# 3.0 PALEOTSUNAMI AND HISTORICAL TSUNAMI DATABASES

### 3.1 HISTORICAL TSUNAMI RECORDS

The New Zealand historical tsunami database has been compiled by Gaye Downes (GNS Science). This is currently an unpublished database but work is in progress to publish it as an online, searchable database. In this section we summarise the New Zealand historical tsunami database by analysing the distribution of tsunami through time and the sources of the tsunami. Brief descriptions of the most significant historical tsunami are given. The New Zealand historical tsunami database assigns a "validity" ranking of 0–4 for tsunami reports, with 4 being a definite tsunami and 0 being an erroneous tsunami report. For our analysis we only use tsunami with a validity ranking  $\geq 2$  (Table 3.1).

**Table 3.1** Validity rankings for historical tsunami of New Zealand 1985–2011. Very doubtful or erroneously attributed tsunami of validity "0" and "1" (shaded grey) are noted in the historical tsunami database but are excluded from the analysis undertaken in this report. \*All the "3"-level tsunami are pre-1932 except for a tsunami on Tasman Lake (Aoraki-Mt Cook) on 22nd February, 2011, which was caused by ice-calving from the Tasman Glacier. The ice calving was triggered by the M6.3 Christchurch earthquake, but it is debatable whether this was a true tsunami.

Validity		# tsunami	Notes on ages
4	Definite tsunami	60	
3	Probable tsunami	8	Most pre-1932*
2	Questionable/unlikely	12	All pre-1928
1	Very doubtful/highly unlikely	15	All pre-1993
0	Erroneous tsunami report	6	All pre-1870

New Zealand has been affected by at least 80 tsunami from 1835–2011 (Downes, unpublished data). Figure 3.1 shows the distribution of tsunami through time during this historical period. The rates of tsunami appear relatively steady at approximately 4–5 per decade up until c. 2000. From 2000 onward the frequency of tsunami appears to increase, but this is largely due to an increase in data collection from tide gauges. The tide gauges record fluctuations caused by tsunami that would not have been noticed by human observation alone. In the period from c. 1970–2000, the database contains a mixture of tide gauge records and newspaper or written reports of tsunami. Prior to c. 1970 the historical tsunami database is largely reliant upon newspaper reports and writings in historical documents, which are naturally skewed towards recording tsunami that were damaging or very noticeable.



**Figure 3.1** Distribution of tsunami that have been recorded on New Zealand shorelines from 1835–2011. Data from the New Zealand historical tsunami database (Downes et al., unpublished data). This analysis excludes tsunami of low validity ranking (see Table 3.1).

Of the 80 tsunami to have affected New Zealand in historical times (post-1835):

- 27 were from distant sources (> 3 hours tsunami travel time, Figure 3.2)
- 12 were from regional sources (1–3 hours tsunami travel time)
- 28 were from local sources (< 1 hour tsunami travel time)
- 13 were from unknown sources.

The ring of subduction zones around the Pacific Ocean is responsible for most of the distantsource tsunami to affect New Zealand (Figure 3.2). Tsunami from the South American margin along Peru and Chile are most frequent, but New Zealand is also affected by tsunami from the Alaska-Aleutian margin, and the Kamchatka-Kuril-Japan margin, and the south Pacific subduction zones of the Solomon Islands and the Tonga-Kermadec trench (Figure 3.2). Tsunami generated at the Sumatra subduction zone (the M<sub>w</sub>9.3 Indian Ocean tsunami, 2004) and by the Krakatau volcanic eruption (1883) were recorded in New Zealand but did not cause any significant damage.

Regional-source tsunami are typically from the Puysegur trench, southwest of New Zealand and the Tonga-Kermadec trench, northeast of New Zealand (depending on the distance from New Zealand, tsunami generated on the Tonga-Kermadec trench can be classified as distant- or regional-source). Local-source tsunami are predominantly associated with upper plate faults or the plate interface along the Hikurangi subduction zone or the Fiordland-Puysegur subduction zone. Exceptions to this are tsunami that were generated by the  $M_W$ 6.4 1922 Motunau (north Canterbury) earthquake, the  $M_W$  7.3 Buller earthquake and landslide-generated tsunami on Lake Taupo in 1846 and 1910.

Figure 3.3 shows the distribution of locations around New Zealand that have been impacted by historical tsunami, and the proximity of the tsunami source (note that a single tsunami can affect multiple points along the coastline, so there are many more data points on Figure 3.3 than there were individual tsunami). The Northland to Bay of Plenty region has been dominantly affected by distant-source tsunami, with rare regional-source tsunami. The East Coast of the North Island from East Cape to Wellington has been affected by both distant-and local-source tsunami. The northwest Nelson area and Fiordland coast have mostly been impacted by local-source tsunami. All other areas, including the Chatham Islands have been dominantly affected by distant-source tsunami (Figure 3.3).



**Figure 3.2** The distant source areas that have generated tsunami that have affected the New Zealand coastline (1835–2011). Each yellow dot represents an event. The dots are in the approximate source location but they do not accurately represent earthquake epicentres. All distant sources were earthquakes, except Krakatau, which was a volcanic eruption. Note that tsunami triggered at the Tonga-Kermadec Trench may be classified as regional if their source is close to New Zealand.



**Figure 3.3** The proximity of tsunami sources for tsunami that have affected the New Zealand coastline since 1835 (Downes, unpublished data). Distant source: > 3 hours travel time; regional source: 1–3 hours travel time; local source: < 1 hour travel time. Note that an individual tsunami event may affect the coastline at multiple points, so each data point does not represent a separate event. Points overlap in some locations but are intended to give a general impression (specific location details are recorded in the database).

