Tsunami hazard for the Bay of Plenty and eastern Coromandel Peninsula

NIWA Client Report: HAM2004-084
June 2004

NIWA Project: BOP04216

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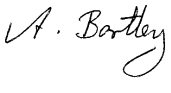

Executive Summary

Environment Bay of Plenty (EBOP) and Environment Waikato (EW) joined together to set up a three-year Joint Tsunami Research Project to assess the tsunami hazard and associated risk for the eastern seaboard of the Waikato/Bay of Plenty region from Colville Channel to East Cape.

The Year 1 phase was carried out by GeoEnvironmental Consultants Ltd, involved field investigations of paleo-tsunami deposits through the collection of sediment cores and partial laboratory analysis. The methodology and interim results were reported in Goff (2003).

For the Year 2 phase, NIWA was contracted by Environment Bay of Plenty, with GeoEnvironmental Consultants Ltd and the Institute of Geological & Nuclear Sciences Ltd (GNS) as sub-contractors.

The Year 2 (2003/2004) Contract Brief for the Joint Tsunami Research Project was:

- Continue with line of enquiry from Year 1 on the paleo-tsunami work carried out by GeoEnvironmental Consultants Ltd, but focus on the detail from two of the existing cores in peat/sand where the chances of getting an intact record are higher e.g., Waihi Beach. Split open promising cores and carry out geo/radio/chemical analyses;
- Interpretation of core data in context of BOP/Coromandel geology, volcanism and plate tectonics;
- Update and compile a historical catalogue of tsunami events for BOP/Coromandel;
- Integrate the historical and paleo-tsunami catalogues, with input from previous steps, keeping to a geological timeframe of several thousand years only;
- Investigate existing sources of historical sea-level records and resonance modelling work to identify hot-spots for remote tsunami in BOP/Coromandel region, and also investigate potential upper limits to storm-tide/wave run-up associated with coastal barriers like Ohiwa, Jacob's Creek, Waihi Beach to aid interpretation of cores and put tsunami events in context of sea-inundation hazards;
- Write a Year 2 draft report by 20 June 2004 that ties together the above tasks, integrating the results in the context of an overview of the "potential" for each type of geo-source, both remote and local, to cause a hazard threat to BOP/Coromandel coastlines.

The development of a credible tsunami hazard profile for the Bay of Plenty and eastern Coromandel has been undertaken by combining data and information from distinctly different sources. These include sea level and tsunami run-up data, eyewitness accounts, marine geophysical surveys, paleo-geological investigations of undisturbed sediment cores inland from the coast and numerical modelling of tsunami resonance behaviour in the overall region. This report describes the tsunami hazard profile for the Bay of Plenty and Eastern Coromandel Peninsula undertaken in Year 2 for the Joint Tsunami Research Project.

Tsunami: causes and categories

The word *tsunami* is used internationally, and is a Japanese word meaning "harbour wave or waves". They are generated by a variety of geological disturbances, particularly large seafloor earthquakes, submarine landslides (which may be triggered by an earthquake), volcanic eruptions (e.g., under-water explosions or caldera (crater) collapse, pyroclastic flows[1] and atmospheric pressure waves), large coastal-cliff or lakeside landslides, and very occasionally a meteorite (bolide) splashdown.

In each case, a large volume of water is disturbed suddenly, generally affecting the whole water column from the floor of the ocean to its surface, creating a train of waves radiating outwards (similar to the wave train produced by a pebble thrown into a lake) until the waves either dissipate or they collide with a shoreline. Tsunami waves can arrive at nearby shores within minutes, or travel across the deep ocean basins at speeds in excess of 500 km/hr. Very large sources (disturbances) are required to cause tsunamis that are damaging at great distances from the source. On the other hand, tsunamis that are generated locally (i.e., near our shores) do not need such a large disturbance to be damaging.

Tsunamis can be classified into categories either by the distance from their source to the area impacted, or more relevant for emergency management purposes, the travel time to the impacted area and the length scale of impact. For this report, three categories are defined:

- local source/local impact event (within say 30 to 60 minutes travel time and affecting several 10's of km of coast);
- regional source/regional impact event (within 3 hours travel time and likely to affect most of the Bay of Plenty and eastern Coromandel);
- distant (remote) source/national impact event (longer than 3 hours travel time and likely to affect several regions).

Paleo-tsunami record

Given that the post-European historic record is relatively short (160 years), geological field investigations and geo-chemical analysis of sediment cores opens up the possibility of detecting, interpreting and dating large paleo-tsunami events to extend the tsunami hazard record for the region. Year 1 of the Tsunami Hazard Study focused on selecting and obtaining cores from potentially undisturbed sites in the Bay of Plenty and Eastern Coromandel (Goff, 2003).

Locations for the paleo-tsunami field sites investigated were: Otama Beach (near Whangapoua); Waihi Beach, Ohiwa Harbour; and Jacobs Creek (between Waiotahi Estuary and Opotiki). Evidence has also been gleaned from additional sites that have been investigated in previous paleo-tsunami studies in both the eastern Coromandel and the Bay of Plenty (Goff, 2002a, b).

[1] A pyroclastic flow is a ground-hugging avalanche of hot ash, pumice, rock fragments, and volcanic gas that rushes down the side of a volcano at 100 km/hour or more, and can have temperatures greater than 500°C. In a coastal setting, such flows can disturb the surface waters causing a tsunami.

By combining detailed visual, geological, geochemical and radio-carbon analyses of the sediment cores and expert interpretation of the results based on the geological context of the Bay of Plenty and Coromandel region (e.g., volcanism, tectonic uplift, Holocene sedimentation), a record of probable paleo-tsunami events has been developed.

In summary, over the past 4000 years a total of two major regional paleo-tsunami events have been recorded in sediment cores—one in AD1302–AD1435 (with some evidence for two separate major events in this period) and an earlier event at 2500–2600 years BP. There is also evidence in various sediment cores that up to four localised paleo-tsunami events have occurred in the Bay of Plenty—in AD1600–AD1700 (local subsidence a factor?), AD1200–AD1300, 1600–1700 years BP, and 2900–3000 years BP.

A key point of these paleo-tsunami investigations is that the resolution used is only capable of identifying tsunami events with run-up height larger than 5 m (Goff, 2003). This lower cut-off limit arises from several factors including: a) a paucity of undisturbed sites due to on-going coastal development; b) the resolution of the methods used to detect and confirm paleo-tsunami deposits; c) sediment core locations are behind elevated sand-spit barriers and at least 250 m inland, with the paleo-tsunami deposits appearing to be undisturbed by storm surges and overtopping impacts. Any further estimation of tsunami run-up heights from paleo-tsunami deposits would need to be investigated using a tsunami wave model (with realistic land topography) once a credible source-generation scenario can be constructed.

Historical tsunami record

In historical times (since 1840), tsunamis are known to have affected places along the Bay of Plenty and Eastern Coromandel coastline at least eleven times. The historical eyewitness and newspaper accounts of the behaviour and impacts of these tsunami events are detailed in Table 1 of Section 3.

Information on the historical tsunami events is based on data and information built up over the last two decades, particularly from the University of Waikato (Earth Sciences Dept.) and the GNS Tsunami Database. The latter revises and updates the earlier databases with new accounts found as the result of recent comprehensive investigation of historical newspaper accounts.

Some recent tsunami events have been small (<0.5 m in wave height) and usually were only detected by sea-level gauges. Such small to very small events were usually not noticed prior to the installation of sea-level gauges, and hence it is probable that the Bay of Plenty has experienced many more of these small tsunamis, particularly from distant sources, than the historical tsunami database indicates.

The most substantial tsunamis to have affected the Bay of Plenty and eastern Coromandel areas in the last 160 years were generated by “remote” or distant sources. The largest tsunamis, in 1868, 1877 and 1960, were generated by very large earthquakes in the subduction zone along the Chile and southern Peru coastlines of South America—directly opposite and facing New Zealand’s eastern seaboard. A further event occurred in August 1883, probably generated by an atmospheric pressure wave from the Krakatau eruption in Indonesia. It caused run-up heights of up to 1.8 m in the Bay of Plenty–Coromandel region. In pre-European history, there are indications that a large earthquake off the

Cascadia region (east coast of Canada/US Pacific Northwest coast) in 1700 could have impacted New Zealand. Recent overseas model simulations of this event (using paleo-tsunami evidence, as well as Japanese historical records) show that the wave heights may have been substantial in some regions of New Zealand, possibly over 1 m in parts of the Bay of Plenty and Coromandel. Further modelling is required to confirm better estimates of the run-up in New Zealand.

Since European settlement around 1840, no "local source/local impact" or "regional source/regional impact" events are known to have affected the Study region. However, this is not unexpected as fault ruptures tend to have return periods of 100s to 1000s of years and volcanic eruptions, return periods of 1000s to 10,000 years or more.

Tsunami hazard profile (Bay of Plenty and eastern Coromandel)

Table E.1 summarises the past tsunami profile of the Bay of Plenty and eastern Coromandel by combining known historical tsunami events (back to 1840) with the paleo-tsunami events identified in this Study over the past 4000 years.

Table E.1: Summary of the known past tsunami events across both the Eastern Coromandel and Bay of Plenty region (combining historical records back to AD1840 and paleo-tsunami signatures back 4000 years). Note: y BP = years before present.

Tsunami run-up height (est.)	**<0.5 m***	**0.5–1 m**	**1–3 m**	**3–5 m**	**>5 m**
No. of events:	>6	1	4–5	?[#]	5 or 6
Year(s):	June 2001	Nov 1922	May 1960		*Regional-scale*
	July 1998		Aug 1883		AD1302–1435
	Oct 1994		May 1877		2500–2600 y BP
	June 1977		Aug 1868		*Local-scale*
	Jan 1976		1700?		AD1600–1700?
	Mar 1964				AD1200–1300
					1600–1700 y BP
					2900–3000 y BP

*Many events of <0.5 m run-up may have occurred, but gone unnoticed before instrumental sea-level records became available.

[#] No events >3 m run-up in the historical records, and paleo-tsunami analyses at this stage can only resolve events with >5 m run-up.

Based on these results, it would appear that the return periods for given tsunami heights for Tauranga Harbour (listed in Table 2 of the main report), are probably too long (e.g., 322-year return period for a 2.5 m high tsunami), especially if these Tauranga Harbour return periods are applied to the entire Bay of Plenty and eastern Coromandel coast. The inference that the return periods are probably lower for the open coast is based on: 1) the five or six local/regional events >5 m run-up height from the paleo-tsunami record that may have occurred over the past 4000 years; 2) that a further four historical events have produced run-up heights between 1 to 3 m since 1840; and 3) a further distant tsunami possibly reached run-up heights of 1 to 3 m in 1700. Further analysis is required to determine a realistic tsunami return-period profile for various sections of the region's coastal margin.

Potential local and regional tsunami sources

While the post-European historical record since 1840 contains no known tsunami events generated from local or regional sources that had run-up heights >1 m, the paleo-tsunami record contains at least five or six events, most of which may have been caused by such sources.

