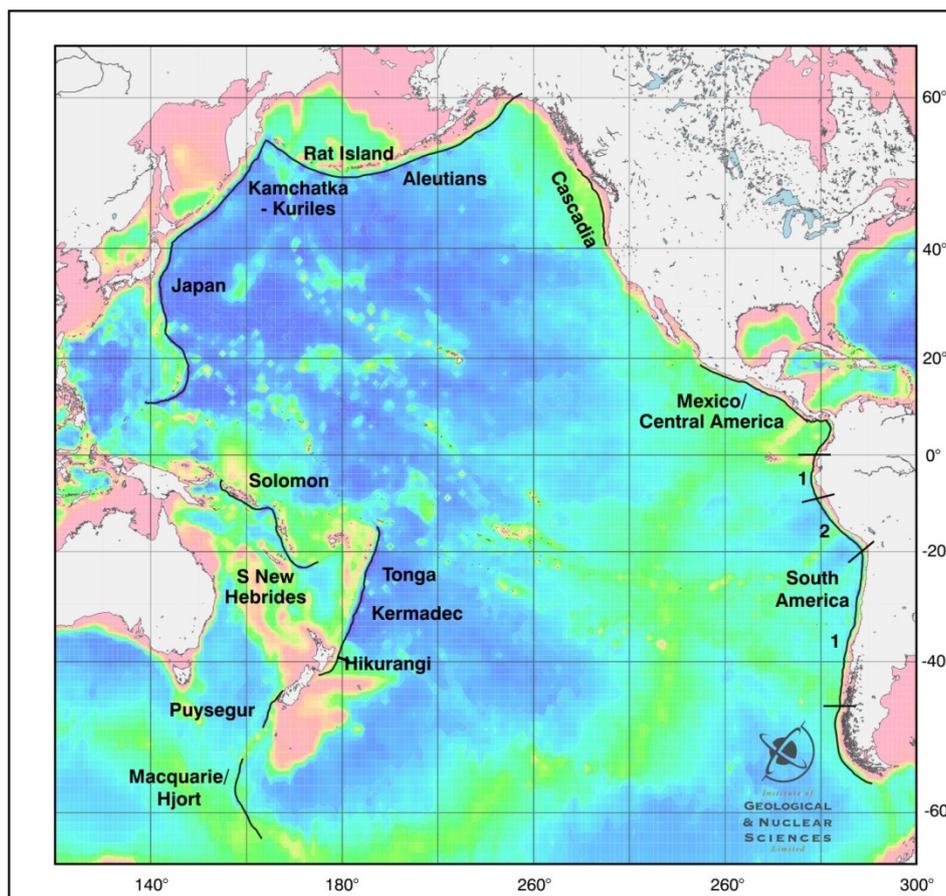


Figure 5.1 Subduction margins in the circum-Pacific region. The South American coastal margin can be partitioned into regions that propagate tsunami either south-westward toward eastern New Zealand (region 2), or direct tsunami further northward, thus more strongly affecting the north Pacific (region 1). The 1868 tsunami was generated in region 2, while the larger but less damaging (in New Zealand) 1960 tsunami originated in region 1.

For example, the distant-source tsunami that caused the most damage to New Zealand in historical times was caused by an earthquake on the southern coast of Peru in 1868 (region 2 of Figure 5.1; see also Section 3.2); the magnitude of this earthquake is estimated to be about $M_W 9.1$. This area of South America's coastline is oriented in such a way that the tsunami energy is more effectively directed towards New Zealand than it is for other sections of the South America coast, where the majority of the energy tends to be directed further north. Should a larger earthquake than the 1868 event occur along this part of the coast, the effects in New Zealand could be considerably greater than occurred in 1868.

In contrast, the 1960 tsunami, although caused by a much larger earthquake ($M_W 9.4$, possibly $M_W 9.5$; see Section 3.2), occurred on a part of the South American plate boundary that is not as well oriented to New Zealand as the 1868 location (southernmost region 1 of Figure 5.1). It produced a smaller tsunami in New Zealand than would have occurred had the source location been better oriented. Nevertheless, the 1960 tsunami caused run-ups of up to 4 m in parts of the North and South Islands. The 2010 Chilean tsunami, caused by an earthquake of magnitude $M_W 8.8$ that occurred to the north of the location of the 1960 earthquake, had run-up heights of up to 1 m in parts of the North and South Islands.

The magnitude of the 1960 earthquake, at $M_W 9.4-9.5$, is probably close to the upper limit for earthquakes for the whole South American coastline (and worldwide). It is uncertain whether other portions of the South American subduction margin are capable of producing earthquakes of this size, or whether most segments have more frequent but relatively smaller earthquakes of $M_W 8-9$. When the earthquake catalogue for the whole South American

margin is analysed, the frequency of large earthquakes appears to decrease with increasing magnitude—for every unit increase in magnitude the frequency of earthquakes drops by about a factor of ten.

Computer models (Power et al., 2004), combined with historical observations, suggest that there is minimal risk of a damaging tsunami in New Zealand generated by South American earthquakes with magnitudes of less than 8.5.

The historical records of Peru and Chile, which are hundreds of years longer than New Zealand's, indicates that large earthquakes and tsunami have occurred relatively frequently in the last 450 years (Table 5.1). From the table it can be seen that over 450 years there were nine earthquakes with magnitudes of $M_w > 8.5$ that caused run-up heights near the earthquake source that were similar to or greater than those produced by the 1868 event. Those earthquakes prior to 1868 probably produced significant tsunami in New Zealand prior to European settlement. The average return period (50 years) is about the same as has occurred in the last 160 years, and indicates the frequency of potentially damaging tsunami from South American sources in New Zealand.

Paleotsunami work conducted by Cisternas et al. (2005) suggests that tsunami comparable in size to the 1960 event occur in southern Chile with an average interval of slightly less than 300 years.

As tsunami from South America approach New Zealand from the east, the east coast will be more affected than the west coast. However, waves do propagate around New Zealand, as well as through Cook Strait, and the west coast will have significant waves in some cases.

Table 5.1 Large South American earthquakes that have produced tsunami with maximum tsunami heights greater than 8 m locally (extracted from Gusiakov, 2001; additional information on the 2001 and 2010 earthquakes from the NGDC tsunami database). Events shown in bold are either known to have caused, or had the potential to have caused a significant impact in New Zealand comparable to the 1868, 1877 and 1960 tsunami. The magnitudes for early events (shown by grey shading) may have large errors. Note: M_b or M_s – body wave or surface magnitude; M_w –moment magnitude; M_t –tsunami magnitude [Abbreviations: S = south; N = north;].

Year	Mth	Day	Lat. (°N)	Long. (°E)	M_b or M_s	M_w	M_t	Max. run-up at source (m)	Source	Max run-up in NZ (m)
1562	10	28	-38.70	-73.20	8.0			16	S. Central Chile	
1586	7	9	-12.20	-77.70	8.5			26	Off Lima, Peru	
1604	11	24	-18.50	-70.35	8.4			16	Africa, N. Chile	
1657	3	15	-36.80	-73.00	8.0			8	Conception, S. Chile	
1687	10	20	-13.50	-76.50	8.5			8	Callao, Lima, Peru	
1730	7	8	-32.50	-71.50	8.7			16	Valparaiso, Chile	
1746	10	29	-12.50	-77.00	8.0	8.6	9.2	24	Callao, Lima, Peru	
1806	12	1	-12.10	-77.10	7.5			6	Peru	
1835	2	20	-36.50	-72.60	8.5			14	Conception, S. Chile	
1837	11	7	-42.50	-74.00	8.5		9.2	8	Corral, S Chile	
1859	10	5	-27.00	-70.40	7.7			6	Caldera, Chile	
1868	8	13	-17.70	-71.60	8.8	9.1	9.0	18	Arica, S. Peru	4; (10 Chatham Islands)
1877	5	10	-21.06	-70.25	8.8	9.0		21	Iquique, N. Chile	~3.5
1922	11	11	-28.31	-70.28	8.3	8.7		9	Caldera, Chile	~1
1929	8	9	-23.60	-70.40				8	N. Chile	
1960	5	22	-38.31	-72.65	8.6	9.5	9.4	25	S. Chile	~4
1960	11	20	-6.64	-80.55	6.9	7.7	7.7	9	N. Peru	
1996	2	21	-9.71	-79.86	6.6/7.5	7.8	7.8	5	Peru	
2001	6	23	-16.26	-73.64	6.7/8.2	8.4	8.2	7.0	Camana, S. Peru	
2010	2	27	-36.122	-72.898	7.2/8.5	8.8	8.8	29.0	Conception, S. Chile	~1

5.1.1.2 Mexico and Central America

The largest well-recorded historical earthquakes in the Mexico and Central America area (Figure 5.1) have magnitudes of less than M_W 8.5, too small generally to produce a damaging Pacific-wide tsunami. However an earthquake in 1787 is estimated at $\sim M_W$ 8.6 (Suarez and Albin, 2009), and the possibility of still larger earthquakes cannot be excluded (Hjorleifsdottir et al., 2012). The coastline of this region is well oriented for directing tsunami towards New Zealand, and the possibility of earthquakes that produce tsunami large enough to be damaging in New Zealand cannot be ruled out.

5.1.1.3 Cascadia

The Cascadia margin refers to the boundary between the Juan De Fuca and North American tectonic plates between northern California and Vancouver Island (Figure 5.1). The Cascadia plate interface has an extensive paleoseismic record based on analysis of turbidites (Goldfinger et al., 2012). The turbidite record suggests 40 large earthquakes in the past 10,000 years, 19 of which are thought to have been whole-region ruptures of about magnitude 9.0, and the remainder are thought to be segmented ruptures with typical magnitudes of 8.2-8.6. Between these large events the Cascadia plate interface appears to be relatively seismically inactive.