Most historical tsunami that have affected New Zealand shorelines have been caused by earthquakes (Figure 3.4). Of the 80 tsunami to have impacted New Zealand, 44 were definitely caused by earthquakes, 9 were caused by earthquakes and associated landslides, and 8 tsunami were caused by landslides alone. One tsunami was a meteo-tsunami caused by the Krakatau eruption in 1883. Meteo-tsunami are generated by air-pressure disturbances (e.g. the blast from the Krakatau eruption); there are three other suspected meteo-tsunami in the historical database. Five tsunami have no known cause, and the remainder of the tsunami (13 events) have uncertain causes but are suspected to have resulted from earthquakes and/or landslides. The distribution of tsunami generated by different causes seems to have no particular spatial pattern (Figure 3.4).



**Figure 3.4** The causes of tsunami that have affected the New Zealand coastline since 1835 (Downes, unpublished data). Note that an individual tsunami event may affect the coastline at multiple points, so each data point does not represent a separate event. Points overlap in some locations but are intended to give a general impression (specific location details are recorded in the database).

### 3.2 LARGE HISTORICAL TSUNAMI

According to the historical tsunami database (Downes, unpublished) the five largest historical tsunami in New Zealand were generated by: the  $M_W 8.2$  Wairarapa earthquake in 1855, a  $M_W 7.1$  earthquake 50 km offshore of Gisborne in March 1947, and distant earthquakes in South America in 1868, 1877 and 1960 (Figure 3.5).



**Figure 3.5** Estimated tsunami runup values for the five largest tsunami in New Zealand between 1835 and 2011. Note the scale varies between boxes. Local-source tsunami have high runups, but are typically smaller in spatial extent, distant-source tsunami have widespread effects but lower runups. The runup values are from the New Zealand historical tsunami database (Downes et al., unpublished data), and include a combination of actual measured values and estimates based on descriptions given in newspaper (and other) reports.

#### 1855 Wairarapa Earthquake

The 1855 earthquake, which ruptured the Wairarapa Fault east of Wellington, generated a tsunami with a maximum known run-up of 10 m at Te Kopi in eastern Palliser Bay and up to 4-5 m in several locations in Wellington and along the northern Marlborough coast (Figure 3.5). The Rongotai isthmus and Miramar were reportedly covered in water to about one metre depth, rushing in from Lyall Bay and from Evans Bay. In Lambton Quay, the tsunami was 2–2.5 m high, washing into shops that fronted on to what was then the beach. Waves swept around Wellington Harbour and through Cook Strait for more than 12 hours, being observed as far south as the Clarence River Mouth and at least as far north as Otaki, where the run-up was probably about 2-3 metres. It is estimated that at least 300-500 km of coastline was affected, with run-ups of 1 m or more. The first waves arrived within minutes in Wellington and within an hour of the earthquake at Otaki and Marlborough. While submarine and coastal landslides may have contributed to the tsunami, the coseismic displacement of the sea bed, by as much as 6 m vertically upward near Turakirae Head on the south Wellington coast, was probably the main cause. Tides continued to be disturbed for the following week, possibly because of large aftershocks, perhaps with accompanying landslides.

#### 1868, 1877 & 1960 South American Earthquakes

Three tsunami, in 1868, 1877, and 1960 generated by great ( $M_W \ge 8$ ) earthquakes in South America caused widespread damage and disruption along the east coast of the North and South Islands and in the Chatham Islands (Figure 3.5). The 1868 tsunami caused the only death in New Zealand attributable to tsunami since European settlement. The tsunami was generated by a magnitude ~M9.1 earthquake off southern Peru/northern Chile. The greatest near-source run-up recorded for the 1868 tsunami was 18 m (Integrated Tsunami Database for the World Ocean). In New Zealand, run-ups of 1-4 m occurred on the mainland, and up to 10 m in the Chatham Islands. Considerable damage to houses, boats, shops, wharves, jetties, and boatsheds occurred along the whole eastern seaboard. At Tupuanga (Tupuangi) on the northwest coast of Chatham Island, the dwellings of an entire Māori village were washed away, the 60-70 residents escaping after the first of three large waves reached the floor of their dwellings. The tsunami severely impacted Great Barrier Island, eastern Bay of Plenty, Napier, Canterbury (especially Banks Peninsula), and Oamaru. It even reached the West Coast with waves of 1-2 m reported in Westport. Damage was more limited than could have been expected because the largest waves of the tsunami arrived within an hour or two of low tide at locations south of Napier. Smaller waves that occurred near high tide also caused damage.

The 1877 tsunami was caused by a magnitude ~M9 earthquake off northern Chile, about 400 km south of the source of the 1868 event. The tsunami was up to 21 m high near its source, but in New Zealand the effects were generally not as extensive or as well recorded in historical documents as the 1868 tsunami. Nevertheless, the tsunami had peak run-ups of 3.5 m. Many of the places strongly affected in 1868 were again affected in 1877, but there were some notable differences showing the effect of the source location (Figure 3.5). The tsunami was again evident for several days, and again damage was limited by the largest waves arriving at or near low tide along a large part of the east coast.

The 1960 tsunami was generated by a massive,  $M_W$ 9.4–9.6 earthquake in the subduction zone off central Chile. It was the largest earthquake in the 20th century. According to the Integrated Tsunami DataBase (ITDB), it caused a large local tsunami (maximum run-up 25 m) that resulted in US\$550 million in damage and 1,000 deaths. Another US\$24 million in

damage and 61 deaths occurred in Hawaii, and in Japan the waves were more than 6 m high and caused 199 fatalities and US\$50 million in damage. There is as yet no estimate of the cost of the damage in New Zealand. As with the 1868 event, run-ups of 1–4 m occurred along the whole eastern seaboard from Northland to Southland, and in the Chatham Islands (Figure 3.5). In places, some of the largest waves of the tsunami arrived within an hour or two of low tide, particularly in the lower half of the North Island and northern half of the South Island. Considerable damage was done to houses, boats, shops, wharves, jetties, port facilities, and boatsheds, as well as threatening the lives of several people in Hawke's Bay, Gisborne and Banks Peninsula.