A comprehensive summary of sources that could potentially generate a tsunami event has been complied from previous and recent geophysical investigations including seafloor mapping and seismic profiling of faulting systems, underwater volcanoes and underwater landslides together with knowledge on past behaviour of volcanoes. On a national scale, Bay of Plenty and eastern Coromandel face quite a diverse range of potential sources for generating a tsunami locally (up to 30 to 60 minutes travel time to coast and local impact) or regionally (up to 2 to 3 hours travel time to coast and regional impact).

Potential tsunamigenic sources (with potential local and regional scale impacts) are:

a) Subduction interface earthquakes that occur in the Tonga-Kermadec-Hikurangi region associated with the Pacific/Australian plate boundary. This source occurs beneath the eastern margin of the North Island and the Kermadec Ridge, where the Pacific Plate underthrusts (subducts) to the west. It is not yet certain if the entire subduction zone is a potential tsunamigenic hazard;

b) There are many upper plate faults in the northern Hikurangi continental shelf margin from Mahia to Ruatoria, some of which may be capable of substantial tsunami generation south of East Cape. However, earthquakes in this region are unlikely to cause large tsunami impacts in the Bay of Plenty as coastal-trapped waves propagating northwards along the coast would be substantially dissipated as they moved around East Cape into the Bay of Plenty;

c) Landslide sources in the Hikurangi margin off East Cape include giant complex landslides such as Matakaoa and Ruatoria that are likely to be, but are not necessarily, triggered during large earthquakes. Such large events could have very long return periods of 10's–100's of thousands of years. However, smaller landslides are more likely within the Matakaoa complex and in the submarine canyons of Bay of Plenty. Further mapping of the Bay of Plenty continental margin is required to determine whether landslide scars are present and in what frequency. Modelling would be required to determine what dimensions and mechanisms of a landslide source would result in a tsunami inundation hazard along the Bay of Plenty coast;

d) Undersea volcanism in the Tonga-Kermadec system (and more distant) is another potential source of tsunamis. At least 23 submarine volcanoes, of the active southern Kermadec arc, occur within 400 km of the Bay of Plenty coastline, three of which (Rumble II West, Brothers and Healy) are silicic calderas. Of these, Healy is thought to have been formed by catastrophic submarine pyroclastic eruption. Larger volcanoes (e.g., Havre and Macauley) are known further north along the Kermadec Ridge volcanic arc, within 970 km of the coastline. Data assembled for this project indicate that one or possibly two paleo-tsunami events inferred for the Bay of Plenty and eastern Coromandel at around AD1302–1435, may be associated with

eruption and/or collapse of the Healy caldera in the Kermadec Ridge. One of the paleo-tsunami events occurs in association with Loisels Pumice, interpreted to be derived from the Healy caldera (Section 4). This event could perhaps coincide with collapse of the Healy caldera cone. There appears to have been an earlier event in the Loisels Pumice-related period (AD1302–AD1435) that may have been associated with the initial (or subsequent?) submarine eruptions at Healy. Seafloor multibeam mapping reveals many of the 23 southern Kermadec volcanoes undergo cycles of sector collapse. Whether such collapses are large single catastrophic events or small repetitive movements is presently unknown;

e) Regional active faults provide many candidate sources of tsunami for Bay of Plenty and eastern Coromandel. They include normal faults in the offshore Taupo Volcanic Zone, both on and off the continental shelf. The major zone of active rifting extends between Whakatane and Tauranga, with faults between Matata and Whakatane accommodating a significant proportion of the total crustal extension. The larger faults with significant seafloor traces include the Whakaari/White Island and Rangitaiki Faults in the offshore Whakatane Graben. Normal faulting in the Taupo Volcanic Zone rarely exceeds 2 m single event vertical displacement, but the larger boundary faults may be capable of larger surface ruptures. Whether fault rupture with modest displacement is capable of generating destructive tsunamis is uncertain, and requires numerical modelling of the seafloor disruption and the propagating tsunami wave;

f) Offshore volcanic sources in the Bay of Plenty and southeastern Coromandel, include Tuhua/Mayor Island and Whakaari/White Island. Whakaari/White Island has previously been discounted for tsunami generation potential due to its deep-water location and any tsunami produced is likely to propagate eastwards away from the coast. Forthcoming multibeam mapping of the Bay of Plenty continental shelf and slope around Whakaari/White Island will provide more updated information on the potential tsunami hazard from this source. However, for Tuhua/Mayor Island, previous modelling studies by the University of Waikato indicate that the credible pyroclastic eruptions of a "Mt St Helens" scale (1 km^3) could produce a tsunami that would impact an area from Tairua to Maketu, peaking at 0.5 m between Whangamata and Tauranga. An eruption ten times bigger with a pyroclastic flow of Krakatau scale (10 km^3) would peak at around 5 m at the coast. Recent geophysical data from Tuhua/Mayor Island indicates the last caldera collapse, associated with the largest eruption, occurred about 6,300 years ago and included the transport of large pyroclastic flows into the sea, which probably generated a tsunami. There is a possible link between Tuhua/Mayor Island pyroclastic flows entering the sea and the ~6300 yr BP event preserved in the sediment cores from Waihi Beach (Section 4.2). Numerous smaller submarine volcanoes occur on the Bay of Plenty continental shelf and slope closer to the coast (within 100–150 km) that also merit investigation;

g) Tsunamis generated by atmospheric pressure-waves or pyroclastic flows from large onshore volcanic eruptions in the Taupo Volcanic Zone (e.g., Okataina Volcanic Centre) or Mt Taranaki are another possibility. The potential for these is little known, but the direct volcanic impacts are likely to overwhelm the additional impact and consequences of any associated tsunami in the Bay of Plenty and eastern Coromandel.

Locations in the region vulnerable to distant and regional source tsunami

Incoming tsunami waves from a distant or regional source can "pick-out" and excite the natural resonant period of a harbour or bay, causing the wave to amplify in height and persist longer in certain areas compared with other parts of the coast. This pattern of more vulnerable areas due to resonance effects changes with the wave periods present within any given tsunami. This means different tsunami events may preferentially impact different areas to those impacted by previous events, especially if a distant South American tsunami is compared with a regional-impact tsunami event. However, combining results from resonance modelling for tsunami wave periods of 75 and 90 minute oscillations with historical accounts of tsunami damage and wave observations, has highlighted some areas of the Bay of Plenty and Eastern Coromandel region that are potentially more vulnerable than other areas. These are listed below, but only apply to "distant source/national impact" and "regional source/regional impact" tsunamis.

Highest vulnerability

- Open coast from Otama Beach to Port Charles and out to Great Mercury Island (especially Whangapoua embayment and Port Charles).
- Mercury Bay (especially Whitianga).
- Open coast between Mt. Maunganui/Mauao and Maketu (especially Kaituna River and Maketu).
- Open coast between Matata and Torere (especially river entrances e.g., Opotiki, Torere).
- Papatea and Whangaparaoa Bays near Cape Runaway.

Moderate vulnerability

- All other open coast areas.
- Tauranga Harbour?
- Ohiwa Harbour?

Further high-resolution modelling is required to ascertain the relative vulnerability of harbours, estuaries and river mouths to "distant" and "regional" tsunami sources. More geophysical information is required to rank the various possible sources of tsunami generation, and additional tsunami wave modelling is needed before relative vulnerabilities of areas in the region can be determined. However, accurate modelling of tsunami run-up behaviour and impacts along the coastal margin, including rivers and harbours and overland flows, will depend on the acquisition of high-resolution bathymetry and land topography.

Finally, a tsunami that is not amplified substantially by resonance may still be dangerous in all parts of the coast (e.g., a run-up of 1 m is considered dangerous, especially coinciding around high tide).

1. Introduction

1.1. The Brief

Environment Bay of Plenty (EBOP) and Environment Waikato (EW) joined together to set up a three-year Joint Tsunami Research Project to assess the tsunami hazard and associated risk for the eastern seaboard from Colville Channel to East Cape.

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b) Interpretation of core data in context of BOP/Coromandel geology, volcanism and plate tectonics.

c) Update and compile a historical catalogue of tsunami events for BOP/Coromandel.

d) Integrate the historical and paleo-tsunami catalogues, with input from previous steps, keeping to a geological timeframe of several thousand years only.

e) Investigate existing sources of historical sea-level records and resonance modelling work to identify hot-spots for remote tsunami in BOP/Coromandel region, and also investigate potential upper limits to storm-tide/wave run-up associated with coastal barriers like Ohiwa, Jacob's Creek, Waihi Beach to aid interpretation of cores and put tsunami events in context of sea-inundation hazards.

f) Write a Year 2 draft report by 20 June 2004 that ties together the above tasks, integrating the results in the context of an overview of the "potential" for each type of geo-source, both remote and local, to cause a hazard threat to BOP/Coromandel coastlines.

1.2. Report content

This report describes the findings of a tsunami hazard study for the Bay of Plenty and eastern Coromandel Peninsula undertaken in Year 2 for the Joint Tsunami Research Project. The paleo-tsunami fieldwork and the preliminary laboratory work that followed were previously discussed in the Year 1 report (Goff, 2003).

The primary focus for this second report is to define the overall tsunami hazard that potentially threatens the Bay of Plenty and eastern Coromandel region.

This report contains a profile of the tsunami hazard drawn from historical tsunami events, blended together with viable pre-historical tsunami events based on interpretation of sediment cores. This interpretation takes into account the context of past coastal evolution in the Bay of Plenty/Coromandel, regional geology, volcanism and plate tectonics.

The tsunami hazard for the region is described in terms of potential sources of tsunami generation, particularly focusing on the processes that could generate potentially more damaging local-source tsunami, as distinct from remote (distant) tsunami sources.

The tsunami risk (exposure, vulnerabilities, susceptibilities) to exposed coastal communities in the Bay of Plenty and eastern Coromandel region will be developed in the Year 3 study.

2. Tsunami—a natural hazard

Tsunami is one of New Zealand's underrated natural hazards. The last major tsunami to hit New Zealand shores was caused by the Chile earthquake of May 1960, some 44 years ago. As a result of this lengthy quiescent period, most people now have a low expectation that a tsunami will pose any danger in their lifetime, according to the recent 2003 National Coastal Community Survey (Johnston et al. 2003).

2.1. Definitions

The word *tsunami* is used internationally, and is a Japanese word meaning "harbour wave or waves". In the past, people called them "tidal waves" but this is a misnomer as they are not generated by tides. Instead they are generated by a variety of geological disturbances, particularly large seafloor earthquakes, submarine landslides (which may be triggered by an earthquake), volcanic eruptions (e.g., under-water explosions or caldera (crater) collapse, pyroclastic flows[2] and atmospheric pressure waves), large coastal-cliff or lakeside landslides, and very occasionally a meteorite (bolide) splashdown (de Lange, 2003).

In each case, a large volume of water is disturbed suddenly, generally affecting the whole water column from the floor of the ocean to its surface, creating a train of waves radiating outwards (similar to the wave train produced by a pebble thrown into a lake) until the waves either dissipate or they collide with a shoreline. Tsunami waves can arrive at nearby shores within minutes, or travel across the deep ocean basins at speeds in excess of 500 km/hr. Very large sources (disturbances) are required to cause tsunamis that are damaging at great distances from the source. The most common sources of these tsunamis are very large earthquakes along the subduction zones that ring the Pacific. However, meteorite impact and very large volcanic events are also possible sources. On the other hand, a tsunami that is generated locally (i.e., near the Bay of Plenty/Coromandel shores) does not need such a large disturbance to be damaging and life threatening, but it would only affect a limited area of the region's coast.