The last great Cascadia earthquake occurred in 1700 AD; it was identified from historical tsunami records in Japan, and is consistent with geological evidence from the United States and Canada (Atwater et al., 2005). This date is earlier than written records in New Zealand, as it is in the United States and Canada, and the only means to estimate likely impact in New Zealand is by using numerical modelling. Japanese researchers have estimated the magnitude of the 1700 AD event at M_W 9.0. An earthquake of this magnitude is expected to produce a tsunami with amplitudes up to about 1 m in many parts of New Zealand, and possibly 1–3 m in Banks Peninsula and the Chatham Islands. The dimensions of the Cascadia subduction zone appear to set an upper limit on the magnitude of possible earthquakes at about M_W 9.2; at this magnitude many areas of New Zealand would be expected to experience waves with amplitudes between 1–3 m.

5.1.1.4 Alaska and the Aleutians

The plate boundary between Alaska and the Aleutians (Figure 5.1) is a highly active source of great ($M_W > 8.0$) earthquakes and tsunami in the Pacific. Historically, three earthquakes—the 1964 M_W 9.4 Alaska, the 1957 M_W 8.7–9.1 Rat Island, and the 1946 M_W 7.9 Aleutian earthquakes—have caused run-ups of up to 2 m along the north and east coasts of New Zealand, but not at any of the urban centres.

The historical record here is too short to reflect the full range of tsunami that New Zealand might experience from the Alaskan and Aleutians region. However, most parts of the coastline produce tsunami that are not particularly well directed to New Zealand, with exception of the area around the source zone of the 1957 Rat Island earthquake.

5.1.1.5 Kurile Islands, Kamchatka

The largest earthquake in the Kurile Islands-Kamchatka area (Figure 5.1) to produce a tsunami recorded in New Zealand is an M_W 9 earthquake south of Kamchatka Peninsula in 1952. Near the earthquake source this event produced a maximum run-up of nearly 19 m,

with a maximum run-up in New Zealand of over 1 m in Gisborne. A larger tsunami, with a maximum run-up of 63 m locally and 15 m at a distance of over 1000 km away, was recorded in 1737 from an earthquake with an estimated magnitude of M8.3. Its effects in the larger Pacific area are unknown. The capacity of the area to produce earthquakes with magnitudes greater than the M9.0 in the historical record is unknown.

5.1.1.6 Japan

The subduction zones off Japan (Figure 5.1) are some of the most active in the Pacific. The region also has one of the longest historical records of large earthquakes and tsunami, spanning over a thousand years. Until recently, no earthquakes offshore Japan were thought to have reached magnitude 9, although there are many historical events over magnitude 8. In New Zealand's historical record prior to 2011, only very small waves with amplitudes of less than a metre were recorded from Japanese earthquakes. The largest historical subduction thrust earthquake in Japan was of M_w 9.0 rupturing much of the northern Japan Trench in March 2011 (see also Section 3.3). This event produced tsunami with runup heights of as much as 35 m at locations along the northeast Japan coastline, with inundation of areas up to about 5 km inland from the coast on the Sendai plain, and significantly further along the banks of rivers. In New Zealand wave amplitudes of up to about 1 m were recorded at various tide gauges (Borrero et al., 2012). A marine threat warning was issued throughout New Zealand for this event, and anomalous waves and currents related to the tsunami were observed for several days after the earthquake. There was some flooding of residential houses at the head of the bay at Port Charles in the Coromandel. We expect that earthquakes on Japan's subduction boundaries much larger than the March 2011 event are rare (i.e. with recurrence intervals of several thousands of years, if they occur at all), although events up to M_w 9.5 cannot be entirely ruled out. The worst case M_w 9.5 scenario would lead to wave heights in New Zealand approximately 2–3 times the ones observed from the March 2011 event. Fortunately, the propagation path from Japan to New Zealand is studded with islands that are thought to protect New Zealand from wave amplitudes of more than 2–3 m, even in this worst case.

5.1.1.7 Solomon Islands, Papua New Guinea

Historically, the Solomon Islands and Papua New Guinea (Figure 5.1) have produced few earthquakes over magnitude 8.5, although we cannot rule out this possibility. In Papua New Guinea, the primary source that could affect New Zealand would be the New Britain Trench. However, the PNG mainland and numerous islands located between the New Britain Trench and New Zealand would scatter the waves. However, great earthquakes on the San Cristobal Trench just to the southwest of the Solomon Islands could pose a significant tsunami hazard to New Zealand. Modelling has shown that the Lord Howe Rise behaves as a waveguide, steering tsunami waves from the Coral Sea region towards New Zealand. Historically, few tsunami from this region have produced tsunami heights exceeding 1–2 m at a large distance from the source. The M_w 8.1 earthquake in the Solomon Islands on 2 April 2007 produced a tsunami with a maximum reported run-up height of 12 m in the Solomons and killed 52 people; the largest recorded waves in New Zealand from this event were just over half a metre in amplitude.

5.1.1.8 Northern New Hebrides

The northern part of the New Hebrides subduction zone is a distant source, while the central and southern parts (see Section 5.2.1.1) are regional sources, using our criteria based on travel-time. The possibility exists that a very large earthquake could rupture multiple fault

segments. Near Vanuatu in the central part of the New Hebrides region, large earthquakes with magnitudes of 8.5 or less have created tsunami with run-ups of 12 m locally. The northern part of the New Hebrides subduction zone is not well oriented to direct tsunami towards New Zealand, although modelling suggests that undersea ridges in the Tasman Sea will direct some of the tsunami energy towards New Zealand.

At the New Hebrides Trench, the Australian plate is moving northeast and being subducted beneath Vanuatu and the adjacent area containing a complex series of rifts and transforms in the North Fiji Basin. Plate movement velocities determined by GPS indicate the plates are converging at the New Hebrides Trench at rates of 4–16 cm/year (Calmant et al, 2003; Power et al, 2012). The GPS data in Vanuatu suggest that a large portion of the plate interface of the New Hebrides subduction zone is strongly locked together between earthquakes (Power et al., 2012), which will presumably result in a major subduction thrust earthquake in that region when the accumulated plate boundary strain is eventually released.

There have been dozens of earthquakes ($> M_W 7.5$) in the New Hebrides Trench region over the last century, including two estimated $M_W 8.0$ events in 1878 and 1920. However, we cannot rule out the possibility of rupture of the northern New Hebrides Trench over a larger area, causing an $M_W 9$ or larger earthquake (Wright et al., 2011).

5.1.1.9 Summary comments

Few areas can, with certainty, be excluded as a source of damaging tsunami until all earthquake sources are considered and numerical modelling has revealed the extent, or lack, of a threat. For global distances, there are significant uncertainties about the potential for tsunami from sources in northern South America, Cascadia (western USA), Mexico, Central America, Alaska and the Aleutians. The 2004 Sumatra and 2011 Tohoku (Japan) tsunami have, however, changed the perspective regarding the potential for earthquakes of magnitude 9 or above on many subduction zones. It has become apparent that many conclusions were incorrectly drawn because they were based on historical records that were in fact too short to include the largest earthquakes which occur infrequently. At this time the main plausible way to determine a limit on the maximum size of earthquake that a subduction zone could experience is from the total length of the subduction zone, i.e., subduction zones cannot experience earthquakes larger than a cut-off magnitude determined by their length (McCaffrey, 2007); this cut-off is lower for shorter subduction zones.

5.1.2 Landslides

The role of submarine landslides and their potential to produce local, regional and Pacific-wide tsunamis has undergone critical international scientific review and debate in recent years, particularly as a result of a devastating tsunami in 1998 in Papua New Guinea. Some scientists have attributed this larger-than-expected tsunami to the magnitude and seismic characteristics of the generating earthquake, others to the occurrence of an offshore landslide a few minutes after the earthquake (Geist, 2000; Tappin et al., 2001). This has led many tsunami researchers to recognise that submarine landslides may play a greater part in generating local tsunami than previously thought. Submarine landslides have also been argued to have added substantially to the trans-Pacific tsunami resulting from a 1946 earthquake in the Aleutians (Fryer et al., 2004). They argue that the narrow “beam” of devastating tsunami that swept Hawaii and the Marquesas Islands, and had a run-up of 4 m in Antarctica, was the result of a 200 km³ landslide triggered by the $M_W 7.9$ subduction earthquake. Others (e.g., Tanioka and Seno, 2001) have suggested the earthquake had very

large slip for its apparent magnitude, such that it would fall into the category of so-called “tsunami earthquakes”.

Modelling indicates that huge sector collapses (1000–5000 km³) of the flanks of the Hawaiian volcano chain could produce Pacific-wide tsunami, as well as very large local tsunami of hundreds of metres (McMurtry et al., 2004). While it is likely that flank collapses of this scale would produce large tsunami in New Zealand, their return periods from any one source are well in excess of the 2500-year return period covered in this study. Therefore, no landslides at global distances are considered viable tsunami sources within the 2500 year period.

5.1.3 Volcanoes

Other than the potential for flank collapse on the slopes of volcanoes, no volcanoes in the historical record are known to have directly produced significant tsunami at great distances. The great 1883 eruption of Krakatau, Indonesia, produced huge local tsunami with some run-up heights exceeding 40 m, but tsunami-like water level oscillations observed at great distances from the volcano have been attributed to a coupling of an atmospheric pressure wave with the ocean. These waves, given the name *rissaga*, or atmospheric tsunami, are outside the scope of this review. Not enough is known about their mechanisms to include them as a tsunami source for this review. Nevertheless, oscillations in New Zealand following the Krakatau eruption included 1.8 m (measured peak to trough) waves at Whitianga and in the anchorage area at Auckland (although only 0.9–1.2 m at the Auckland docks) (de Lange and Healy, 1986).