## 1947 Gisborne Earthquakes

In March 1947, a 120 km long stretch of coast, from Mahia Peninsula northwards, was struck by a tsunami, 30 minutes after a moderately felt earthquake. The earthquake was located about 50 km offshore from the coastline north of Gisborne and 10–15 km west of the Hikurangi Trough. Although described by some as severe and prolonged, the earthquake was not widely felt along the nearest coast and the shaking caused no damage. The maximum intensity of MMI 4 (on the Modified Mercalli Intensity scale) is considerably less than the expected intensity for the earthquake's moment magnitude  $M_W7.0-7.1$  and surface wave magnitude  $M_S7.2$ , and was even somewhat low for its local magnitude of  $M_L5.9$ . This type of earthquake is called a tsunami earthquake. Tsunami earthquakes are characterised by a slow rupture pattern and produce tsunami of greater size than expected, given the earthquake magnitude ( $M_S$ ).

The March 1947 tsunami was not only observed along the coastline from Mahia Peninsula to Tokomaru Bay (Figure 3.5), but also probably at Waitangi, and possibly at Tuapeka, in the Chatham Islands. The maximum run-up height of the March 1947 tsunami was ~10 m at a near-deserted beach about 20 km north of Gisborne (Figure 3.5). Here, the 16 m span wooden bridge on the main road near Pouawa was swept 800 m inland and all except one room of the only house nearby was destroyed, with the five occupants surviving. The Tatapōuri Hotel and other houses were damaged further south and near Mahia. Another tsunami earthquake ( $M_L 5.6$ ,  $M_S 7.2$ ,  $M_W 6.9$ –7.1; Doser and Webb, 2003) in May 1947 in a similar source area to the March event produced a tsunami that again impacted the Gisborne region coastline. Estimated runups of up to 6 m occurred at Waihau Bay and up to 5 m at Tolaga Bay; minor damage was reported.

The tsunami earthquakes of March and May 1947 have been the subject of recent investigation by GNS Science (Bell et al., 2009; Wang et al., 2009). Bell et al. (2009) proposed a source model for the March 1947 event involving rupture on or near a subducted seamount located on the shallow part of the plate interface. This unusual situation produced an anomalously large tsunami because the physical presence of the seamount promoted shallow rupture (hence more deformation of the seafloor), the concave profile over the seamount focussed the tsunami waves toward a narrow stretch of coastline, and the slow rupture occurred at a similar speed to the tsunami wave propagation, resulting in water "piling-up" and amplifying the tsunami. Tsunami earthquakes similar to the 1947 events are a problem for public tsunami hazard awareness because the relatively low severity of ground shaking associated with such earthquakes may not alert people to the need to evacuate, yet such earthquakes can produce anomalously large tsunami with short travel-times, so self-evacuation is the best form of mitigation.

Further discussion of the 1947 earthquakes appears in Sections 4.2.1 and Section 5.3.1.2.

### 1947 Aleutian Island Earthquake

A tsunami generated by the 1946 M7.4 earthquake in the Aleutian Islands caused minor damage and 1–2 m run-ups over limited parts of the New Zealand coastline. This event is important, as it is the only distant earthquake under  $M_W 8.5$  to have had a significant effect in New Zealand. The 1946 Aleutian earthquake was a tsunami earthquake similar to, but much more distant than, the 1947 Gisborne earthquakes described above. In the near-source area of the Aleutian earthquake, tsunami runups of up to 42 m were recorded, and far-field effects were felt across the Pacific, including 159 deaths in Hawaii (Okal and Hébert, 2007; Okal et al., 2003). In New Zealand the greatest impact on the main islands was along the east coast of the northern North Island (north of Whangarei), with water heights above sea level at the time reaching 1.2 m and causing minor damage to a bridge at Tutukaka. Great Barrier Island, Tolaga Bay and Stewart Island were also affected, with inundation heights of about 1–1.2 m above sea level.

### 3.3 RECENT TSUNAMI EVENTS 2005–2011

In the period from 2005–2011 (since the 2005 report of Berryman) there have been four tsunami to have affected New Zealand shorelines. These were the July 15<sup>th</sup>, 2009, Dusky Sound tsunami, the September 29<sup>th</sup>, 2009, South Pacific tsunami, the February 27<sup>th</sup>, 2010, Chile tsunami and the March 11<sup>th</sup>, 2011, Tohoku (Japan) tsunami. None of these tsunami caused any significant damage in New Zealand, but the potential threat level was high and Civil Defence warnings were issued in all cases.

## 15<sup>th</sup> July, 2009, Dusky Sound tsunami

On July 15<sup>th</sup>, 2009, a  $M_W7.8$  earthquake on the subduction interface beneath southern Fiordland created a tsunami that affected the near-field region of Dusky Sound (Figure 3.6) (Beavan et al., 2010a; Clark et al., 2011c; Prasetya et al., 2011). The earthquake ruptured an ~80 x 50 km<sup>2</sup> patch of the plate interface (Beavan et al., 2010a) and it was the largest earthquake in New Zealand since the 1931 Napier earthquake. It was only because it occurred in the remote and largely unpopulated area of Dusky Sound that there was not significant injury or damage caused by the earthquake and tsunami. The earthquake occurred at night (9:22 pm). The ensuing tsunami was recorded by eyewitnesses on boats within Dusky Sound, and recorded instrumentally by a DART buoy in the Tasman Sea and by tide gauges in New Zealand and Australia (Figure 3.6) (Prasetya et al., 2011). The largest instrumental record was 0.98 m peak-to-trough at Jackson Bay, south Westland, ~ 260 km from the epicentre (Figure 3.6).



**Figure 3.6** The location of the 2009 Dusky Sound earthquake and the locations where the tsunami was recorded (JB: Jackson Bay, PP: Passage Point, BH: Bluff Harbour, DI: Dog Island). PT: peak-to-trough tsunami measurement; Amp: tsunami amplitude.