Tsunamis can be classified into categories either by the distance from their source to the area impacted, or more relevant for emergency management purposes, the travel time to the impacted area and the length scale of impact. For this report, three categories are defined:

- local source/local impact event (within say 30 to 60 minutes travel time and affecting several 10's of km of coast);

[2] A pyroclastic flow is a ground-hugging avalanche of hot ash, pumice, rock fragments, and volcanic gas that rushes down the side of a volcano at 100 km/hour or more, and can have temperatures greater than 500°C. In a coastal setting, such flows can disturb surface waters causing a tsunami.

- regional source/regional impact event (within 3 hours travel time and likely to affect most of the Bay of Plenty and eastern Coromandel);

- distant (remote) source/national impact event (longer than 3 hour travel time and likely to affect several regions).

Tsunami waves differ from the usual waves we see breaking on the beach or in the deep ocean, particularly in their length between wave crests. In a tsunami wave train, the distance between successive wave crests (or wavelength) can vary from several kilometres to over 400 km, rather than around 100 metres for waves at the beach. The time between successive tsunami wave crests can vary from several minutes to a few hours, rather than a few seconds. Out at sea, a tsunami would not be noticed because of these large wavelengths and small wave heights of less than 0.5 m. But as tsunami waves reach shallow coastal waters, they slow down and steepen rapidly, sometimes reaching heights of 10 m or more. Shallow bays and harbours tend to focus the waves and cause them to bounce around and amplify (or resonate), which is why the Japanese called them "harbour waves". Tsunami waves that overtop natural coastal beach ridges and barriers can surge considerable distances inland in low-lying areas (order of 100's of metres to a kilometre or more depending on the wave run-up height and the "roughness" of the land cover and built environment).

Key definitions to quantify tsunami are:

- Tsunami period (minutes)—the time between successive wave peaks. This can fluctuate during any particular event and vary between different locations within the same region. Periods are usually in the range of a few minutes (e.g., "local source/local impact" tsunami) to an hour or more for a "distant source/national impact" tsunami.

- Tsunami height (m)—taken as the vertical crest-to-trough height of waves, but it is far from constant, and increases substantially as the wave approaches the shoreline. Usually only used in conjunction with measurements from a sea-level gauge to express the maximum tsunami height near shore.

- Tsunami run-up (m)—a more useful measure of the tsunami hazard is the maximum run-up height, expressed as the vertical height the seawater reaches above the instantaneous sea level at the time (including the tide). This measure still has the drawback that it depends markedly on the type of wave (rapidly rising and falling, a bore, or a breaking wave) and on the local slopes of the beach and foreshore areas, so it is highly site-specific.

- Inland penetration (m)—the maximum horizontal distance inland from the shoreline or mean-high-water mark inundated by the tsunami. It depends on

the tsunami run-up and local topography, barriers and slopes within the coastal margin.

2.2. Tsunami wave behaviour

The behaviour of any given tsunami wave-field that arrives at any particular coastal locality can vary substantially, depending on several factors, including the generating mechanism, the location, size, and orientation of the initial source (disruption), source-to-locality distance, local seabed and coastal margin topography and the "birds-eye" plan shape of the coastline. Conversely, all tsunami from the same source area with similar generating mechanisms will propagate to a coastal locality in a similar manner, in which case scenario modelling can be very useful in determining local vulnerability to tsunami hazards.

The size and distance of the source makes a substantial difference in the type of waves observed at any locality: large linear fault ruptures will cause more regular (periodic) waves over long stretches of coastline, while smaller "point" sources (underwater landslides, volcanoes etc) generate radially-dispersive, unstable shorter sequences of waves that will peak over a short distance of coast (de Lange and Healy, 1999). Some "point" sources may initially produce an even distribution of wave energy around the generation region. However most generating mechanisms will involve a highly directed distribution of wave energy towards a particular region or country (for a "regional" or "distant" tsunami) or a particular locality (for a "local" source).

The arrival of a tsunami wave-train (i.e., it isn't just one wave) is often manifest by an initial drawdown of the level of the sea (much faster than the tide), but for other events, the first sign may be an initial rise in sea level. The waves that propagate towards the coast seldom break before reaching the near shore area, and the larger waves will appear to have the whole ocean behind them. Thus the larger waves will move relentlessly forward inundating the coastal margin, until they reach maximum run-up height before receding temporarily. Other tsunamis occur as an advancing breaking wave front or bore, which is the type of wave most people associate with a tsunami. Most tsunamis reaching the New Zealand coast historically have behaved as a non-breaking wave, although have tended to form bores within shallow estuaries and river mouths–see Section 3 for historical observations (de Lange and Healy, 1999).

A tsunami wave-train that impinges on the coast in one area can often reflect back offshore, spreading out in a circular wave front. The nearshore part of that reflected wave front can become trapped at the coast and move downcoast parallel to the shore to other localities. These secondary waves are known as coastal-trapped waves. They move quite slowly in shallow water arriving at other localities many minutes or even hours after the initial tsunami wave impact on the coast. This has implications for emergency managers in determining when the tsunami is no longer a danger

2.3. Tsunami risk

Tsunami damage and casualties are usually a result of three main factors:

- Inundation and saltwater contamination by potentially large volumes of seawater could flood roads, buildings, and farmland (causing long-term saltwater damage to pasture or crops).

- Impact of swiftly-flowing water (up to 30–60 km/hr), or travelling bores on vessels in navigable waterways, canal estates and marinas, and on buildings, infrastructure and people where coastal margins are inundated. Swiftly-flowing water, or bores, can also cause substantial coastal erosion and scour of road carriageways, land and associated vegetation. The return or out-rush flows generated when a large tsunami wave temporarily recedes are often the main cause of drowning, as people are swept out into deeper water.

- Debris impacts—most casualties arise from the high impulsive impacts of floating debris picked up and carried by the up-rush (inundating) and down-rush (receding) flows.

Mitigation of the tsunami risk to a developed region such as the Bay of Plenty and eastern Coromandel may differ somewhat for "local source/local damage" events compared with "distant source/national damage" events. Tsunami hazard mitigation measures are primarily achieved through:

- land-use planning controls on coastal developments e.g., hazard risk zones;

- public education and awareness of tsunami hazards and consequences;

- community awareness of the appropriate personal response to Emergency Management warnings as well as "natural" indicators of "local" or "regional" source tsunamis (e.g., by association with strong earthquake shaking, offshore volcanic explosion, and/or unusual behaviour of the sea);

- appropriate Emergency Management Response plans for each of "local", "regional" and "distant" source tsunami events (e.g., signage, public preparedness for "local" and "regional" tsunamis (see previous point), inundation and evacuation maps, adequate knowledge of likely impact, and adequate and timely warnings with associated evacuation plans (for "distant-source" tsunami).

3. Historical tsunami events

3.1. Historical catalogue of events

Tsunamis are relatively common around the New Zealand coast, especially the eastern seaboard, with an average of 12 to 13 events >1 m high occurring every century somewhere around the country (de Lange, 2003).

In historical times (since 1840), tsunamis are known to have affected places along the Bay of Plenty and Eastern Coromandel coastline at least twelve times (Table 1).

Table 1. Summary of tsunamis, their effects and their sources that have been recorded in the Bay of Plenty and Eastern Coromandel area in the historical era (1840–1996). Abbreviations: HWM – high water mark; RRF – rapidly rising and falling water levels; BC – coastal bore; BS – coastal stream/river bore; BW – breaking wave.

TSUNAMI IMPACT DATA								TSUNAMI SOURCE DATA	TSUNAMI IMPACT: DESCRIPTIVE ACCOUNTS/COMMENTS
Year	Date (local) Duration (D)	Location	Max crest-trough ht. (m)	Run-up (max height relative to sea level at time) (m)	Max inundation height above HWM (m)	Max horiz. extent of inundation past HWM (m)	RRF, BC, BS, BW	Source Source location	
1840	28-Jul	Whitianga (Mercury Bay)						Storm, not tsunami	According to McKay (1949) a wave of character like 1947 March 25 Gisborne tsunami threw *HMS Buffalo* on shore at Whitianga (Mercury Bay), and wrecked her, while fish were thrown ashore between Hick's Bay and Te Araroa. Other sources, for example, Riddle (1996), indicate that the *Buffalo* foundered in a storm, not a tsunami. Credibility of reference to fish stranded at Hick's Bay (see entry below) is unknown.
1840	August?	Hicks Bay		≥ 1	≥ 1			Source unknown, possibly storm. Reliability low	Fish thrown ashore between Hicks Bay & Te Araroa, at the same time as the wrecking of the *Buffalo* at Whitianga 1840 (McKay 1949). However, wrecking of the *Buffalo* was due to storm, not tsunami.

1868	15-Aug D>24 hrs	Port Charles		About 2.5	About 2.5		BS, RRF	M 8.8+ earthquake, S Peru/ N Chile	Tide rose to "unprecedented extent" washing away timber and doing "other" damage. (Weekly Press 22 August 1868); Serious flood, owing to rise in tide of 6 ft [1.8 m] in 5 minutes. The wharf, 7,000 ft of timber carried away, plus other [unspecified] damage. (Daily Southern Cross 22 Aug 1868); At 0200 water rushed into houses on the flat near sawmill. Flat about 3 ft [0.9 m] vertically above spring tides, and mill buildings on flat built on blocks 2 ft [0.6 m] high. Water was over 2 ft [0.6 m] high in houses, and considerable difficulty experienced in removing people. Some swam to high ground. During the same day, water receded below LWM at ebb tide, and then returned suddenly as a big wave of several feet in height. The flow and recession occurred several times during low water. It drove logs up creeks, carrying away the whole of stacked timber and depositing it on the flat. Boats and a tramway were washed away. Some logs floated to Pakiri and also a boat. No damage to the mill machinery. One boat at mouth of creek held fast and was not damaged. (Daily Southern Cross 2 September 1868); Water said to have risen 6 ft [1.8 m] in five minutes. (Weekly News 22 August 1868)
1868	15 Aug	Huruhi Harbour, Mercury Island		2?	≥ 2	18 m		M 8.8+ earthquake, S Peru/ N Chile	During night of 14-15 August tide 60 ft further inshore than usual on flat in front of house, shells, seaweed and fish being scattered high along beaches round the harbour. Ten ton barge was set adrift from anchorage, later found 1/4 mile [400 m] up a salt water creek jammed between the banks. No noise heard during night. (Buchanan, Cameron (1977): *Ahuahu (Great Mercury Island)*: *Memoirs of Cameron Buchanan, resident of Mercury Island, 1859-1873*. Mercury Bay District Historical Society, Whitianga, N.Z.)
1868	15 Aug D 12 hrs	Maketu	≥ 1.5	≤ 0.9	?		BS, RRF	M 8.8+ earthquake, S Peru/ N Chile	On afternoon of 15 August, at time of high water, water in river receded to below LWM, then suddenly rose again. Fluctuations observed till evening. Rock turned over near mouth of river. (NZ Herald 28 August 1868)