5.1.4 Bolide impact

As an island nation surrounded by a large deep sea, New Zealand has a tsunami hazard from impacts of asteroids and comets. This hazard is real, finite and determinable, but the probability of a damaging tsunami from bolide impact is low. One such large event is known to have occurred on Earth within recorded human history—a meteor exploded over Constantinople on a clear afternoon in 472 AD, hitting the city with a wave that knocked sailboats flat in the water.

Asteroids and comets are collectively known as Near Earth Objects (NEOs) when they approach close to Earth, especially if their closest approach is less than the distance to the moon. If they enter the Earth’s atmosphere, they are collectively called bolides. The visible track of a bolide across the sky is a meteor, or shooting star. The solid objects that sometimes are recovered later are meteorites. A meteorite survives its passage through the atmosphere and hits Earth about once every two hours.

Current technology allows us to detect and track the larger NEOs (larger than a few metres in diameter) and calculate their probability of hitting Earth, days, weeks, and sometimes months in advance of their closest approach. The larger the body, the further out it can be identified and tracked. At any time, there are always some NEOs, and many approaching. (A current list of NEOs can be viewed at <http://neo.jpl.nasa.gov>, and is updated at least daily). If a NEO large enough to be of concern were likely to hit the Earth, substantial advance warning would be given; in fact several warnings have been made public before very near misses. All significant objects on a collision course can be tracked, and their likely impact site on Earth predicted, with known uncertainty, some substantial time in advance of impact. Large bolides, however, have never been so common that they have featured prominently in human history.

Numerical estimates of the frequency of impact of a meteorite of sufficient size within a distance range of New Zealand that could cause a damaging tsunami appear to have a recurrence interval many times longer than the 2500 years considered in this project (see Appendix 2 for details of the calculation). This estimate of long recurrence intervals for meteorite generation of damaging tsunami is consistent with their scarcity in human records. Because of the apparent long return period for a damaging tsunami generated by meteorite to affect New Zealand we do not consider this source further in our tsunami source descriptions.

While most bolide impacts occur as distant sources of tsunami to New Zealand, they may also occur at regional or local distances.

5.2 REGIONAL SOURCES

The warning time for tsunami from regional sources is about 1–3 hours, and presents a real challenge to monitoring and warning agencies. To locate an event, evaluate its tsunami potential and issue a warning in so short a time is problematic, requiring pre-planning and scenario development. Self-evacuation of residents will be required at short notice. As outlined in the following sections, regional source tsunami may represent a significant hazard and risk, and these may be catastrophic on rare occasions.

Regional sources include earthquakes and volcanoes (eruption and flank collapse) from tectonically active regions to the north of New Zealand, and south of New Zealand from about 50-60°S. Regional sources of tsunami to the east and west are highly unlikely (but note that the Solomon Islands subduction zone is a *distant* source that primarily affects the west coast of New Zealand, even though it occurs in the 'SW Pacific region'). Hence, the coasts most at risk from regional source tsunami are the northern half of the North Island and the southern half of the South Island.

The following sections outline what is known about the historical impact of regional source tsunami, about the sources of potentially damaging tsunami, and what has been learnt and can be learnt from numerical modelling, and from geological studies of pre-historical tsunami. They form the basis for what is known about the frequency and magnitude of events that New Zealand might expect to experience.

5.2.1 Earthquakes

In New Zealand's historical record, the largest earthquakes along the arc between New Hebrides (Vanuatu), Kermadec Islands and Tonga have been less than magnitude 8.5. Only two of these are known to have caused tsunami with run-ups in New Zealand approaching 1 m. Although the record of run-ups in New Zealand may be incomplete, we would expect a large event in historical times to have been noted.

To the south of New Zealand, only a few large earthquakes have occurred since the 1960s, when the installation of a worldwide seismic network allowed large earthquakes to be identified and located. The only three large earthquakes in the last 40 years had magnitudes between 7.8 and 8.4, and all were in areas of the plate boundary where earthquakes with predominantly horizontal (strike-slip) movement along the fault occur. These earthquakes do not usually generate large tsunami and none had run-up of > 1 m in New Zealand (along the south and west coasts of the South Island).

In this section we address the potential of each subduction zone at regional distances, to generate tsunami that could produce tsunami heights of 2 m or more in New Zealand, and within the 2500 year return period considered in this project.

The 2011 Tohoku earthquake demonstrated that there may be very long intervals of time between the largest earthquakes on a subduction zone (the most recent earthquake of similar size in Tohoku was in 869 AD). Consequently the ability to infer a maximum potential earthquake size from 100–200 years of historical data is very limited.

5.2.1.1 Southern New Hebrides

The central and southern parts of the New Hebrides subduction zone are regional sources according to our criteria based on travel-time, while the northern part is a distant source (see 5.1.1.8). Large earthquakes with magnitude up to 8.5, causing tsunami with run-ups of 12 m locally, have occurred near Vanuatu in the central part of the New Hebrides region. The central part of the New Hebrides subduction zone is not well oriented to direct tsunami towards New Zealand. The southern part is well oriented, but here the record of earthquakes is probably complete only from 1960 onward.

The historical record of earthquakes in the Southern New Hebrides is short, and complete for major earthquakes ($M_W > 7.5$) only over the past century. This is primarily a consequence of their remote location, the largely uninhabited nature of the few islands in the eastern part of the arc, and the lack of surviving oral history accounts from the pre-colonisation cultures on the islands. The largest historical event occurred in August 1901, with an estimated magnitude of 7.9–8.4, and several lives were lost on islands close to the earthquake due to the tsunami (see summary in Power et al., 2012). The observed maximum water heights in Hawaii from this tsunami were up to 1.2 metres, and were significantly larger than those of the 2009 M_W 8.1 Samoa earthquake. On this basis, Power et al. (2012) favour the interpretation of a moment magnitude for this event of around 8.4.

Power et al. (2012) show that earthquakes larger than an M_W 8.0 on the southern section of the New Hebrides trench (Figure 5.1) could present a significant hazard for Northland. An under-sea ridge (the Three Kings Ridge) extends north from Cape Reinga and acts as a waveguide (see the discussion of waveguide effects in Section 4.2.2.2), leading to potentially hazardous wave heights in the northern North Island. Numerical models show that earthquake scenarios ranging from M_W 8.15 to M_W 8.8 on the southern New Hebrides Trench could lead to maximum tsunami heights of 2 to 15 m respectively at highly amplifying sites in the far north, such as the Aupouri Peninsula. Along the eastern and western coastlines of Northland, maximum expected tsunami heights range from less than 1 to ~5 metres as the magnitude ranges from M_W 8.15 to M_W 8.8.

5.2.1.2 Tonga Trench

Historically, earthquakes have not exceeded magnitude 8.5 in the Tonga Trench (between 14–26°S), and the tsunami have not significantly affected New Zealand, principally because of the orientation of the subduction margin. It has been thought that the potential for regional scale tsunami was limited, as the plate interface appears to be uncoupled, so the plates slide past each other relatively freely (Bevis et al., 1995), but the September 2009 Samoa tsunami (also called the ‘South Pacific tsunami’; see Section 3.3) was found to be caused by two near simultaneous earthquakes—one on the subduction interface and the other on an outer rise normal fault (Beaven et al., 2010; Lay et al., 2010), so this assumption is open to question. If the magnitude of earthquakes on the Tonga Trench were to be limited only by subduction

zone length, then earthquakes with magnitudes up to the low 9s would be possible; yet this still appears to be unlikely based on the lack of coupling revealed by current geodetic measurements (though these measurements are not sufficient to fully reveal the extent of coupling, especially close to the trench).

5.2.1.3 Kermadec Trench

The Kermadec Trench is both a regional tsunami source and a local source. Earthquakes caused by rupture along only the northern half of the Kermadec Trench could produce tsunami that take more than an hour to reach New Zealand (excluding the Kermadec Islands), while rupture of the Kermadec Trench south of $\sim 33^\circ\text{S}$ could produce tsunami that reach some parts of New Zealand in less than an hour. Because of the orientation of the Kermadec Trench, even for tsunami originating on the southern part of the trench, the most strongly affected coasts are usually more than one hour travel time away.

The c. 1400 km long Kermadec Trench has a moderate level of historical seismic activity. In the 33 years from 1976 to 2009, 544 earthquakes with a magnitude greater than 5 originated on the shallow part (≤ 40 km depth) of the plate interface (based on thrust mechanisms) (Power et al., 2012). Power et al. (2012) used the seismicity data, and the available geodetic data, to assess the potential for large tsunamigenic earthquakes, and modelled the consequences of such events.

There have been three significant earthquakes on the southern half of the Kermadec Trench (e.g., south of 29°S) since the beginning of the 20th century, namely the earthquakes of 2 May 1917 ($M_{8-8.6}$), 14 Jan 1976 ($M_{7.8-8}$), and 20 October 1986 ($M_{7.9}$). The 1976 earthquake occurred near Raoul Island and was likely a low-angle thrust earthquake along the main subduction plate interface. This earthquake caused a tsunami that was widely observed in the Pacific, primarily on tide gauges, but also as observed run-ups. Several yachts were damaged in Tutukaka harbour near Whangarei in New Zealand (Bay of Plenty Times, 15 January 1976; Downes, pers. comm.). The 1986 earthquake is consistent with normal faulting within the subducting Pacific Plate (see summary in Power et al., 2012), and only very small (< 10 cm) water height changes were observed at tide gauges in Hawaii, French Polynesia and Samoa.

The ability of the Kermadec Trench to produce earthquakes larger than about $M_{\text{W}} 8.5$ is not well known. Comparisons with other subduction margins similar to the Kermadecs (Mariana, for example) would lead us to suspect that the plate interface is relatively weakly coupled, i.e., the plates are sliding past each other relatively freely and without building up elastic energy that is then released as earthquakes. However, analysis of GPS data from Raoul Island (the only existing GPS site on the Trench) indicates that the plate interface is probably strongly coupled or locked, at least in that part of the Trench (Power et al., 2012).