Within Dusky Sound strong currents pulled on boat anchor lines, vessels hit the seafloor and turbulent water was observed, but no damage was caused. Post-tsunami field reconnaissance in Dusky Sound found very little disturbance to the shoreline, except in one location where the tsunami deposited shells and starfish 2.3 m above, and 8 m inland of the high tide line (Clark et al., 2011c). Tsunami modelling suggests the tsunami elevations in the near-field area would have been 0.5–2 m, with flow speeds of 3 m/sec (Prasetya et al., 2011). No tsunami damage was recorded, probably due to the absence of shoreline infrastructure and the coincidence of the tsunami with a low tide.

### 29<sup>th</sup> September, 2009, South Pacific tsunami

The September 2009 South Pacific tsunami was triggered by an earthquake doublet (two near-synchronous earthquakes) of  $M_W 8.0$  and  $M_W 8.1$  at the northern end of the Tonga Trench (Beavan et al., 2010b). Within 10–15 minutes of the earthquake, tsunami struck Samoa, American Samoa and the northern islands of Tonga, causing 189 fatalities and millions of dollars of damage.

The tsunami was first recorded in New Zealand 4.2 hours after the earthquake at Moturiki Island (Tauranga) and on the Chatham Islands (Figure 3.7; this analysis only includes sealevel gauge data from NIWA; the Geonet-operated tide gauge data has not yet been analysed for this event). Other parts of New Zealand took up to eight hours for the tsunami to arrive. The maximum peak-to-trough measurement was 0.89 m recorded at Kaingaroa on Chatham Island, the mainland of New Zealand generally saw peak-to-trough measurements of 0.3–0.6 m (Figure 3.7). The maximum height waves arrived 5–19 hours after the earthquake.

Within minutes of the earthquakes the Pacific Tsunami Warning Centre issued a tsunami warning for the wider South Pacific, including New Zealand.



**Figure 3.7** Maximum peak-to-trough measurements of the 29th September, 2009, South Pacific tsunami. This coverage only includes sea-level gauge data provided by NIWA (Rob Bell). The Geonet-operated tide gauges have not yet been analysed for this event.

## 27th February, 2010, Chile (Maule) tsunami

The 27<sup>th</sup> February, 2010, Chile tsunami (also called the Maule tsunami) was triggered by a  $M_W 8.8$  earthquake on the central Chile subduction zone. A ~500 km long segment of the plate interface slipped up to 15 m in the earthquake (Vigny et al., 2011). In Chile the death toll was 521, with 124 of those due to the tsunami. Hundreds of kilometres of the Chilean coast was affected by the tsunami, which had a maximum runup of 29 m (Fritz et al., in press). The eastern Pacific islands of the Juan Fernandez Archipelago and Easter Island also suffered tsunami damage.

In New Zealand the first tsunami waves were detected on tide gauges on the Chatham Islands 11.6 hours after the earthquake. The tsunami arrived on the east coast of the New Zealand mainland 13–14 hours after the earthquake and the first arrivals reached the west coast 16–18 hours after the earthquake. The maximum peak-to-trough values for the tsunami were 1.93 m (Lyttleton), 1.9 m (Chatham Islands, Figure 3.8) and 1.8 m (Gisborne, Figure 3.9). The maximum tsunami peak-to-trough heights arrived 20 hours after the earthquake at Lyttleton, 13.6 hours at the Chatham Islands, and 21.4 hours at Gisborne, so there was a time lag of 2 to 7 hours between the first tsunami waves and the largest tsunami waves. Gisborne had the highest amplitude measurement of 1.05 m.



**Figure 3.8** Illustration of the arrival of the 2010 Chile tsunami on the Chatham Islands tide gauge. The red line is the tide gauge reading and the black line shows fluctuations with the tidal effect removed. The green bar illustrates how the peak-to-trough value is obtained, and the blue line illustrates the amplitude measurement.



**Figure 3.9** Maximum peak-to-trough measurements of the 27th February 2010 Chile (Maule) tsunami at tide gauges around New Zealand. Data processed by Paul Lehmann.

Immediately following the Maule earthquake, a Pacific-wide tsunami warning was issued by the Pacific Tsunami Warning Centre. The New Zealand Ministry of Civil Defence & Emergency Management issued a tsunami warning for New Zealand. There was sufficient delay between the tsunami generation and arrival that the tsunami warnings were well-publicised in New Zealand and many coastal activities on the Sunday morning were cancelled.

# 11<sup>th</sup> March, 2011, Japan (Tohoku) tsunami

The Tohoku tsunami of 11<sup>th</sup> March, 2011, was triggered by a  $M_W$  9.0 earthquake at the Japan trench (Simons et al., 2011). In the near-field the tsunami was devastating, causing 15,700 deaths, with 4,600 people missing (as of 28<sup>th</sup> August, 2011, IOC-UNESCO Bulletin). The tsunami affected >2000 km of the Japan coastline; over 290 km of the coast had tsunami runups of >20 m, with a maximum runup of 39.7 m reported (Mori et al., 2011). The maximum instrumentally recorded amplitudes around New Zealand were up to about 1 m, and inundation of a small residential area occurred in Port Charles on the Coromandel Peninsula. The impacts of the tsunami on New Zealand are summarized in Borrero et al. (2012).

## 3.4 APPLICATION OF THE NEW ZEALAND HISTORICAL TSUNAMI RECORD

The written historical record covers only 165 years, and this is too short a time to reflect the full range of possible events that New Zealand might experience. Many large earthquakes have recurrence intervals in hundreds of years for the smaller events ( $M_W 8.5$ ) to several thousand years for the largest earthquakes (e.g.  $M_W 9.5$ ). Also, the historical record of small tsunami, or tsunami in the early years of our history, in sparsely populated places, or in remote places, such as Fiordland, is almost certainly incomplete. Nevertheless, New Zealand's historical tsunami database is one of the most comprehensive databases in the South Pacific.