1868	15-Aug D>24 hrs	Opotiki	3?	≥ 1.8	<1, possibly 0		BS, RRF	M 8.8+ earthquake, S Peru/ N Chile	First appeared at 08.30, at low water. Wave about 6 ft [1.8 m] high rushed up the river at rate of 6 to 7 knots, filling the river to HWM. Water remained high for several minutes, then retreated and was low tide again 1/4 hour later. Several smaller ebbs and flows continued until 13.00, only one being like the first. This was a wall of water 3 ft [0.9 m] high, again bringing the level in the river nearly up to HWM. At 14.00 the tide was at its proper HWM, remaining there for nearly 2 hours without any change. It then ebbed gradually. On 16 August, a sudden rise of 1ft 6 in [0.5 m] at 13.00. Estimated height of first wave, about 10 ft [3 m]. (Letter from W Mair, at Opotiki, to Dr Hector, dated 17 Aug 1868) At 8.30 am at about low water, a bore of no less than 6 ft [1.8 m] high rushed up the river with great velocity swamping boats. In a few minutes river filled to HWM, remained high for several minutes, then retreated and in 15 min was at low water again. Several small ebbs and flows followed. At 11.30 a bore of about 3 ft [0.9 m] rushed up river at rate of 6-7 knots, reached nearly to high water, and retreated again over 15 min. Several smaller bores followed and at about 14.00 the tide rose to its usual level and scarcely any perceptible change occurred until 16.15, when it commenced (contd) to ebb. On 16 August, sudden rise of about 1½ ft [0.5 m] during flood tide at about 13.00. (Letter from W Mair to Captain Hutton, dated 7 September 1868 [seems to be different from other transcribed letter], Te Papa Archives MS??)
1868	15 Aug	Opape	3?	≥ 1.8	<1, possibly 0		BS, RRF	M 8.8+ earthquake, S Peru/ N Chile	Similar effects as at Opotiki.
1868	15 Aug	Torere	3?	≥ 1.8	<1, possibly 0		BS, RRF	M 8.8+ earthquake, S Peru/ N Chile	Similar effects as at Opotiki.
1868	15 Aug	Raukokore, near Cape Runaway		3?				M 8.8+ earthquake, S Peru/ N Chile	First wave broke at 04.00-05.00. Receding water swept boats, canoes, timber out to sea. Schooner, 3-4 miles off shore "fetched" heavily and broke foremast. (Letter from W Mair, at Opotiki, to Dr Hector, dated 17 Aug 1868); First wave came in at about 04.00, and swept boats, canoes and ? out to sea. Wave thought to come from eastward. (Letter from W Mair to Captain Hutton, dated 7 September 1868 [seems to be different from other transcribed letter], Te Papa Archives MS??)

1868	15-Aug	Cape Runaway		3.0				M 8.8+ earthquake, S Peru/ N Chile	3 m wave between 4–5 am. (Fraser database 1998) GLD comment: Location probably Raukokore (based on letters in Te Papa, see Raukokore effects column.
1877	11-May	Port Charles	2.4-3.5	≥ 1.2 to 1.8	Est. 1.5		RRF	M 8.8+ earthquake, N Chile	Tide ebbed and flowed 20min all day. Ave height =2.5 m, Max=3-3.6 m (Fraser database 1998)) Tidal wave washed 100 logs (stored) up creek or its branch. A few logs over bank. Awakened about an hour before daylight by sound of water round house, saw logs washing upstream. Whole flat area under water. Tide rushing in & out all day about every 20 minutes. Effect at high water in evening less than in morning. Tide rose and fell an average of 8 ft every 20 min all day. Sometimes as much as 10-12 ft. At 1400 it possibly rose and fell more. At near LW it came up to level of wharf and went back nearly to reef. Punt broke away, but no damage. (Thames Advertiser May 17)
1877	11-May	Mercury Island		Est. ≥ 1.5?				M 8.8+ earthquake, N Chile	Two vessels dragged anchors and were driven ashore, but were floated at next tide with no damage. Tide rose to alarming height, but no damage done. Current estimated at 8-10 knots. Only damage was a fence washed away. (Thames Advertiser May 17)
1877	11-May	Tauranga		1.2-1.8	0.6-1.5? above HWS			M 8.8+ earthquake, N Chile	Tide fluctuated of 3 m several times during day, continued on the 12th. (Fraser database 1998) The tide rose suddenly several feet higher than spring tides, and kept rising and falling a foot at short intervals all day. (Thames Advertiser May 12) Tide rose from 2-5 ft higher than spring tides and receded rapidly. Several other rises during the day. A number of houses on the beach flooded. (Thames Advertiser May 14) No damage reported. At 0800, tide rose suddenly 2-3 ft above the usual spring tides, and receded rapidly. Sudden rises of a foot throughout day. (NZ Herald May 12)
1883	29-Aug	Thames		1.5				Pressure-wave tsunami, attributed to eruption of Krakatau	Tide became full during ebb flow. (Fraser database 1998)

1883	29-Aug	Coromandel		0.9				Pressure-wave tsunami, attributed to eruption of Krakatau	Wave was seen at low tide, then tidal fluctuations.(Fraser database 1998)
1883	29-Aug	Whitianga		1.8				Pressure-wave tsunami, attributed to eruption of Krakatau	The water rose 1.8 m during ebb flow. (Fraser database 1998)
1883	29-Aug	Tairua		1.8				Pressure-wave tsunami, attributed to eruption of Krakatau	A rise and fall of 1.8 m was observed. (Fraser database 1998)
1883	30-Aug	Maketu		0.9				Pressure-wave tsunami, attributed to eruption of Krakatau	A 0.9 m bore swept up the Kaituna River. (Fraser database 1998)
1922	12-Nov	Whitianga		0.9				M8.3-8.5 earthquake, Chile	Max rise to HWM. Rises & falls at intervals of 20 minutes throughout day, diminishing towards evening. Fluctuations on lesser scale next morning (13th). (GNS files, unpublished data)
1937	April	Opotiki		0.9				Unknown, possibly not tsunami.	A rise of 0.6 m above the road at the bridge over the Waiotahe River at about 9 pm, followed by a fall, and a further rise to 0.9 m above the road. Newspaper accounts record abnormal tidal levels were also reported at Thames and Ngatea, and a car was caught in a tidal creek at Matata following a sudden rise in the tide. Reliability of this event is low as the date is uncertain and there may be confusion with a flooding event due to high intensity rainfall during the same month. (de Lange and Healy 1986b)

1947	25-Mar	East Cape		0 or <0.3				M7.1 (slow) earthquake off Poverty Bay at 38..85°S 178.87°E	Nothing unusual noticed, but houses were located remote from sea. (GNS files, unpublished data)
1947	25-Mar	Bay of Plenty		0.0				M7.1 (slow) earthquake off Poverty Bay at 38..85°S 178.87°E	No effects recorded. (GNS files, unpublished data)
1948		New Zealand		0.0				M7.8-8 earthquake, Chile	No effect recorded on tidal records for Sept 9 at Lyttelton, Dunedin, Nelson, Greymouth, Auckland (Marine dept correspondence to Seismological Observatory Sep 28 1948.)
1950	13-Mar	Bay of Plenty		1.0				M5.3 earthquake at 38.4°S 178.0°E	Unusual disturbances along Bay of Plenty coasts. (Fraser database 1998). [Note: the earthquake is only of moderate magnitude and well inland. Likelihood of its producing a tsunami is low.]

1960	23-May	Whitianga		2.5				M9.5 earthquake, Chile	11 boats were also swept away, but they were recovered. (Fraser database 1998). Worst hit in BOP and eastern Coromandel with repeated surges of up to 7 ft (2.1 m) (am May 24) Some launches alternately stranded and riding high at moorings, others broke away but were recovered. (Bay of Plenty Times May 24 1960) On May 24, boats broken away from moorings. No major damage, although rise was 6-8 ft above normal and 2 to 3 houses on beachfront were flooded. (Bay of Plenty Times May 25 1960) 11 small craft swept out to sea, gardens and airport hangar flooded, fish left floundering on shore road. Sea surged up river, swept over road and into foreshore gardens. The aerodrome was flooded and 3 aircraft moved to safety when water reached over the wheels. 11 small craft swept up river or out to sea rescued. Estimated that river ran out at 25 knots, a tugboat at full steam ahead went backwards. (Auckland Star May 24 1960) Sea swept over road, flooding foreshore gardens. Aerodrome flooded, water over wheels of aircraft in hangar. Later aircraft located 11 small boats swept out to sea, or up the river. Water running out of river at estimated 25 knots early am May 24. (Evening Post May 24 1960) Surge reached several feet above high water at about 2100 (May 23). Boats broke adrift, were swept out or capsized. Resident described water a rapidly swirling river. Some boats recovered. Tide full at 17.55 (May 23) and the sound of surge against the tide could be heard [section of report missing, rest abandoned until rest of report obtained] (NZ Herald May 24 1960) Vessel hit bottom at Whitianga Wharf am May 24. Water surged up and down at 20-minute intervals. Buffalo wreck exposed. (NZ Herald May 25 1960). [Also see eyewitness account in Appendix 1]
1960	23-May	Great Mercury Is.						M9.5 earthquake, Chile	The water was reported to bubble and whirl. (Fraser database 1998)
1960	23-May	Mercury Cove		3.0				M9.5 earthquake, Chile	Sandbanks which were normally covered by deep water were exposed. (Fraser database 1998)
1960	23-May	Mercury Bay		2.3		-		M9.5 earthquake, Chile	Oscillations every 40 min for four days. (Fraser database 1998)
1960	23 May	Whangamata						M9.5 earthquake, Chile	Pipi bank exposed by surges [no detail on wave height]. (NZ Herald May 25 1960)

1960	23-May	Mount Maunganui	2.1	1.4				M9.5 earthquake, Chile	Max oscillation occurred between 4:30 pm and 5:00 pm. (Fraser database 1998). "Widely varied fluctuations recorded" during night and morning, more obvious at Mount than 3 miles into Tauranga Harbour. Watchhouse keepers at wharf first noticed fluctuations at 10pm, May 23. Began moderately, followed by a surging 2 ft (60 cm) rise at midnight, followed by drop of 2 ft by 1.15, in next hour a series of falls "for 2 ft", then minor fluctuations, followed at 3.45 by a rise of 3 ft. From 3.15 am-4.30 water level dropped 3.5 ft and between 4.30 and 5.00 fluctuations ranged up to 4.5 ft. After that tide rose and fell 2 ft every hour. According to Capt. Carter, normal range in 6-hr period was 5 ft, but from 00.00-7.00am range extended over 7 ft. (Bay of Plenty Times May 24 1960).
1960	24-May	Tauranga		~1				M9.5 earthquake, Chile	Max fluctuation occurred on the 25th. (Fraser database 1998). Many rises and falls causing no problems. Tide gauge at northern end of railway wharf recorded sudden rises and falls from 3.3 ft- 1ft during night. Gauge still unsettled at noon, May 24. (Bay of Plenty Times May 24 1960) Tides still irregular, but fluctuations not as large as May 24. At 11.30 pm (May 24), fall of 2.5 ft in 30 min. (Bay of Plenty Times May 25 1960) As Auckland Star reports. (Evening Post May 24 1960) Considerable scouring. At Sulphur Point (Bay of Plenty Times May 24 1960)
1960	24-May	Kaituna River		1?		-		M9.5 earthquake, Chile	(Fraser database 1998). Damage more serious than at Maketu. A portion of rock causeway at Ford's Cut washed away; minor damage to temporary work at the pumping station at the end of Ford's Rd. A recording taken at river mouth at 02.30 am May 24 showed a rise of 2 ft above normal high tide level. Between 4 am and 7 am four peaks recorded 1.5 ft above normal high tide mark. (Bay of Plenty Times May 25 1960)
1960	24-May	Maketu		1?		-		M9.5 earthquake, Chile	0.8 ha of land was lost during 24-27 of May. (Fraser database 1998). Erosion at Maketu Domain said to be speeded up, with 4-5 ft carved off the sea front, according to Tauranga drainage engineers. (Bay of Plenty Times May 25 1960) Little said to remain of recreation area. (NZ Herald May 25 1960).
1960	24-May	Whakatane		<1				M9.5 earthquake, Chile	(Fraser database 1998). No excessive movement noticed; small craft not affected. (Bay of Plenty Times May 24 1960) Tides erratic on May 24. Water dropped 1 ft in 5 min, then rose 1.5 ft in next 5 min. (Bay of Plenty Times May 25 1960).