Power et al. (2012) conducted numerical modelling of four plausible scenarios for Kermadec Trench rupture, ranging from earthquakes of $M_{\text{W}} 8.5$ (restricted to the far southern Kermadec Trench) to $M_{\text{W}} 9.4$ (rupture of the entire Kermadec Trench). The $M_{\text{W}} 9.4$ scenario was included, as the possibility of rupture of the entire trench cannot be ruled out empirically (McCaffrey, 2007), although if such an event occurs it is likely to be very rare (i.e., with a recurrence interval of several thousand years). The numerical results show that tsunami generated from the southern and/or middle sections of the Kermadec subduction zone pose a greater hazard to the coast of New Zealand than tsunami generated along the northern Kermadec Trench. For tsunami generated on the northern Kermadec Trench, the majority of energy travels towards the open Pacific, as well as through the South Fiji Basin to the

northwest toward Norfolk Island and New Caledonia. In contrast, if the southern and/or middle sections of Kermadec plate interface are ruptured, refraction effects due to the sloping continental shelf of North Island will gradually bend tsunami waves onshore, and thus much more energy will be directed toward the coasts of the Northland and Auckland regions.

The largest event modelled by Power et al. (2012) was an M_w 9.4 earthquake rupturing the plate interface along the entire Kermadec Trench. Numerical models show that such an event could produce tsunami causing tremendous damage throughout the coastal areas between Gisborne and Northland. Along the coasts of the Raukumara Peninsula and the Bay of Plenty, tsunami wave amplitudes around 5–10 m would be expected. On many parts of the northeastern coasts of Great Barrier Island and Northland, calculated tsunami run-up heights would be about 15–20 m above the normal level. Tsunami waves of over 10 m amplitude also would strike the southwestern coast of the Aupouri Peninsula and Ahipara Bay on the west coast of northern Northland. However, if such giant events do occur, they would be very rare indeed.

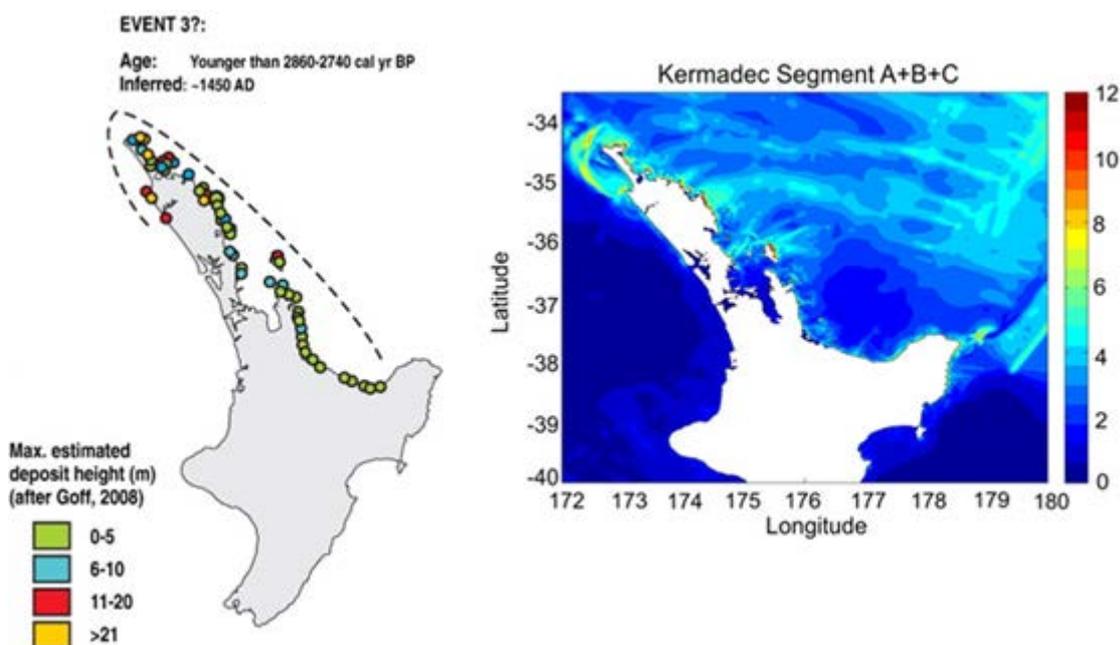


Figure 5.2 (Left) Paleotsunami deposits and their height distribution for the event with an inferred date of ~1450AD (Goff, 2008; Goff et al., 2010); (Right) maximum water level distribution (offshore) in modeled M_w 9.4 Kermadec Trench scenario (Power et al., 2012).

There is a remarkable similarity in the distribution of estimated tsunami heights caused by the hypothetical M_w 9.4 event and the distribution of paleotsunami deposits assigned to a tsunami in approximately 1450 AD by Goff et al. (2010) (Figure 5.2). As most of the tsunami energy originating on the northern half of the Kermadec Trench passes to the north of New Zealand, an earthquake on just the southern part of the trench could achieve a similar distribution of run-up heights at a lower magnitude, though this would probably still need to be at least M_w 9. Unless a more plausible explanation for the paleotsunami data can be made, the possibility of such events needs to be taken seriously.

For the scenario in which a 300 km long section of the southern Kermadec Trench (between ~36°S to ~38°S) ruptures in an M_w 8.5 event, water levels 3–5 m above mean sea level would occur along the northeastern coasts of Northland and the Auckland region, as well as the coasts of Gisborne and the Bay of Plenty. Waves with amplitudes of over 10.0 m would strike the northeastern coasts of Great Barrier Island. Water level increases of 1.0–3.0 metres would occur along the southwestern coast of Northland.

5.2.1.4 South of New Zealand (including Macquarie Ridge)

Most plate boundary zones in the Southern Ocean have horizontal (strike-slip) movement and large earthquakes in these zones are unlikely to produce large tsunamis. There are no highly active subduction zones in the Southern Ocean. The Hjort Trench (56°S–60°S) and subduction zone is the only part of the margin where significant vertical deformation could be expected in a large earthquake, and the orientation of the zone plate boundary there would partially direct tsunamis towards New Zealand. However, recent studies of the Hjort trench area (Meckel et al., 2003) suggest subduction in this region is immature with little significant down-dip movement, so large thrust earthquakes are unlikely to occur.

Historically, large earthquakes along the Macquarie Ridge (M_w 8.1 earthquakes in 1989 and 2004), and further south near the Balleny Islands (M_w 8.1 in 1998) have involved strike-slip (horizontal) movement, producing small tsunamis (less than 50 cm) in southern New Zealand. The effects of an M8.3 earthquake on the Macquarie Ridge in 1924 have not yet been well researched.

The Puysegur subduction zone, immediately to the south of New Zealand, is a local source and described in Section 5.3.1.7.

5.2.2 Volcanoes

There are 26 volcanoes (>10 km in diameter) along the active Taupo - Kermadec arc that lie between 300 km and 1000 km from mainland New Zealand (Figure 5.3), we will refer to these as the Kermadec Volcanoes. Three “scenarios” of how these volcanoes represent possible regional tsunami sources are:

- catastrophic submarine silicic eruption and caldera collapse,
- large catastrophic sector collapse,
- frequent small avalanches on edifice flanks.

5.2.2.1 Catastrophic submarine silicic eruption and caldera collapse

Submarine eruptions of silicic-type magma can occur in a series of explosive pulses, each of which can generate tsunamis. Associated caldera collapse, such as occurred at Kratatau in 1883, is another possible tsunami source.

South of 30°S, four silicic caldera complexes with explosive styles of eruption have been surveyed—Macauley, Havre, Brothers and Healy (see Figure 5.3). A fifth caldera (Rumble II West) has a partial silicic composition, and thus may generate tsunamis on occasion. Macauley is the largest caldera and the source of the Sandy Bay Tephra pyroclastic eruption 6300 years ago. Estimates of the eruption volume vary; Latter et al. (1992) estimated 100 km³, Lloyd et al. (1996) estimated a lower limit of 1–5 km³. Recent sea floor mapping reveals an unfilled caldera ~10.8 km long and ~8.2 km wide (Wright et al., 2006)—that can be interpreted to represent the eruption of 35–58 km³ of material. Havre is a silicic caldera volcano mantled in pumice of unknown age, but the pumice is interpreted to be older than the Sandy Bay Tephra eruption. Havre erupted in July 2012, producing an ash plume and a pumice raft estimated to cover circa 20,000 km² (volume ~1 km³) but no tsunami was observed. Brothers and Healy volcanoes have <3.5 km wide calderas, and consist of explosive-type lavas (Wright and Gamble, 1999). Healy was probably formed by catastrophic submarine rock and ash flow eruptions, with the destruction of a 2.4–3.6 km³ volcanic cone and formation of a caldera. The eruption is tentatively correlated with part of the Loiseles

Pumice of c. 600 years ago, which is found along much of the eastern North Island coastline (Wright et al., 2003).

5.2.2.2 Large catastrophic sector collapse

Seafloor mapping reveals that many of the southern Kermadec volcanoes have undergone large-scale mass-wasting or sector collapse. The volumes of material involved in each sector collapse are currently undocumented. However, an upper limit to any individual sector collapse is probably 4–5 km³, as evinced by the collapse of the western flank of Rumble III (Wright et al., 2004). Both the age of the Rumble III collapse in particular, and the frequency of large sector collapse in general are unknown, but possibly have recurrence intervals of >10,000 years for any one volcano.