The frequencies of occurrence for distant, regional and local source tsunami of specified runup somewhere in New Zealand based on the historical record are only first estimates, and may severely under- or overestimate the hazard. The historical record contains no local volcanic events, no great (M>8) local or regional plate interface earthquakes, and large earthquakes have occurred on only a small proportion of a large number of local sources. To calibrate frequency relationship requires multiple events of each type.

For risk management, and to provide all the necessary information for appropriate response in a tsunami warning situation, the historical record is at best indicative. It is, however, very useful for understanding the behaviour of tsunami in New Zealand, for public education, and for calibrating and validating numerical models. Paleotsunami can be used to supplement the historic record and New Zealand's paleotsunami record will be described in the next section.

# 3.5 PALEOTSUNAMI RECORDS

### 3.5.1 Description of paleotsunami

Paleotsunami are tsunami that occurred prior to written records. The evidence for their occurrence typically comes from the sediments and debris that they deposited in the coastal zone (tsunami deposits), occasionally from the marks of erosion they left in the landscape, or from archaeological sites and oral traditions (see, e.g., Atwater, 1987; Nanayama et al., 2003). Studies of coastal sediments can be used to build up a record of paleotsunami that inundated coasts in the past. Such records extend the tsunami record much further back in time than the historical and instrumental record, thereby improving our knowledge of tsunami hazard. Tsunami deposits, in addition to providing evidence for the occurrence of past tsunami inundation, can also provide information about their sources, and their frequency and magnitude in the following ways:

# Sources

- The aspect and length of coast over which a tsunami deposit is found can provide information about the direction and distance offshore of the source (and thereby whether it was a local, regional or distant event).
- The type of source can sometimes be inferred from co-existence of the tsunami deposit with physical evidence of deformation (e.g. subsidence and liquefaction features imply a local earthquake source).
- Correlation of the deposit with a known tsunami-causing event can be used to infer a source where high-resolution age control is available.

# Frequency

• Where a long geological record of tsunami deposits exists, it is possible to estimate recurrence intervals for paleotsunami. This type of information is particularly important where no large tsunami have occurred in historical times, but where large events are represented in the geological record frequently enough to suggest they will occur again in the future.

# Magnitude

- Sedimentary deposits are usually evidence of large paleotsunami because small tsunami are unlikely to leave obvious evidence of their occurrence in the geological record.
- The physical extent of tsunami deposits along and across coastal topography, as well as the height above sea level that deposits reach, provide minimum estimates for tsunami inundation distance and run-up height once any vertical tectonic movement is accounted for.

Although paleotsunami datasets have a unique contribution to make to tsunami hazard assessment, there are some major limitations that must be taken into account. For a start, paleotsunami datasets will always be incomplete because:

- Many paleotsunami are not represented in the geological record:
  - Not all tsunami leave a recognisable deposit.
  - Not all deposits are preserved for long periods of time.
- Many paleotsunami cannot be identified:
  - Not all deposits contain unique tsunami signatures.
  - Deposition is patchy, so evidence may be missing from a particular site.
  - Storm surge deposits may be misinterpreted as tsunami deposits and vice versa.

Despite the limitations of paleotsunami research, it has a vital role in identifying areas that have been impacted by tsunami. With detailed work, the source, magnitude and frequency of past tsunami can be elucidated. Paleotsunami research extends the record of events beyond the historical period, which is too short to capture the full range of events that can potentially affect the New Zealand coastline.

## 3.5.2 The New Zealand paleotsunami database

Paleotsunami research since the 2004 Indian Ocean tsunami has increased markedly, both internationally and in New Zealand. In New Zealand, paleotsunami have been identified at many places around the coastline as a result of targeted research by a few scientists. Identification of paleotsunami in New Zealand has provided evidence for the occurrence of past large events and has improved awareness of New Zealand's tsunami risk. Despite the recent increase in paleotsunami research, there is still a lack of coverage of key sites and little detail at many of the sites that have been studied. Paleotsunami research is time-consuming, so the focus of many studies has been on the initial identification of tsunami deposits. Additional work that is crucial for the assessment of tsunami source, frequency and magnitude, such as detailed mapping of the extent of the deposit, high-resolution age control, and investigation of multiple events at any one site, is yet to be carried out in many cases.

Recently a paleotsunami database<sup>19</sup> for New Zealand has been compiled by Goff (2008) and Goff et al. (2010c). This database describes 293 observations around the New Zealand coastline of likely-to-possible paleotsunami, which are related to between 35 and 40 paleotsunami (i.e., there are multiple observations that are attributed to the same event).

The New Zealand paleotsunami database contains a mixture of formally published research (peer-reviewed journal articles), non-formally published research (e.g. student theses, newsletters, reports, conference proceedings) and unpublished work (e.g. personal communications). A significant number of the database entries are based on the compilers' reinterpretation of published work which described deposits but did not specifically relate them to tsunami. These entries indicate further study is required to confirm or refute their interpretation as paleotsunami. Despite the variable quality of the source material and debate about interpreting paleotsunami deposits, the New Zealand paleotsunami database is a valuable resource for describing the distribution of features of potential-to-likely paleotsunami around the New Zealand coastline. Later in this section we will describe some recent and well-verified paleotsunami studies from around the New Zealand coastline.