1960	24-May	Opotiki		1.5				M9.5 earthquake, Chile	(Fraser database 1998)
1964	29-Mar	Tauranga		0.15				M9 earthquake, Alaska	(Fraser database 1998)
1976	14-Jan	Tauranga		0.10				M7.8 earthquake, Kermadec area	(Fraser database 1998)
1977	22-Jun	Tauranga		0.15				M7.2 earthquake, Kermadec area	(Fraser database 1998)
1994	6-Oct	Tauranga		0.1				M8.3 earthquake, Kuril Islands	Oscillations actually less than 0.1 m but unknown so 0.1 m used. (Fraser database 1998)

Information in Table 1 is based on data in de Lange and Healy (1986a), Fraser (1998), and data in the GNS Tsunami Database. The latter revises and updates Fraser's (1998) database (which in turn is an update of de Lange and Healy (1986a)) with new data, which comprise information from GNS historical archives and new accounts found as the result of a recent comprehensive investigation of historical newspaper accounts. It should be noted that the database has not yet been brought up to date with small events that have occurred since 1994 (e.g., the 1998 Papua New Guinea and 2001 Peru tsunamis were detected in New Zealand; both <0.25 m peak-to-trough height at the Moturiki Island sea-level gauge).

Some tsunamis listed in Table 1 were small and only detected by sea-level gauges. Such small to very small events were usually not noticed prior to the installation of sea-level gauges, and hence it is probable that the Bay of Plenty has experienced many more of these small (<0.5 m height) tsunamis, particularly from distant sources, than the historical tsunami database indicates. It should be recognised also that the record of significant tsunamis is probably incomplete in the 19th century either because the effects at isolated communities were not reported or because there were no communities to report them. Nevertheless, it is considered unlikely that catastrophic or severely damaging events in the Bay of Plenty area would have escaped notice had they occurred during the era of European settlement (post-1840).

The most substantial tsunamis to have affected the Bay of Plenty and eastern Coromandel areas in the last 160 years were generated by "distant" (remote) sources. The largest, in 1868, 1877 and 1960, were generated by very large earthquakes in the subduction zone along the Chile and southern Peru coastlines of South America—directly opposite and facing New Zealand's eastern seaboard.

Since European settlement around 1840, no "local source/local damage" or "regional source/regional damage" events are known to have occurred in the Study region. However, this is not unexpected as fault ruptures tend to have return periods of 100's to 1000's of years and volcanic eruptions, return periods of 1000's to 10,000 years or more.

3.2. Commentary on effects of substantial historical tsunamis

3.2.1. 1868 Peru tsunami

The 1868 tsunami was generated by a magnitude M~9 earthquake off southern Peru/northern Chile, in almost the same seabed location as the recent Peru M8.3 earthquake and tsunami in June 2001. The 2001 tsunami was small in New Zealand (unnoticed by the public), the largest effect being recorded on sea-level gauges at the Chatham Islands and Lyttelton Harbour, with tsunami wave peak-to-trough heights

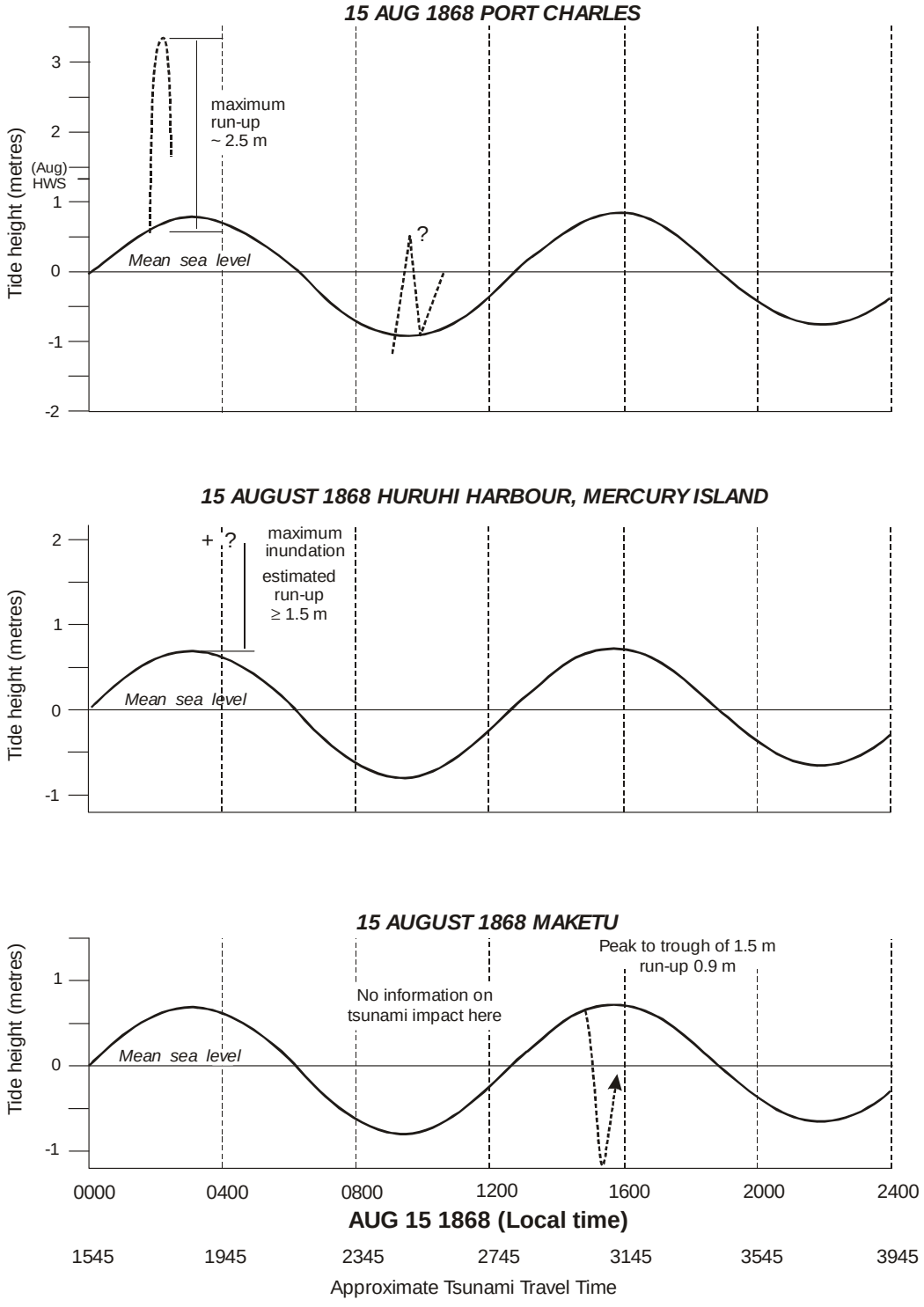
being about 0.6 m at both locations.[3] The greatest run-up height attained by the 2001 tsunami near its Peru source was about 7 to 8 m. In contrast, the greatest near-source run-up recorded for the 1868 tsunami was 18 m (HTDB/PAC, 2001).

In the Bay of Plenty and eastern Coromandel, the 1868 tsunami is known to have (see Table 1 for further details):

- inundated houses to a depth of 0.6 m, and washed away a tramway, boats and stacked timber at Port Charles (est. run-up height about 2.5 m);
- swept about 20 m inland over a flat area at Huruhi Harbour, Mercury Island;
- caused bores at Maketu (est. run-up height ≤0.9 m);
- caused bores at Opotiki (up to 3 m bore, est. run-up height 1.8 m) swamping boats, with similar effects observed at Opepe and Torere;
- damaged boats, and swept canoes, boats and timber out to sea at Raukokore (near Cape Runaway) with wave heights estimated to be 3 m above normal tidal levels (i.e., run-up height). These were previously thought to have occurred at Cape Runaway (de Lange and Healy, 1986a).

Figure 1 shows graphically the known, somewhat sparse, information on the elevations attained by the waves in relation to the tidal levels at the time. These have been estimated using the NIWA Tide Forecaster www.niwa.co.nz/services/tides. The greatest inundation effects at Port Charles, Mercury Island and Raukokore occurred near high tide in the morning of 15 August 1868 (NZST) some 17–18 hours travel time from the source, some 1 to 1.5 hours after the expected arrival of the first waves. This suggests the very first waves were considerably smaller. At Opotiki and Maketu the effects were only observed many hours after the damaging waves had been observed at the other locations mentioned (Figure 1).

[3] Note: the June 2001 Peru tsunami was measured at NIWA's Moturiki Island gauge at Mt. Maunganui peaking at 0.25 m peak-to-trough height some 24 hours after the earthquake off Peru. The first wave-front was detected 16 hours after the Peru earthquake.