5.2.2.3 Frequent small landslides and debris avalanches

All Kermadec volcanoes, to varying degrees, show evidence of small and frequent landslides and debris avalanches (Wright et al., 2006). Typically these collapses are <1 km³. The timing and frequency of such failures is almost entirely unknown, but the one example based on repeat multi-beam surveys of Monowai volcano reveals the collapse of 0.03 km³ of material between 1998 and 2004 (Wright unpublished data). Similar shallow failures, typically 10–300 m thick, occur on all southern Kermadec volcanoes. The recurrence interval of such events is unknown but could be 10 years for any one volcano.

5.2.2.4 Summary of Kermadec volcanoes as tsunami sources in New Zealand

No historical records exist of volcanic activity in the Kermadec chain producing tsunami in New Zealand or elsewhere. Therefore we have little basis for modelling possible tsunami from activity in the Kermadec volcanoes (Table 5.2, Figure 5.4). In general, the volumes of the eruptions associated caldera collapses and the scale of sector collapse features so far identified are significantly (at least an order of magnitude) smaller than has been proposed in the literature for damaging tsunami effects at distances of 1000 km or so. Additionally, a numerical model of a 1 km³ rock and ash avalanche entering the sea from Mayor Island in the Bay of Plenty indicated only a 0.5 m tsunami on the coast about 30 km distant (de Lange and Prasetya, 1997) so we expect that events with volumes typically 10 times larger but at 10–30 times the distance will have effects no larger than indicated by the modelling of the Mayor Island event. However, significant doubts remain about the source characteristics, and about the effectiveness of rock and ash flows/avalanches and collapsing high altitude eruption columns in producing tsunami that could be damaging at the 300–1000 km distances between the volcanoes and New Zealand. Volcanic unrest in the Kermadec volcanoes leading to a major eruption is expected to have a long lead time, so an extended period of preparation prior to any tsunami should be possible.

5.2.3 Landslides

No landslide sources, at regional distances, have been thus far identified that are sufficiently large or frequent enough to justify the inclusion of regional distance landslides in the tsunami source model for this study. However, further consideration of this potential source, by searching for giant landslides such as the Matakaoa and Ruatoria features of eastern North Island (section 5.3.2.1) along the Tonga-Kermadec and Puysegur-Macquarie margins is warranted.

Table 5.2 Summary of available data from Kermadec chain volcanoes. * = local sources < 100 km from New Zealand.

Volcano	Edifice/Caldera Volume (km ³)	Eruptive Volume (km ³)	Collapse Volume (km ³)	Age of Last Event (yrs before present)	Frequency (yrs)
Macauley	17.4	100 <5 35-58		6,300	?
Havre	6.8	1-10		<1	?
Brothers	2.8	~5		>~5,000	?
Healy	2.4-3.6	10-15		600	?
Rumble III			4.4	Unknown	?10,000
Generic volcano			0.03	<1	?100
Mayor Is.*		~1		6,300	~10,000
White Is.*			0.01	?100	?100

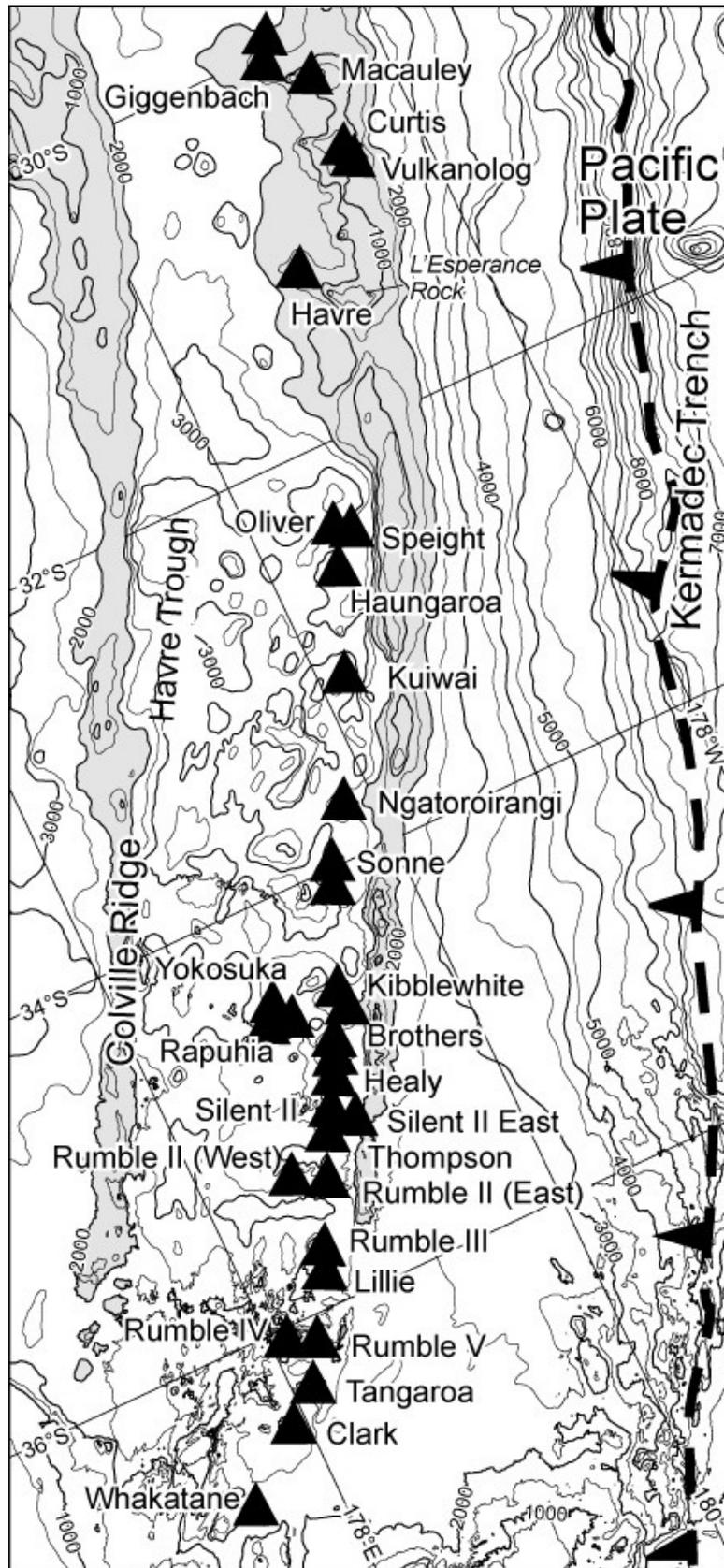


Figure 5.3 Distribution of submarine volcanoes along the southern Kermadec arc between 30°S and 36°30'S (after Wright et al., 2006).

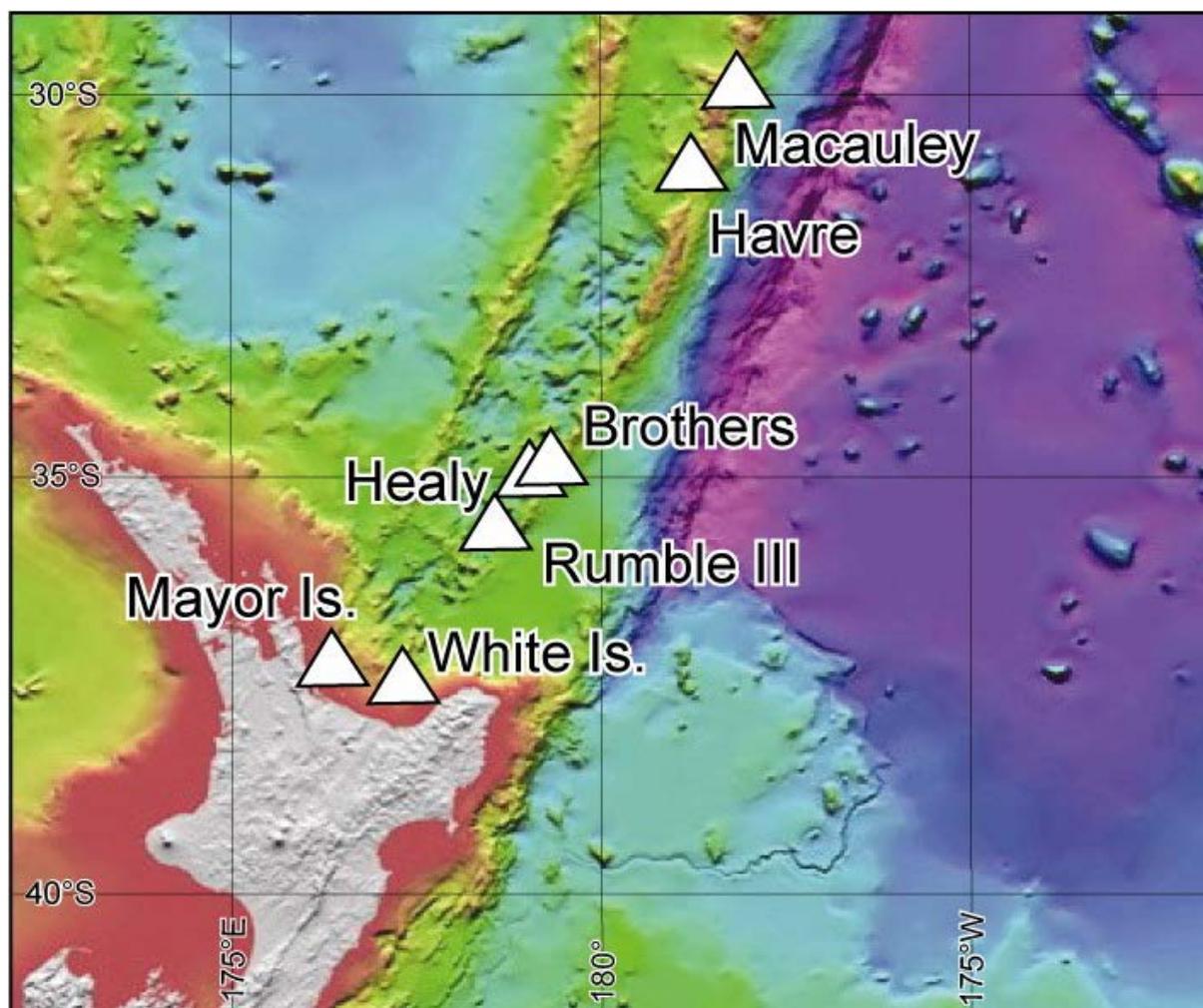


Figure 5.4 Location of possible volcano sources for tsunami along the southern Kermadec arc.