### Geographic distribution and validity of paleotsunami deposits

Figure 3.10 shows the distribution of sites with paleotsunami evidence around the New Zealand coast. This figure shows the sites based on the validity of the evidence: larger dots in warm colours show the sites with excellent evidence for paleotsunami, smaller dots in cooler colours show sites with less certain evidence. In general a wealth of paleotsunami evidence exists along the east coast of the North Island (with the exception of the East Cape region), in the Wellington region, along parts of the east coast of the South Island, and on the Chatham Islands. Paleotsunami evidence also exists along the west coast of New Zealand but is less abundant than east coast records. The spatial distribution of paleotsunami evidence is approximately consistent with the distribution of historical tsunami observations (compare Figure 3.10 with Figure 3.3). It is also consistent with the location of New Zealand's most active offshore faults (the Hikurangi margin, Cook Strait, Bay of Plenty) and with the direction of tsunami coming from Pacific Rim subduction zone sources.

GNS Science Consultancy Report 2013/131

<sup>&</sup>lt;sup>19</sup> The Goff (2008) and Goff et al. (2010c) database includes information about historical events, but we exclude these from our description of the paleotsunami database because the Downes et al. (in prep) historical tsunami database (used in the previous sections) is a more comprehensive resource.



**Figure 3.10** The distribution of paleotsunami evidence around the New Zealand coastline from the New Zealand paleotsunami database (Goff, 2008; Goff et al., 2010c). The symbols are graduated according to the validity of the data.

### Ages of paleotsunami deposits

Ages of paleotsunami in the New Zealand paleotsunami database are shown in Figure 3.11 and Figure 3.12. Paleotsunami evidence is usually associated with an age range; sometimes relatively small (±50 years) but in most cases larger (>200 to 1000's of years). In Figure 3.11 and Figure 3.12 the midpoint of the age ranges plotted. Much evidence of paleotsunami (~160 entries in the database) is estimated to be related to tsunami that occurred between 1300AD and 1600AD (Figure 3.11). Tsunami within this age range are spread around the New Zealand coastline (Figure 3.12). However, much of this evidence probably relates to the same tsunami, particularly where the sites occur in proximity (Figure 3.12). In most cases the age estimate is a relative age and not based on absolute dating methods such as radiocarbon dating. In many instances, a deposit has been dated in one location then deposits nearby in a similar stratigraphic position are assumed to be of the same age. There are 52 entries in the database related to tsunami >2000 years old. The oldest estimated paleotsunami evidence is 2.51 Ma (Goff et al., 2012), and is inferred to be related to the Eltanin asteroid impact.

The age distribution of the evidence for paleotsunami deposits in New Zealand raises some interesting questions. In general one would expect the evidence for paleotsunami to increase toward the present day because more recent events would leave fresher, more distinctive deposits/erosion scars, the deposits would be shallower (i.e., closer to the surface so more accessible for geological studies), and there would be less opportunity for the evidence to be reworked. The fact that paleotsunami evidence in New Zealand is most frequent at ~1500AD and then tapers off is unusual. Many well-studied paleotsunami deposits are reliably dated to this time period so it is likely there was a cluster of large tsunami in the 14<sup>th</sup> and 15<sup>th</sup> centuries. However, there may be some inflation of the frequency at this time through inaccurate correlations. This could occur when one deposit that is well-dated is used for correlation to a number of other deposits in the same region without independent verification of age.



**Figure 3.11** Estimated dates of paleotsunami deposits through time. Paleotsunami deposits have been binned into age ranges according to their midpoint age. Note that multiple records of paleotsunami deposits may relate to the same tsunami event. This plot excludes information from Māori oral records that have an age range estimate of AD 1250–1800.



**Figure 3.12** The distribution of paleotsunami evidence around the New Zealand coastline from the New Zealand paleotsunami database (Goff, 2008; Goff et al., 2010c). The symbols are coloured according to the estimated date of the paleotsunami.

### Elevation of paleotsunami deposits

The present-day elevation of paleotsunami deposits can yield information about the minimum height of the tsunami. The elevation of discrete deposits or erosional features can be measured, and, once vertical tectonic movement is taken into account, the measurements can be used to estimate the minimum runup of the paleotsunami. The elevations of deposits in the New Zealand paleotsunami database are shown in Figure 3.13. The red to yellow dots show deposits above present-day mean sea level. The blue dots show sites where paleotsunami deposits have been found in cores. Many types of paleotsunami evidence do not have a measurable elevation (hollow squares, Figure 3.13); this evidence includes Māori oral records and secondary geomorphic changes such as sand dune remobilisation.

The highest inferred-paleotsunami deposits are 60–65 m and occur on the west coast of the Waikato region. While this is an alarmingly high elevation, the evidence consists of a pebble layer within sand dunes. This type of data is of debateable paleotsunami origin; it is often assigned a paleotsunami origin because few other explanations are satisfactory. It is however unsatisfactory to conclude proof of a paleotsunami origin through lack of alternative

explanations, but to date there has been little rigorous investigation of such pebble layers to understand their mechanism of deposition.

Most paleotsunami deposits are between 0–5 m above sea level. While this does not seem high, it should be remembered that the deposit elevation gives a minimum wave runup and the tsunami may have been many metres higher. Many of the most reliable indicators of paleotsunami have been found in core samples, e.g. Cochran et al. (2005); Goff et al. (2010b); Nichol et al. (2007b). The core samples usually come from back-barrier wetlands and extend to below mean sea level, so they are not a useful proxy for runup measurements (but they can reliably indicate minimum inundation distance).



**Figure 3.13** The distribution of paleotsunami evidence around the New Zealand coastline from the New Zealand paleotsunami database (Goff, 2008; Goff et al., 2009b). The symbols are coloured according to the elevation at which the paleotsunami evidence is found. Empty square boxes indicate evidence that lacks elevation data, for example Māori oral records, or evidence of secondary dune mobilisation.