15 AUG 1868 PORT CHARLES
Tide height (metres)
3
2
(Aug) HWS
1
0
-1
-2
maximum run-up ~ 2.5 m
?
Mean sea level
15 AUGUST 1868 HURUHI HARBOUR, MERCURY ISLAND
Tide height (metres)
2
1
0
-1
+ ?
maximum inundation
estimated run-up ≥ 1.5 m
Mean sea level
15 AUGUST 1868 MAKETU
Tide height (metres)
1
0
-1
Peak to trough of 1.5 m run-up 0.9 m
No information on tsunami impact here
Mean sea level
0000
0400
0800
1200
1600
2000
2400
AUG 15 1868 (Local time)
1545
1945
2345
2745
3145
3545
3945
Approximate Tsunami Travel Time

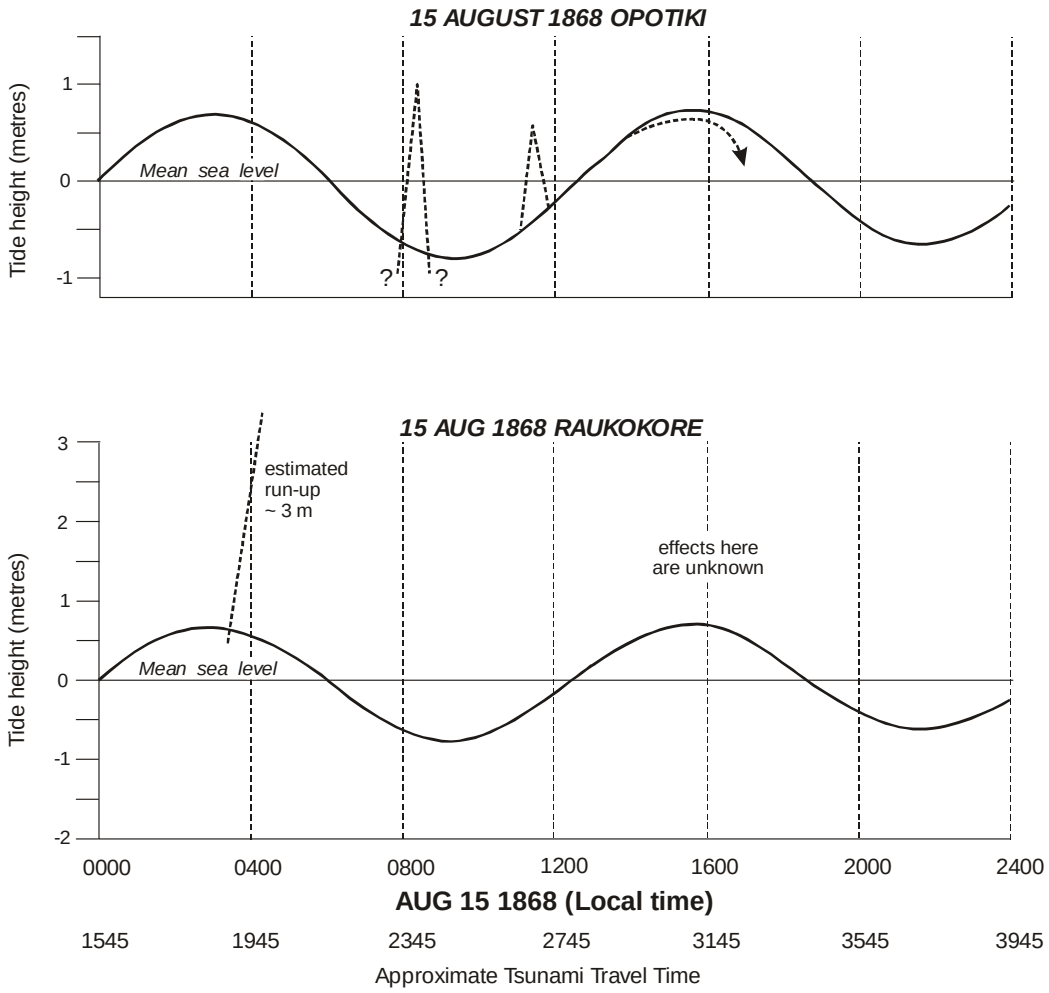


Figure 1: Known information on tsunami elevations in relation to predicted tides in Bay of Plenty and eastern Coromandel resulting from the 1868 Peru tsunami.

The paucity of information at each of the locations described above means there is uncertainty as to whether the tsunami waves that caused the greatest inundation were the largest, as there may have been larger waves at lower tides that did not reach far beyond the high-tide mark. At all locations, water levels oscillated for 24 to 48 hours after the first arrival of the tsunami, some waves probably still large enough at some places to endanger people and boats (i.e., about 1 m wave height, or 2 m from crest to trough) for at least 12 hours.

3.2.2. 1877 Chile tsunami

The 1877 tsunami was caused by a magnitude M~9 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, nor as well recorded in historical documents, as the 1868 tsunami. Many of the places

strongly affected in 1868 were again affected in 1877, but there were some notable differences.

In the Bay of Plenty, the 1877 tsunami is known to have (see Table 1 for further details and de Lange and Healy (1986b)):

- inundated the whole of the flat area at Port Charles, reaching houses and washing away 200–300 logs from the sawmill with water levels varying by up to 3–3.5 m every 20 minutes for 8 to 12 hours; also the wharf was damaged (estimated tsunami run-up about 1.2–1.8 m, possibly higher);

- caused several vessels to drag their anchors and a fence to be washed away at Mercury Island (run-up unknown, but probably no less than 1.5 m);

- flooded houses on a beach at Tauranga (no detail on specific location) with waves 0.6–1.5 m above high water springs mark (estimated tsunami run-up height about 1.2–1.8 m).

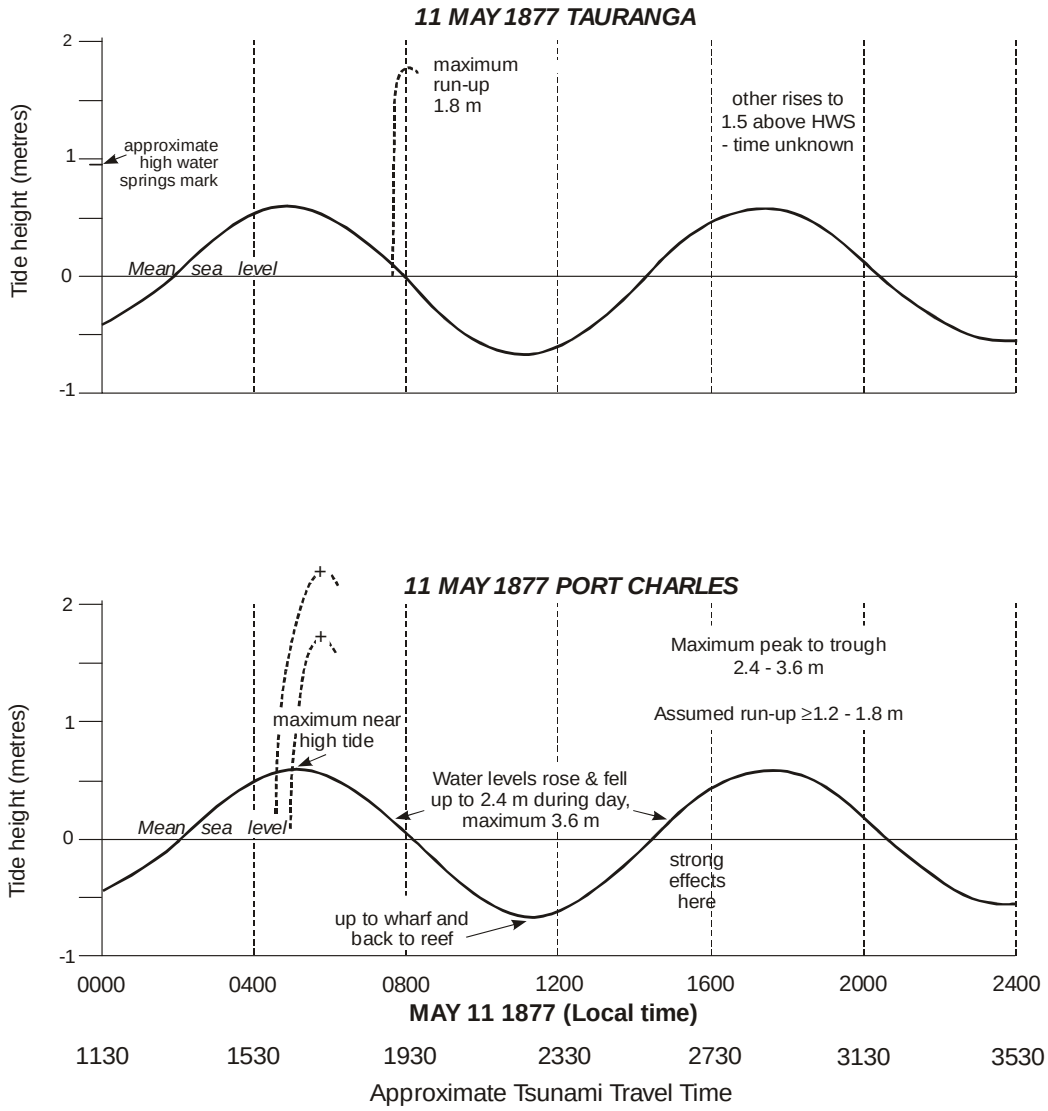


Figure 2: Known information on tsunami elevations relative to predicted tides in Bay of Plenty and eastern Coromandel resulting from the 1877 Chile tsunami.

Figure 2 shows graphically the known information on the elevations attained by the waves in relation to the tides (estimated using tide forecaster, www.niwa.co.nz/services/tides). As in 1868, the greatest effects at Port Charles occurred near high tide, some 15–17 hours after the initiating earthquake, and an hour or so after the expected first wave arrivals. The timing of the maximum inundation at (or possibly near) Tauranga, when waves were said to reach 1.5 m above High Water Springs mark, is not known.

3.2.3. 1960 Chile tsunami

The 1960 tsunami was generated by a massive magnitude M9.0–9.5 earthquake off central Chile, the largest earthquake in the 20th century. Along the Chilean coast, the

resulting tsunami reached a run-up of up to 25 m. The elevations reached in New Zealand were considerably less than those experienced in Japan and Hawaii (maximum tsunami run-ups over 6 m), where there was loss of life. Unfortunately, descriptive accounts in newspapers of the effects in New Zealand are not very extensive. There are, however, sea-level gauge records for several New Zealand ports (Heath 1976), including the port gauge at Tauranga Harbour (Figure 3). Note that the "smooth" line in Figure 3 is the predicted tide level at the entrance to Tauranga Harbour, while the wavy line represents the actual sea-level signal from the tide gauge at the port (Railway Wharf) during the tsunami event.

The first waves of the tsunami arrived at New Zealand tide gauges within half an hour of each other about 12.5 hours after the generating earthquake. The records show that the arrival of the tsunami was emergent at all stations taking several wave cycles over several hours for larger waves to develop. Also evident are the large differences in response to the tsunami from one tide gauge location to another. These differences are the combined effect of the tsunami travel path, the characteristics of the tsunami waves (principally, the periods represented in the wave train and the amplitude), and the resonant response properties of the harbours. Essentially when a harbour or coastal bay has resonant periods similar to those that are present in the incident waves, large amplification of the incident waves will occur at places within the harbour or bay.