5.3 LOCAL SOURCES

By definition, tsunami generated by local sources arrive at the nearest coastline within an hour, and many can arrive within minutes. New Zealand's location astride a plate boundary means that it experiences many large earthquakes, some of which cause local-source tsunami. It is also exposed to local-source tsunami from submarine and coastal landslides, and island and submarine volcanoes.

5.3.1 Earthquakes

Local earthquakes have the potential to produce catastrophic tsunami, with 10 m or more run-up, over a short length of coast (local impact, i.e., tens of kilometres of coast) or over a longer length of coast (regional impact, i.e. hundreds of kilometres of coast). The impact depends on the extent of fault rupture and seafloor deformation, which in turn depends on the magnitude of the earthquake. The tsunami resulting from a very large, 200–300 km long rupture of the Hikurangi Trough plate-interface on the east coast of the North Island could affect 200–300 km or more of the nearby coast, with large run-ups. Such an event could cause significant to severely damaging waves along much of the east coast and in the Chatham Islands.

Some coasts are more at risk from tsunami than others because of their proximity to areas of high local seismic activity, but no part of New Zealand coastline can be considered completely free from local source tsunami hazards. The tsunami hazard is also present around the shores of our larger freshwater lakes, although consideration of this hazard is not within the scope of this study.

The probabilistic tsunami hazard model described in Chapter 6 uses earthquake magnitude-frequency distributions for the subduction zones around New Zealand that have been derived for use in the preliminary Global Earthquake Model (GEM) from the various studies described below.

Information on crustal faults comes from historical earthquakes and from the mapping of active faults in the offshore areas around New Zealand—these have been the primary methods for determining the local potential sources for earthquakes producing tsunami. The model for these faults used for the present probabilistic study is based on the offshore faults in the New Zealand Hazard Model (NZSHM; Stirling et al., 2012) and removing those too small, or with the wrong fault mechanisms, to produce significant tsunami. Some additional faults were tentatively added to the model following group discussions, in particular faults along the Hikurangi Outer Rise, the west coast of the South Island, and in the Tasman Bight (see Appendix 5 for further details).

Stirling et al. (2012) produced a series of maps showing the locations of the faults in the NZSHM, and this reference may be used to locate the faults described in the following sections.

5.3.1.1 Kermadec Trench

The Kermadec Trench is both a regional and a local source for tsunami. Earthquakes that rupture only on the northern half of the Kermadec Trench produce tsunami that take more than an hour to reach New Zealand (excluding the Kermadec Islands), while those whose ruptures include portions of the trench south of $\sim 33^{\circ}\text{S}$ produce tsunami that reach parts of New Zealand in less than an hour. Because of the orientation of the trench, even for tsunami originating on the southern part of the trench, the most strongly affected coasts are usually more than one hour travel time away. The main description of the Kermadec Trench as a tsunami source is in Section 5.2.1.3.

5.3.1.2 Tsunami sources in offshore eastern North Island

We recognise that a significant source of large vertical seafloor displacements during earthquakes is in conjunction with the Hikurangi subduction margin off the eastern North Island. Tsunami could be generated by large to great earthquakes ($M7.5\text{--}9.0$) on the plate interface itself from slip between the two opposing plates (Wallace et al., 2009), or by rupture of steeper faults that extend upward through the Australian plate (see Figure 5.5).

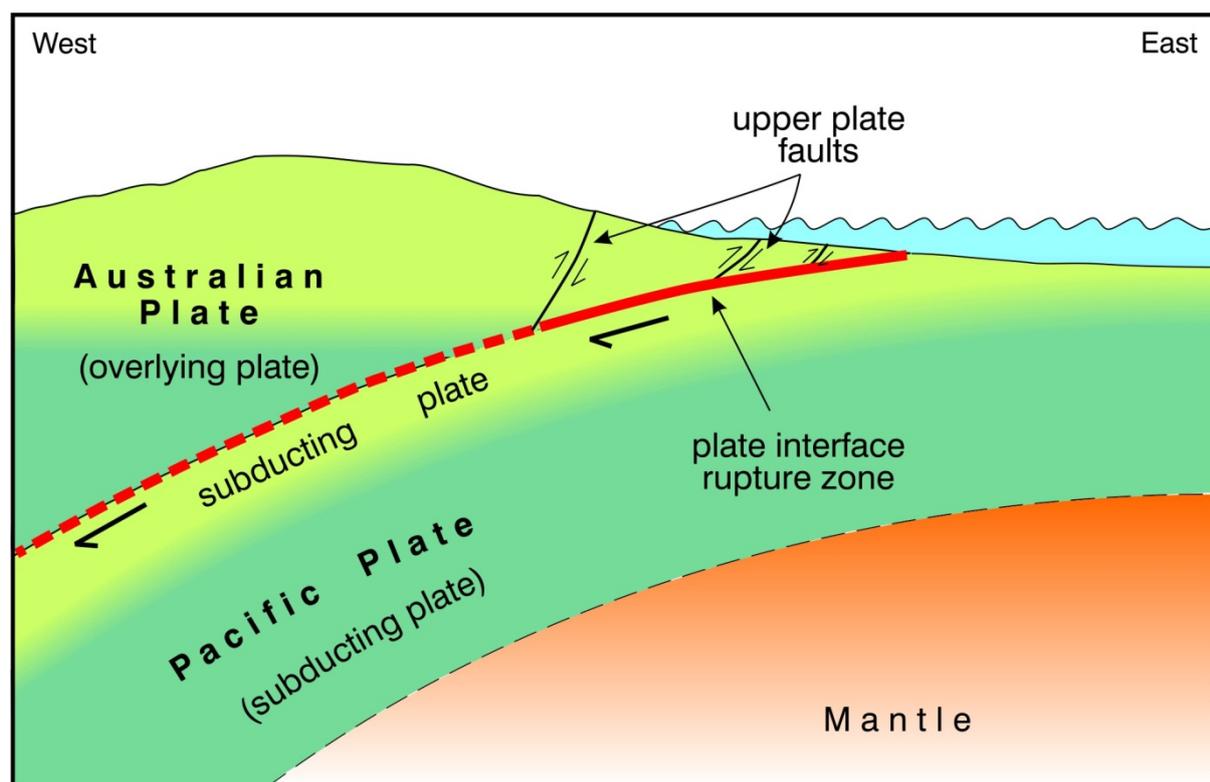


Figure 5.5 Schematic cross-section through the Hikurangi subduction zone.

NIWA scientists have mapped many faults in the offshore area from the inner shelf (~ 50 m water depth) to the deep ocean (>2000 m water depth) of the Hikurangi subduction margin (e.g., Barnes et al., 1998). This mapping has defined the subduction front and structural features on the shelf and slope. Many of these structures mimic onshore faults and folds, having lengths of tens of kilometres and heights of up to 500 m.

Some information on fault slip rates comes from studies of specific faults such as the Lachlan Fault, offshore of Mahia Peninsula (Barnes et al., 2002) and from the presence of uplifted marine terraces of Holocene age (c. <10,000 years) along the east coast (Berryman et al., 1989; Berryman, 1993; Berryman et al., 2011). Data from these linked studies provide a basis for assigning fault parameters to other structures when location and fault length are the only data available. For example, studies of the uplifted marine terraces on Mahia Peninsula and the offshore seismic stratigraphy of the Lachlan Fault show that this fault is capable of generating a large surface earthquake that ruptures the sea bed every 615–2333 years (Barnes et al., 2002). This is confirmed by the evidence for five uplift events on the peninsula in the last c. 5000 years (Berryman, 1993). Other data that provide tie-points for assigning fault parameters include the source dimensions and magnitude of the M 7.8 Hawkes Bay earthquake of 1931.

Research summarised in the New Zealand Seismic Hazard Model indicates more than 80 faults in the continental shelf and slope part of offshore eastern North Island occur in bedrock that is interpreted to be strong. Further offshore the fault structures have developed in softer, weaker rocks and it is unclear whether these are strong enough to break independently in large earthquakes. It may be that these “outer margin” faults only rupture in association with major subduction thrust earthquakes. Currently faults are treated as independent tsunami sources in our tsunami hazard model if they are represented as independent seismic sources in the national seismic hazard model (Stirling, 2012); however the assumptions regarding these outer margin faults warrant further work. Earthquake magnitudes and recurrence

intervals have been estimated. based on the length of the faults (as they are expressed on the sea floor) and estimated slip rates, Outer Rise earthquakes, occurring in the Pacific plate where it bends prior to subduction, are a potentially important tsunami source that has so far received little attention. A major component of the Samoa 2009 tsunami was caused by an Outer Rise earthquake, as have some major historical tsunami in Japan. This type of tsunami source is discussed in Appendix 5.

The Hikurangi subduction zone is arguably the most important local-source tsunami hazard posed to New Zealand (and a comparable hazard to the Kermadec Trench subduction zone, which we regard primarily as a regional source). However few data are available on the timing and size of large-to-great earthquakes from the Hikurangi subduction zone. We have developed the source model with a range of possible earthquake sizes and recurrence intervals, based on historical earthquakes, plate motion rates from GPS, and GPS measurements of the accumulation of elastic strain. Elastic strain accumulation measurements help to assess which portions of the plate interface are currently locked or “stuck” and building up strain for future earthquakes.