## Inferred sources of paleotsunami

Due to the nature of the evidence it is difficult to confidently identify the source of a paleotsunami. Hence, we treat all causes in the paleotsunami database as "inferred". Techniques to identify the source of a tsunami include:

- Correlation to known historical earthquakes from other locations. For example, a paleotsunami deposit on the Chatham Islands has been correlated with a large earthquake in 1604 in South America (Goff et al., 2010a).
- Association with evidence of a paleoearthquake, volcanic eruption or landslide. For example, two paleotsunami sand layers found in wetland cores in northern Hawke's Bay are inferred to be related to paleoearthquakes because they occur in association with a sudden (coseismic) subsidence event (Cochran et al., 2005). Correlation of paleotsunami deposits in Abel Tasman National Park and Kapiti Island to the c. AD 200 Taupo eruption has been suggested by Lowe and de Lange (2000).
- Extensive mapping of a paleotsunami deposit such that various source models can be tested. An example of this has been attempted by Goff et al. (2010b), who compiled evidence for three paleotsunami to have impacted the northern half of the North Island. They inferred that the distribution of two paleotsunami deposits matched a Tonga-Kermadec trench earthquake source, and one potentially matched an earthquake source in the Fiji region or a large volcanic eruption in the New Hebrides island group. This technique requires extensive mapping and accurate dating which has rarely been carried out.

Figure 3.14 shows the inferred causes of paleotsunami evidence around the New Zealand coastline. Most paleotsunami evidence is related to earthquake sources (64% of entries in the database). There are smaller numbers of paleotsunami related to land or submarine landslides and volcanic eruptions, and one inferred asteroid impact (at 2.51 Ma) and many examples of paleotsunami evidence with no inferred cause.



**Figure 3.14** The distribution of paleotsunami evidence around the New Zealand coastline from the New Zealand paleotsunami database (Goff, 2008; Goff et al., 2009b). The symbols indicate the inferred cause of the tsunami.

### 3.5.3 Recent paleotsunami research 2005–2011

Since the 2005 review of tsunami hazard and risk in New Zealand (Berryman, 2005), the amount of paleotsunami research in New Zealand has increased in line with a global increase following the 2004 Indian Ocean tsunami. Techniques have improved due to better knowledge of modern and prehistorical tsunami deposits. In this section we review some of the recently published research (Figure 3.15).



**Figure 3.15** The locations of recent paleotsunami studies in New Zealand from 2005–2011. The black lines onshore represent the active faults of New Zealand. The red line shows the Hikurangi-Kermadec trench.

**Northern Hawke's Bay:** Studies by Cochran et al. (2006), and Cochran et al. (2005) have found evidence for two paleotsunami associated with coseismic subsidence events. The tsunami deposits were found in cores from coastal wetlands, and the deposits were characterised by anomalously coarse grain size, chaotically mixed sediment, sharp lower contacts and fossils derived from seaward of the site. Both tsunami deposits were associated with a change in the wetland environment due to a large earthquake. A significant amount of radiocarbon data was obtained and the paleotsunami were dated at c. 7100 and 5550 years BP. By tracing out the extent of earthquake-related deformation, it was inferred the most likely source of the earthquake was a large event (~M7.9) on the Hikurangi margin plate interface. The Cochran et al. (2006) study is significant because it is the first to provide good evidence that the Hikurangi subduction zone has produced large earthquakes and tsunami in the past. Work is continuing on this project at a new location in southern Hawkes Bay which also shows evidence of paleotsunami deposits.

**Okarito Lagoon, Westland:** Nichol et al. (2007a) obtained sediment cores from Okarito Lagoon on the west coast of the South Island that suggest occurrence of a paleotsunami c. AD 1320–1495. The evidence for the paleotsunami consists of coarser-grained sediment overlying a buried soil and it is associated with an increase in lagoon salinity. Some key diagnostic criteria (e.g. evidence of transport from a seaward environment) are absent from the inferred paleotsunami deposit, and there are several other scenarios that could explain

an increase in lagoon salinity. The age of the deposit overlaps with the timing of a large earthquake on the Alpine Fault. Therefore Nichol et al. (2007a) infer the cause of the subsidence and tsunami to be an Alpine Fault earthquake.

**Great Barrier Island:** A sediment core from a back-barrier wetland at Harataonga Bay, Great Barrier Island, contains evidence of a paleotsunami at c. 3000 years BP (Nichol et al., 2007b). The paleotsunami deposit is characterised by its coarser grain size and unusual magnetic properties (which suggest the sand came from a seaward source). Nearby, earlier work had identified a gravel layer within sand dunes that had also been inferred to be a paleotsunami deposit (Nichol et al., 2003). While neither deposit in isolation is unequivocal evidence for paleotsunami, their proximity does suggest a common source, thus making paleotsunami a more likely cause. The significance of the Great Barrier Island studies is that the gravel deposits reach up to 14 m above sea level, and the wetland at Harataonga Bay has a stationary 15 m foredune, implying the paleotsunami had a runup of >15 m. Nichol et al. (2003) speculate the source of the tsunami could have been an earthquake along the Hikurangi subduction zone or a volcanic eruption along the Kermadec arc. The heights are consistent with modelling of an earthquake on the Tonga-Kermadec subduction zone (Power et al., 2012; Lane et al., 2012).

**Otago coast:** The geomorphology of Blueskin Bay and Long Beach, north of Dunedin, was studied by Goff et al. (2009a). Goff et al. (2009a) identified coastal landforms consistent with formation due to erosion and scouring by tsunami (e.g. scour fans and sand dune breaches). At Long Beach, a sand layer was also identified and tentatively assigned a tsunami origin, though it was not characterised in sufficient detail to be considered a strong case for a paleotsunami deposit. Correlations to dated Māori artefacts suggest the inferred paleotsunami occurred sometime in the 14<sup>th</sup> to early 15<sup>th</sup> century; they suggest a Puyseygur subduction zone earthquake (Goff et al., 2009a).