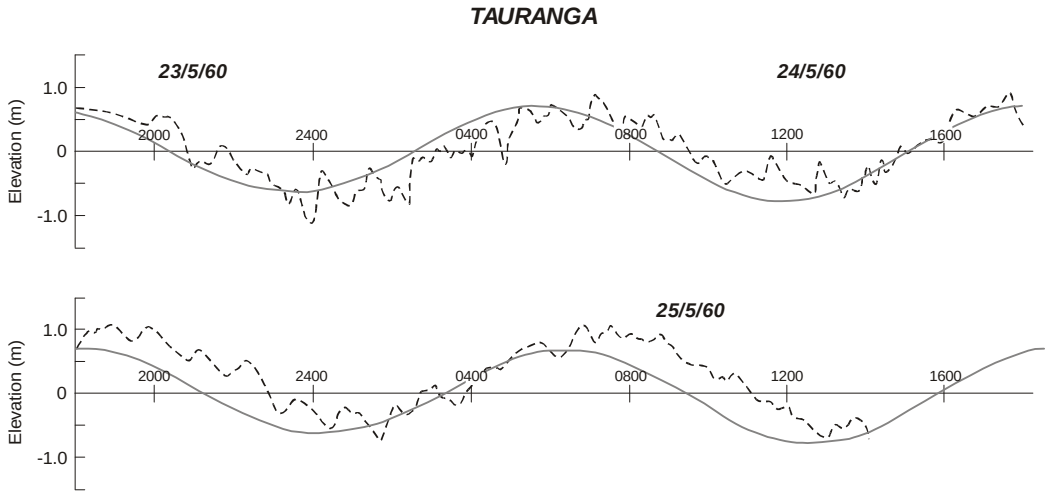


Figure 3: Sea-level gauge record (wavy line) from the Railway Wharf in Tauranga Harbour following the passage of the Chile tsunami on 24–25 May 1960 [Extracted from Heath (1976)]. The predicted tide at Tauranga Harbour entrance, estimated using the NIWA Tide Forecaster, is shown for comparison.

The effects of the May 1960 Chile tsunami in the Bay of Plenty and eastern Coromandel were (see Table 1 for further details):

- At Whitianga, tsunami wave heights of 1.8–2.5 m were observed. At the lowest drawdown, the wreck of the *Buffalo* was exposed. Many boats were

swept from their moorings, some being recovered. A large rock in the sea changed position. At its peak, the tsunami waves inundated the waterfront road, the airport and flooded several houses. [Extracted from de Lange and Healy (1986a) with estimated tsunami run-up of 2.5 m (Fraser database 1998)]. Appendix 1 contains a copy of recollections by Howard Pascoe and 28 other signatories of the events following the Chilean tsunami.

- At Mercury Cove, sandbanks normally covered by deep water were exposed, with estimated tsunami run-up of 3 m (Fraser database 1998).

- At Mercury Bay, oscillations every 40 min for four days (Fraser database 1998).

- At Mount Maunganui, the tsunami was first noted at 10 pm. A 0.6 m surge occurred at midnight, followed by a 0.6 m drop at 01.15 am. Minor fluctuations occurred until 02.45 am, when the water rose 0.9 m. Between 03.15 am and 04.30 am the water dropped 1.1 m. From 04.30 to 05.00 am, the range was 1.4 m, the maximum reported. After 05.00 am the water rose and fell 0.6 m every hour. The total water level range reported was about 2.2 m instead of the normal tidal range of 1.5 m. [extracted from de Lange and Healy (1986a)]. Estimated run-up: 1.4 m (Fraser database 1998).

- At Tauranga, a succession of rises and falls were recorded. The tide gauge at the Railway Wharf showed fluctuations of up to 1 m (see Figure 3). Considerable scouring was observed at Sulphur Point. Fluctuations with an amplitude of up to 0.8 m continued more than 14 hours after the first arrival. [Extracted from de Lange and Healy (1986a)]. Estimated run-up: 1.0 m.

- At Kaituna River, the rock causeway at Fords Cut was severely eroded, and the pumping station at the end of Ford's road damaged by salt water. [Extracted from de Lange and Healy (1986a)]. Estimated run-up: 2.3 m (Fraser database1998).

- At Maketu, the water levels rose and fell many times, in one instance, three times within an hour. Several boats were swept from the estuary and carried far inland. The water reached 0.6 m above high tide level (no time given). During the three-day period that the tsunami affected the area, 3 m of the reserve was removed, leaving a building perched on the edge of the sea, and an estimated 0.8 ha of land lost. However, the area was undergoing erosion prior to the tsunami. [Extracted from de Lange and Healy (1986a)]. Estimated run-up: 2.3 m (Fraser database 1998).

- At Whakatane, fluctuations of 0.5 m were recorded at Whakatane Heads on the first day of the tsunami. [Extracted from de Lange and Healy (1986a)]. Estimated run-up: 0.5 m (Fraser database 1998).

- At Ohiwa Harbour, residents reported extensive changes to Ohiwa Spit and bars within the harbour entrance some time after the tsunami. Gibb (1977) believed the tsunami to have accelerated erosion, but de Lange and Healy (1986a) considered it unlikely, considering the history of erosion prior to the 1960 tsunami.

- At Opotiki, estimated run-up: 1.5 m (Fraser database 1998).

The effects of all other known historical remote-source tsunamis from around the Pacific have been considerably less than 1 m in run-up height throughout the Bay of Plenty and eastern Coromandel region.

The Bay of Plenty has apparently also been affected by the atmospheric pressure wave generated by the explosive eruption of Krakatau, Indonesia, in 1883. This type of tsunami is a wave generated in the sea in response to a pressure wave in the atmosphere, which reportedly can pass many times around the earth. It can imitate a tsunami at large distances not directly affected by the tsunami that may have been generated at the source. For instance, the 1883 Krakatau pressure-wave (rissaga) tsunamis were recorded on tide gauges in Hawaii, Alaska, and Europe, while the tsunami travelling through the ocean was mainly confined to the Indian Ocean with little of it reaching further afield (Choi et al. 2003).

In the Bay of Plenty, de Lange and Healy (1986a) attribute the following effects to the 1883 Krakatau pressure-wave tsunamis:

- At Whitianga, starting 36 hours after the largest explosion, water suddenly rose 1.8 m at low tide then receded, leaving vessels high and dry before the water rose again.

- At Tairua, a 1.8 m rise and fall.

- Over 50 hours after the explosion, a 0.9 m bore, which travelled 4.8 km up the Kaituna River, was reported at Maketu.

Minor effects throughout New Zealand seem to have occurred over a five-day period.

3.3. Summary of historical record

In summary, three main events occurred over historical times (1840 onwards) with tsunami run-up heights of up to 3 m. All three events were generated by “distant” South American earthquake sources, occurring in 1868, 1877 and 1960. An unknown

number of smaller historical events have occurred, particularly events with run-up heights <0.5 m. Before instrumental records, these events probably went largely unnoticed. Since European settlement, no “local” or “regional” source events are known to have caused substantial tsunamis in the Study region.

4. Paleo-tsunamis

Given that the post-European historic record is relatively short (160 years), geological field investigations and geo-chemical analysis of sediment cores opens up the possibility of detecting, interpreting and dating large paleo-tsunami events to extend the tsunami hazard record for a region. Year 1 of the Tsunami Study focused on selecting and obtaining cores from potentially undisturbed sites in the Bay of Plenty and eastern Coromandel (Goff, 2003).

Locations for the paleo-tsunami field sites discussed below are shown in Figure 4. The relevant figure numbers for each site are given at the beginning of each sub-section. Sites are discussed in order from northwest to southeast.

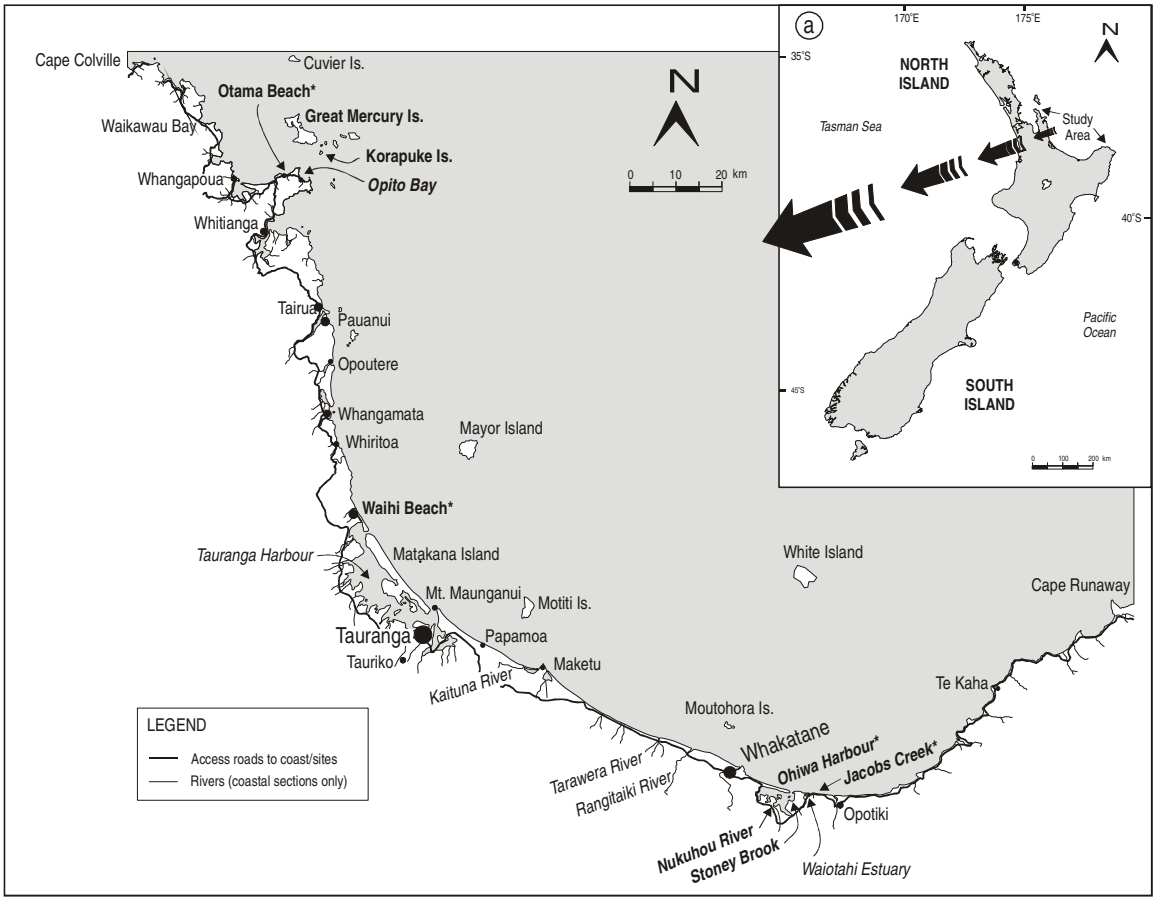


Figure 4: Study area in a regional context—showing access roads used and coastal portion of Bay of Plenty rivers. Main study sites are marked with an asterisk.

4.1. Otama Beach (Figures 5-6)

This is a north-facing beach on the mainland (in the lee of Great Mercury Island). It is about 2 km long and backed by sand dunes that rise up to about 8 m above mean sea level, extending inland for about 200 m. The surrounding hills are drained by the

Otama River that exits to the sea at the eastern end of the beach. The river is partially constrained by the dunes and forms an extensive wetland behind them to the south. This wetland has been heavily modified by farming, roading and drainage. Core OT1 was taken from the wetland about 250 m inland behind the dune system.