GPS measurements show that the southern Hikurangi subduction interface (beneath Wellington and the Wairarapa) is currently “locked” and that strain is building up that will probably eventually be released in a great ($M_W > 8.0$) subduction megathrust event. This is in contrast to the northern part of the Hikurangi margin (adjacent to Hawke’s Bay, Gisborne and Raukumara peninsula) which appears to be a region of more moderate subduction interface earthquakes ($M_W < 7.1$) and steady creep rather than extremely large earthquakes. However, this does not mean that there is no tsunami hazard posed to the Gisborne region by the Hikurangi subduction zone. On the contrary, in 1947, two historic M_W 7.0–7.1 earthquakes offshore of Gisborne led to much larger than expected tsunami with run-ups of up to 10 and 6 m. These earthquakes had classic characteristics of “tsunami earthquakes”. They were located close to the trench where the interface is at very shallow depths, rupture velocities were slow (assumed to be < 1 km/s), and rupture durations were long (at least 40 and 25 seconds for the March and May 1947 events, respectively). A low energy release at high frequencies resulted in low M_L values (5.9; 5.6) compared to M_S values (both 7.2) and M_W values (7.0–7.1; 6.9–7.1), and larger than expected tsunami, with run-ups of 10m and 6m respectively (Downes et al., 2000; Doser and Webb, 2003). Power et al. (2008) suggested that earthquakes might recur in the source area of the 1947 tsunami earthquakes as frequently as every ~70–80 years.

Overall, we expect that the southern Hikurangi margin experiences less frequent (i.e., at ~300-900 year intervals) but extremely large to great ($M_W > 8.0$) subduction earthquakes. The northern Hikurangi margin, on the other hand, probably experiences more frequent, moderately sized earthquakes that are located at very shallow levels along the Hikurangi trench; these could also produce significant tsunami (e.g., the 1947 earthquakes near Gisborne, see Sections 3.2 and 4.2.1 for further details). We also cannot rule out very infrequent subduction thrust events that rupture the entire Hikurangi margin in $M_W \sim 9.0$ earthquakes. If such events occur, they would produce devastating tsunami similar to those observed in Japan in the M_W 9.0 Tohoku event in March 2011. We account for these variations in our assigned parameters for the subduction zone as a whole (Appendix 3), but uncertainties in both the distribution of earthquake magnitudes and their recurrence intervals are large, and the properties determining these characteristics are likely to vary along the length of the subduction zone. Further work is needed to incorporate such varying factors into a tsunami hazard model.

5.3.1.3 Tsunami sources from faults in the Bay of Plenty

There are many active faults in the offshore area of the Ruapehu-White Island volcanic zone. These faults typically have smaller dimensions than the faults offshore of the eastern North Island, and the maximum earthquake that these faults can produce is believed to be about M 7, with 2–3 m of potential seabed displacement on a fault up to ~30 km long. These relatively small sources are not thought to be capable of producing large tsunamis.

5.3.1.4 Tsunami sources from faults near Auckland

The active Kerepehi Fault probably extends into the Hauraki Gulf about 40 km east of Auckland, and is the only offshore active fault known in the Auckland region. The fault can produce earthquakes up to about M 7, similar to those in the Bay of Plenty. At 40 km distance, we consider it unlikely that the fault poses a significant tsunami hazard to Auckland. In addition, de Lange and Healy (2001) and Chick et al. (2001) completed some numerical modelling of a tsunami generated by the Kerepehi Fault source, and found it would not produce a run-up of 2 m or more in Auckland.

5.3.1.5 Tsunami sources from faults in the Cook Strait and offshore Marlborough

Numerous active faults occur in the Cook Strait area and offshore Marlborough (Barnes et al., 1998; Barnes et al., 2008; Pondard and Barnes, 2010), including the offshore southern part of the Wairarapa Fault that in 1855 generated a tsunami with 10 m of local run-up (and up to a 5 m run-up in Wellington). The potential of the active faults for producing tsunamis are based on their length and by assigning earthquake magnitudes based on their onshore continuations in Marlborough and southern North Island. The southern section of the Wairarapa fault ruptured into Cook Strait, with at least 6 m of vertical movement, and this produced the tsunami mentioned above. Using the Abe local source equation (see section 4.2 for details) we calculated that the offshore section of the fault, to produce the tsunami run-up observed in the Wairarapa, Wellington, and Kapiti Coast, would be approximately equivalent to a $M_W 7.7$ earthquake. This approximation appears to be reasonable, based on recent numerical propagation models of this source (Rob Bell, pers. comm., 2005; Cousins et al., 2007). Other Marlborough and Wellington region faults include the Boo Boo, Needles, Chancet, Campbell Banks, Cloudy, Te Rapa, Kekerengu Bank, Palliser-Kaiwhata, Wellington, Ohariu, Awatere and Wairau Faults. For this study we have excluded strike-slip faults such as the Boo-Boo Fault from the hazard model, though the possibility of uplift along the small step-overs in such faults remains, as does the possibility for some residual vertical component to the fault motion (Cousins et al., 2007). Apart from the Wairarapa Fault, the largest earthquakes assigned to offshore faults in this region are M 7.5–7.8.

5.3.1.6 Tsunami sources from faults in the western Cook Strait and offshore Manawatu

An extensive marine survey of the region offshore from the Manawatu-Kapiti area has recently been completed (Lamarche et al., 2005), and has provided valuable new insight into the location and characteristics of offshore faults in the region. These structures have a modest potential to generate tsunamis (maximum earthquake magnitudes of up to $M_W \sim 7.7$), but they may be important as they are located at short distances from urban areas on the Kapiti coast, Porirua and northern South Island. Additional faults extending from the Tasman and Golden Bay area towards Taranaki were tentatively identified for this study; details are given in Appendix 5.

5.3.1.7 Tsunami sources from faults in southern South Island

In the offshore Fiordland region, plate boundary structures that include the Alpine Fault and the Puysegur subduction zone are capable of producing large-to-great earthquakes of $>M 8$, similar to or larger than, the $M_W 7.8$ Dusky sound earthquake in July 2009. This, coupled with early historical records of drownings on the south Fiordland coast, probably by a tsunami in the 1820s, has led to recent numerical simulations of tsunami generation and propagation from these sources (Downes et al., 2005; Prasetya et al., 2011a). Because the Alpine Fault is predominantly a strike-slip fault, the structure is not considered likely to generate significant tsunami except at localised areas where the fault steps from one strand to another and large vertical movements are possible. Thus, the tsunami source tends to be very localised, which could generate a large run-up locally, but is unlikely to travel as far as Invercargill.

The Puysegur subduction interface has the potential to generate major tsunami. The plate interface here is the mirror image of the Hikurangi subduction zone, in that the Australian plate is here subducted beneath the Pacific plate. The relative motion of the plates is highly oblique, though the implications of this for tsunami generation are unclear. The subduction zone has been studied by Hayes and Furlong (2010), and models have been constructed by Downes et al. (2005), Goff et al. (2009), and Prasetya et al. (2011b). The bathymetry off the southern South Island appears to offer some natural protection to southern shores. This is because the water shallows at a substantial distance from the coast and some of the energy is dissipated at the shelf edge.

5.3.2 Landslides

New Zealand is vulnerable to tsunami hazard from both coastal and submarine landslides. In historical times, several earthquake-triggered coastal landslides have resulted in documented localised tsunami. These include a tsunami north of Westport associated with coastal landslides triggered by the 1929 $M_W 7.8$ Buller (Murchison) earthquake, a tsunami near Napier associated with coastal cliff failure triggered by the 1931 $M_W 7.6$ Hawke's Bay earthquake, and a tsunami in Charles Sound during the 2003 $M_W 7.2$ Fiordland earthquake that had a run-up of 4–5 m (Hancox et al., 2003). In addition to earthquakes, wave action and coastal processes, heavy rain and human activities can trigger coastal landslides.

There are no confirmed historical instances of tsunami triggered by submarine landslides in New Zealand. Some reported tsunami do bear characteristics of landslide-source tsunami however, e.g. at Goose Bay south of Kaikoura (Du Bois 2012). Marine geophysical data demonstrate the widespread occurrence of submarine landslide deposits on the sea floor around New Zealand's submerged continent.

5.3.2.1 Submarine landslides

The number of published studies on New Zealand submarine landslides has increased significantly since 2005 (Lewis and Barnes, 1999; Carter, 2001; Collot et al., 2001; Faure et al., 2006; Crutchley et al., 2007; Lamarche et al., 2008; Mountjoy et al., 2009; Mountjoy et al., 2009; Barnes et al., 2010; Joanne et al., 2010; Kukowski et al., 2010; Pedley et al., 2010; Micallef et al., 2012; Mountjoy and Micallef, 2012). In addition to published studies, numerous un-published high-resolution multibeam data sets have been collected by NIWA and other agencies, resolving large areas of New Zealand's seafloor. The combination of published and unpublished data indicate widespread slope failure off the Bay of Plenty, the length of the eastern coast of the South and North islands, Fiordland, and the West Coast of

the South Island. Mapped landslide source areas occur from 2 to >50 km distance from the coastline, and in water depths ranging from 100 to 3000 m.

The size of known landslide source areas varies from the smallest failures resolvable using high-resolution ship-board multibeam data (several tens of metres across) to continental margin scale failures (several tens of kilometres across). The size range is poorly distributed, however, with the majority of source volumes estimated at 10^{-3} – 1 km³ (e.g., Micallef et al., 2012), a few failures at 10km³ scale (Mountjoy and Micallef, 2012; NIWA unpublished data); and two landslides near East Cape at the extreme end of the volume scale (>100km³) (Collot et al., 2001; Lamarche et al., 2008).