**Chatham Islands:** The sedimentary evidence for the AD 1868 tsunami (see section 3.1) on the Chatham Islands was investigated by Goff et al. (2010a). At the same location where 1868 sediments were identified, an older, thicker sandsheet was identified. This sandsheet contained many indicators of tsunami deposition (e.g. erosional base, fining upward, rip-up clasts, unusual paleoecology). The changing pollen assemblages placed some age constraints and Goff et al. (2010a) correlated the event to an AD 1604 earthquake in Peru. Interestingly, a study at another location on the Chathams Island by Nichol et al. (2010) found evidence of the AD 1868 tsunami but no others, even though they had a sedimentary core dating back to 43,000 years BP.

*Kaituna Bay, Northland:* Cores from a wetland at Kaituna Bay contain evidence for three paleotsunami in the past 8000 years (Goff et al., 2010b). The paleotsunami deposits were well-characterised using multiple techniques (e.g. grain size, geochemistry, microfossils) but relatively poorly dated. Goff et al. (2010b) correlated the paleotsunami deposits at Kaituna Bay to other inferred paleotsunami deposits of a similar age around the North Island and used the spatial distribution of the deposits to infer the earthquake source. The events at c. 6500 and 2800 years BP were inferred to be from a Tonga-Kermadec trench earthquake source. The event at c. 1450AD had a wider distribution than the earlier events and was also correlated to inferred paleotsunami deposits in the northern west coast of the North Island. Goff et al. (2010b) suggested a volcanic eruption in the New Hebrides island group may have been the source of that event. The dating of older paleotsunami deposits at Kaituna Bay (and those around the North Island that they were correlated to) has large uncertainties (up to  $\pm$  3000 years). Further radiocarbon dating (with lower uncertainties) is necessary to provide

confidence that the regionally-distributed paleotsunami deposits were from a single (and very large) event.

*Wairarapa coast:* The east coast of the Wairarapa region from Cape Palliser to Cape Turnagain displays a sequence of uplifted Holocene marine terraces. In a study of the ages of the marine terraces (which indicate the timing of past large earthquakes along this coast) Berryman et al. (2011) found evidence of paleotsunami. Thirty-five radiocarbon ages were collected from the marine terraces and approximately 20% of these were anomalously young for their elevation. The anomalously young samples on high terraces often coincided with the age at which lower terraces had been uplifted. It was suggested by Berryman et al. (2011) that the young samples represent paleotsunami deposits that were emplaced by tsunami triggered by the same earthquakes that uplifted younger terraces. The research into the inferred paleotsunami deposits has not yet been comprehensive enough to be confident of their origin, but the Berryman et al. (2011) study corroborates previous paleotsunami studies along the Wairarapa coast (e.g. Goff et al., 2004).

**Big Lagoon, Blenheim:** The Big Lagoon area near Blenheim is the subject of an ongoing study examining evidence for earthquake-related subsidence. Multiple cores have been taken from the lagoon margins and these cores show evidence of paleotsunami deposits. Clark et al. (2011a) identified a sand layer at 3.1 m depth with anomalous microfossil assemblages and a chaotic, poorly sorted sedimentology. It was not well-dated (between 2,000–7,000 years) but it will be the subject of further studies. More recently Clark et al. (unpublished data) have identified a sand layer in Big Lagoon containing evidence for landward-transported marine microfossil assemblages, dated at c. 800 years BP. Like the northern Hawke's Bay study sites of Cochran et al. (2005, 2006) and the Wairarapa coast (Berryman et al., 2011, Goff et al., 2004), the Big Lagoon site is significant because it may hold evidence of paleotsunami related to large subduction earthquakes along the Hikurangi margin (Figure 3.15).

### 3.5.4 Summary of paleotsunami in New Zealand

The New Zealand paleotsunami database (Goff, 2008; Goff et al., 2010) is a valuable and comprehensive resource documenting evidence for paleotsunami in New Zealand. It is a vast improvement on the state of paleotsunami knowledge before 2004. The distribution of paleotsunami evidence around the New Zealand coast resembles the distribution of historical tsunami, and as such, it highlights the areas of the coastline most vulnerable to tsunami hazard. The age of the paleotsunami deposits requires significant improvement; most entries in the database are poorly dated or not dated at all. The apparent increase in paleotsunami around AD 1500 is unusual and deserves further investigation. If it represents a real clustering of tsunami events the cause needs to be understood. The elevation of the paleotsunami deposits can be used to estimate a minimum runup height for paleotsunami. However, some of the best evidence for paleotsunami comes from submarine cores, while some of the highest elevations are from pebble layers which have a tenuous association to paleotsunami. There are some examples where sufficient research has been undertaken to determine the source of the tsunami that deposited a paleotsunami deposit or left other evidence (e.g. erosion/oral record). However, for most entries in the database the tsunami source is only inferred, hence the sources are not reliable.

Age and runup estimates in the New Zealand paleotsunami database are still too scattered and uncertain to give accurate magnitude-frequency relationships. However, the paleotsunami database is a valuable resource for assessing tsunami hazard in New Zealand because our historical record is far too short to capture the range of tsunami that could potentially affect New Zealand shores. As paleotsunami research continues to increase beyond "reconnaissance-level" studies and into detailed multi-proxy, multi-site investigations, the reliability and quality of the database will improve substantially.

Recent paleotsunami studies in New Zealand have covered a wide range of sites, from areas of low seismicity (Otago, Northland), areas of high seismicity along the Hikurangi margin (Hawkes Bay, Wairarapa, Blenheim), areas of high exposure to South American tsunami (Chatham Islands) and regions of high onshore seismicity but with relatively few offshore tsunami sources (Westland). The studies are becoming more rigorous through the use of multi-proxy techniques and an increasing knowledge of the signatures of tsunami deposition and erosion. Increasingly, sufficient data is being gathered so that the tsunami source can be identified. A future challenge will be to bring our knowledge of paleotsunami up to a standard where source models can be reliably calibrated using the inland extent and elevation of paleotsunami deposits, thus ensuring inundation models and tsunami evacuation zones are dependable.

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