While the magnitude of submarine landslides can be relatively easily measured in geophysical data, determining the age of submarine landslides is significantly harder and more costly. Typically only the largest landslides justify individual case studies and thus have known ages. A summary of the details of the largest landslides documented on the New Zealand margin is provided in Table 5.3.

Table 5.3 Summary parameters of the largest submarine landslides documented on the New Zealand margin.

Landslide Name	Volume (km ³)	Upper water depth (m)	Age (kyr)	reference
Matakoa Debris Avalanche	~430	~200	600±150	(Lamarche et al., 2008)
Ruatoria Avalanche	3150±630	~140	170±40	(Collot et al., 2001)
Paritu Debris Avalanche	~30	~800	7.6±0.6	(Mountjoy and Micallef, 2012; Pouderoux et al., 2012a)

As the low number and older ages of the very large-scale mass failures indicates, we expect that such events occur very rarely, and are outside the time frame relevant to risk assessment. To complete a risk assessment, a key piece of information needed is the recurrence interval (return time) for hazardous events. It is generally not possible to gain recurrence-type information from individual mass failure complexes, however, slope-failure-recurrence is being used as a proxy for seismic activity on the Hikurangi and Fiordland margins (c.f. Goldfinger, 2011). The method is based on the assumption that significant earthquakes will trigger widespread slope failure across a large area of a margin. Detailed work by Pouderoux et al. (2012b) across three locations on the Hikurangi margin indicates mean return times for turbidite deposits of 270–430 years. This data is very useful for understanding the frequency of slope disturbance on active margins, however it does not indicate the magnitude of the slope failure at the source, and thus is not directly applicable to assessing the tsunami hazard from submarine landslides.

The best studied and most widely publicised landslide tsunami scenario in New Zealand is for the head of Kaikoura Canyon. An unstable accumulation of 0.25 km³ of sediment there has been inferred, based on local sedimentary processes (Lewis and Barnes, 1999). The shallowest area of the sediment is in only 35 m water depth. Dated sedimentary events within Kaikoura Canyon and further down-slope suggest that sediment failure occurs every 200–300 years. Based on this scenario, Walters et al. (2006) carried out a tsunami simulation and found that for the average-case scenario, the inferred landslide is capable of generating waves up to 13 m above tide level in adjacent Goose Bay, and 2 m at Kaikoura. The arrival time for these waves is very short, at 1 minute and 15 minutes respectively. Large local tsunami in this region are supported by historical records and paleo-tsunami data (Du Bois, 2012).

Very little other research has been carried out to model landslide-tsunami for specific scenarios around New Zealand. Joanne (2008) modelled the potential of smaller failures on the flanks of the Matakaoa Submarine Instability Complex for producing tsunami, and demonstrated a potential for significant inundation at Te Araroa on the East Cape. Kukowski et al.(2010) infer a potential for large-scale failure of a frontal ridge of the Hikurangi Margin. The inferred worst-case scenario is for complete failure of the “Rock Garden” ridge offshore of Hawke’s Bay involving 150–170 km³ material, initiating in ~700 m water depth. Modelling the tsunami potential for this worst case generates a sea-surface disturbance of 40 m, however no wave runout was modelled to assess the effect at the coast for this event.

While it is definitely useful to undertake tsunami assessments for individual landslides based on specific scenarios, in order to determine the hazard and risk to coastal populations it is necessary to consider multiple landslide sources across a broad area. As with earthquake hazards, the uncertainty inherent in magnitude and frequency of submarine landslides best lends itself to probabilistic hazard assessment (Grilli et al., 2009).

5.3.2.2 Coastal landslides

Sub-aerial landslides entering the sea (or lakes), especially into deep water, can generate major but local tsunami. Some historical examples were noted in Section 5.3.2. There is no systematic monitoring of coastal cliff stability around New Zealand. At any time, there are always coastal cliffs with marginal stability, requiring only a minor trigger to collapse them.

We have carried out a qualitative assessment of this hazard in the vicinity of each of the urban centres considered in this project. Criteria for assessing tsunami-inducing coastal-landslide hazard have included:

- topography (steep, high slopes close to water)
- geology (the relevant strength and structure of the rock)
- known landslides (presence and types that can be identified as reaching the water)
- historical evidence – e.g. 1931 Napier, 1855 Wellington, 1929 Murchison (note – all of these landslides are associated with large earthquakes. There is a much lower risk of similar landslides without earthquakes).

Whangarei	Whangarei Heads could pose a small threat. Landslides at Onerahi are too small to cause significant waves.
North Shore	No risk - no steep, large slopes at coast.
Waitakere	Little risk (apart from west coast beaches and north side of Manukau Heads, which have significant landslide potential – possibly waves of a few metres over a distance of up to 1 km).
Auckland	Some risk at St Heliers - Achilles Point - Karaka Bay. Coastal-cliff collapses in the order of 100 m wide, but into shallow water.
Manukau	Probably has greatest risk in Auckland region, especially the north side of the harbour from Green Bay to Manukau Heads.
Tauranga	Although there are many landslides, none seem capable of generating more than small waves, except for a small chance of large failures of Mount Maunganui.

Whakatane	Possibly greatest risk is from Moutohora Island just offshore. It has collapsed to the north pre-historically, and there may be a risk of collapse to the south. Whakatane headland collapse could pose a danger, although the rock strength is good
Gisborne	Low hazard, almost no risk. Possible nearshore uplift caused by landslides, but very rare. Hill at Titirangi has greatest potential to cause waves, but very small – only 100m high and not steep enough.
New Plymouth	Collapse of Paritutu cliffs could cause modest waves, but Whitecliffs is too far away.
Napier	Local risk on east side of Bluff Hill – small rockfalls (but none into sea in the 1931 earthquake). Greater risk is presented by landslides between Napier and Wairoa, as in 1931, but likely to affect only a limited area (<10 km).
Wanganui	Landslides at Castlecliff are unlikely to cause any waves. Greatest risk of wave generation from landslide is from Shakespeare Bluff into the river. The effects are likely to be small.
Kapiti	Some risk between Pukerua Bay and Paekakariki. Brendan's Beach and the restaurant south of Paekakariki are at greatest risk.
Porirua	A small risk at Titahi Bay, but little risk elsewhere.
Lower Hutt	Eastbourne/Gracefield/Seaview is at some risk. Some risk from Wellington fault scarp earthquakes affecting Petone, but this is only likely to occur in association with large earthquakes.
Wellington	Coastal collapse between Ohiro Bay and Sinclair Head is a hazard for south coast bays. Some risk from fault scarp collapse into harbour. Some risk in larger landslides such as at Worsler Bay, but effects likely to be limited, and only likely in the event of large earthquakes.
Nelson	Possible but low likelihood of large-scale movement at Tahunanui causing heave at the toe of the slide out to sea.
Christchurch	A small risk from rockfalls into Lyttelton and Akaroa harbours. No very large-scale landslides are apparent, and most slope instability is shallow failures in loess and regolith.
Timaru	Low cliffs at Caroline Bay/Dashing Rocks pose a negligible risk (high quality rock).
Dunedin	No large landslides, capable of causing large waves, are known adjacent to Otago Harbour. The outer coast cliffs both east and west of the city (Highcliff, Lawyers Head, St Clair cliffs, Tunnel Beach) have potential for landslides large enough to cause waves at coastal suburbs. At least one large prehistoric landslide (Lovers Leap) is known, but in general the rock appears solid.
Invercargill	May be affected by tsunami from very large landslides in Fiordland, but only as a result of a very large earthquake. Otherwise the risk is very low. No apparent risk at Bluff Harbour.

5.3.2.3 Conclusions

The likelihood of coastal landslides inducing tsunami is low except during large earthquakes, in which case other tsunami-generating phenomena are likely to be more important, apart from in the immediate vicinity of the landslides.

The greatest potential for very large landslides is in relatively uninhabited areas of very high relief such as Fiordland, but the risk of such events must be orders of magnitude lower under undisturbed conditions than during earthquakes.

5.3.3 Volcanoes

5.3.3.1 Mayor Island and White Islands

Mayor and White island volcanoes are very close potential source for tsunami. Mayor Island has produced both explosive and lava flow eruptions, and includes three phases of caldera collapse. The last caldera collapse, associated with the largest eruption, occurred 6,300 years ago (Houghton et al., 1992) and included the movement of rock and ash flows into the sea.; this event is probably the only recorded instance of rock and ash flows entering the sea within the New Zealand region. Numerical modelling of a credible 1 km³ (“Mt St Helens scale”) rock and ash flow from Mayor Island, that enters the sea, would produce a 0.5 m high tsunami on the adjacent coast around Whakatane (de Lange and Healy, 1986; de Lange, 1997).

White Island is the emergent summit of a larger submarine volcano. Eruptions have included both lava flow and small explosive eruptions of mostly andesite (of typically moderately explosive style), but including dacite (associated with a more energetic eruptive style), though the eruption history of the volcano is poorly known. A small collapse of the inner crater wall in 1914 produced a debris avalanche that may have entered the sea. The active hydrothermal system weakens the volcano structure and enhances the potential for sector collapse on both the outer subaerial and submarine flanks.

The probability of generation of a significant tsunami from White Island is considered low (de Lange and Healy, 1986; de Lange and Prasetya, 1997), not least because the most likely sector collapse direction is toward the east, and thus any tsunami generated would be directed offshore. Other small caldera volcanoes and associated pumice deposits occur on the outer Bay of Plenty continental slope (Gamble et al., 1993; NIWA unpublished data). Based on the low likelihood of damaging tsunami indicated by these specific modelling studies, we find no reason to add these volcano or landslide sources to the tsunami source model in this project.

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