Geomorphological evidence indicates that there have probably been at least three key phases of significant geological activity: the first forming the heavy mineral rich layers that lie beneath the dunes; the second depositing a discontinuous pebble veneer on the second dune phase; and the third creating two marked phases of dune construction subsequent to pebble emplacement.

The core adds little to this interpretation, although paleo-environmental changes within the wetland appear to be generally related to terrestrial events. The one exception is the burial of an organic-rich layer at 2.23 m. This layer is overlain by a coarse sand that contains a moderate percentage of marine diatoms and fresh granules of Loisels Pumice. This is interpreted as a marine incursion, possibly associated with emplacement of the pebbles or heavy mineral-rich layer. The presence of Loisels Pumice is used to infer an age for this event between AD1302–AD1435 (McFadgen, pers. comm., 2003). Both the Loisels Pumice arrival and Healy Caldera collapse occurred within this timeframe, although, where present, what are believed to be Healy-related deposits underlie Loisels Pumice (McFadgen, pers. comm., 2003; Leahy, 1974). The assumption that Loisels Pumice took some time to arrive on the shore presumes that all the pumice was produced in one event, however pumice could equally have been produced in an earlier (smaller?) event or events and then washed up as the result of a later eruption from the same source. A definitive source for these tsunamis is not available, but the association with an eruption at that time and the occurrence of Loisels Pumice are used to infer an association with the Healy caldera collapse. However, Wright et al (2003) note that the chemistry of only some of the Loisels Pumice types are similar with Healy caldera, so clearly there are other Kermadec Ridge sources active in about the same timeframe, and likewise are potential tsunami sources (Section 5.2).

Geomorphologically, the discontinuous pebble veneer is situated on the seaward side of the second dune phase at Otama Beach (see photos in Goff, 2003). Probably correlative deposits of similar age are found in places throughout the northeast North Island (e.g., Nichol et al. 2003a, 2003b).

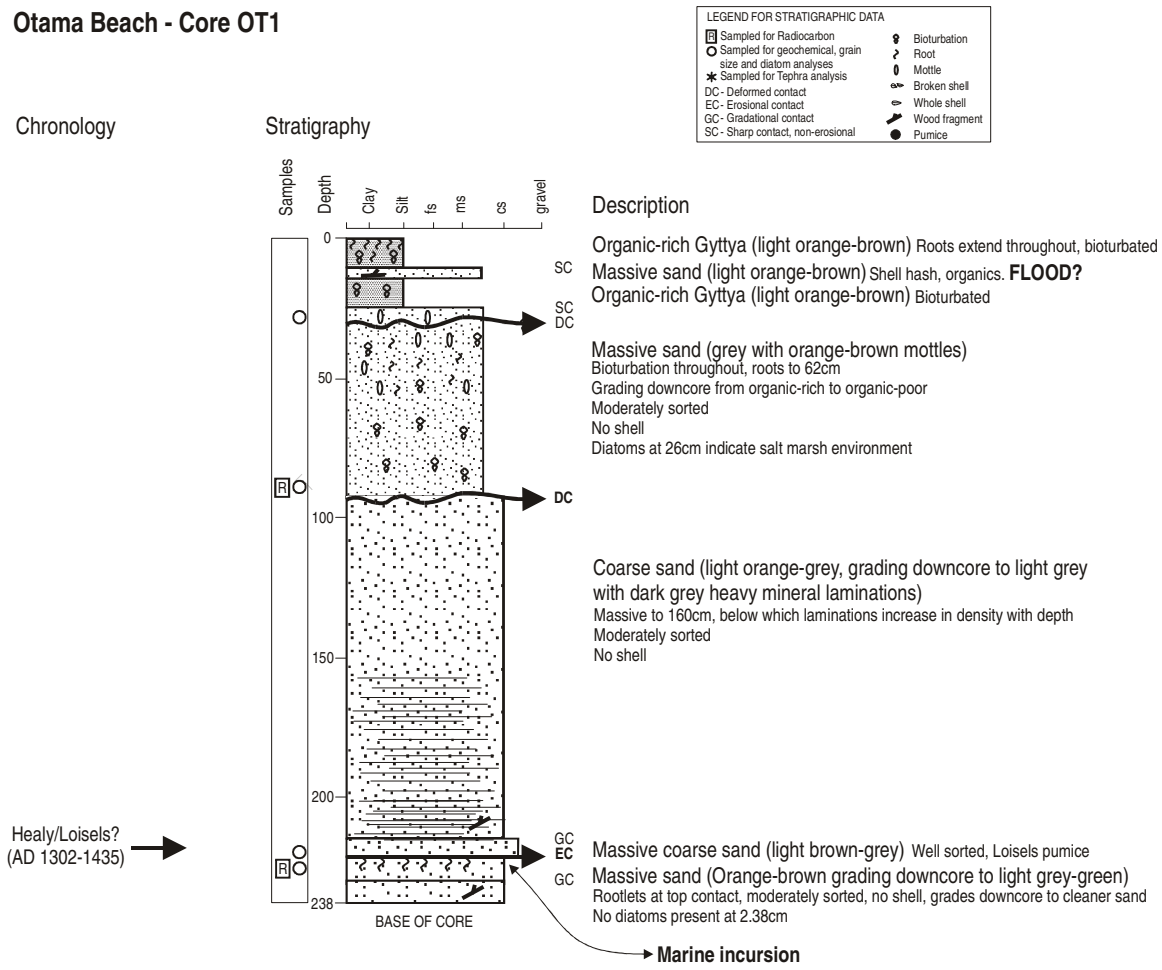


Figure 5: Otama Beach—Stratigraphy and chronology of Core OT1 (legend applies to all subsequent stratigraphic diagrams).

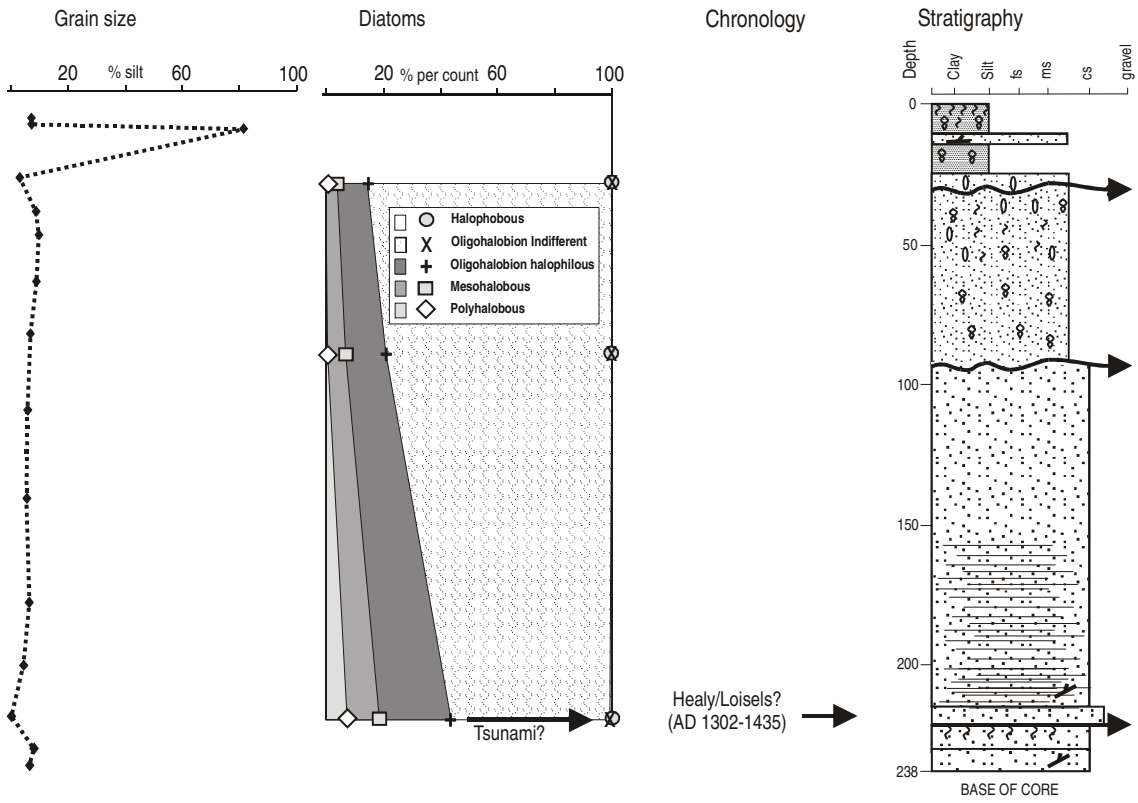


Figure 6: Otama Beach—Core OT1: Stratigraphy, sediment grain size, diatoms, and chronology (refer to text for interpretation of possible tsunamis).

4.2. Waihi Beach (Figures 7-11)

Core WAI 1 was taken from the seaward end of a wetland immediately to the east of Emerton Road and landward of the coastal dune system. This was the seaward end of a transect (WAI 2 was taken from the landward end), with the core taken from a drained wetland 450 m inland opposite the Sea Air Motel. Core WAI 2 was extracted from the landward end of the transect in the Department of Conservation section of this wetland about 1200 m inland. These cores were opened in the laboratory and sampled for radiocarbon, micropalaeontological and sediment analyses. Like the Otama Beach core, these are dominated by sandy material, but are more heterogeneous, containing frequent shell and shell hash units (dominated by cockle—*Austrovenus stutchburyi*).

The stratigraphy of core WAI 1 indicates two possible tsunami inundations around AD1302–AD1435 and at an estimated pre-Taupo age of about 2500 years BP. Identification of these events is complicated by poor diatom preservation and the radiocarbon date of a possibly reworked estuarine shell. The latter seems likely to be

reworked because it is situated immediately beneath a unit rich in Taupo pumice, the only occurrence of Taupo pumice in this core. The timing of the younger event is, like Otama Beach, based upon the unique presence of Loisels Pumice, and is assigned to what is termed a Healy/Loisels age.[4] A unique coarse sand unit with common marine diatoms, and an underlying buried soil mark this event. Subsidence is possible but seems unlikely. The older, pre-Taupo event is marked by a chaotic unit of coarse sand, shell, and shell hash. Similar units have been reported from Ohiwa Harbour cores. Shells within the unit contain numerous sub-tidal species exotic to the estuarine conditions that occurred at the time. The timing of this event has been estimated to ~2500 years BP, however it may well coincide with a radiocarbon dated tsunami inundation reported from a small valley off the eastern side of the Wairoa River south of Tauranga. The Hopping Farm site is 7 km inland (see Goff (2003) report for site details) and the outer bark of a drowned tree produced an age of 2962±52 years BP (WK-11860).

Core WAI 2 has a stratigraphic record extending back over 6500 years, with up to four possible tsunami inundations identified. The youngest event has been correlated with that of WAI 1 (AD1302–AD1435) because of the presence of Loisels Pumice. The unit has markedly coarse sediments, common marine diatoms, sub-tidal shells, and a geochemical signature indicative of tsunami inundation (Chagué-Goff and Goff, 1999). A similar estimate of ~2500 years BP is given here for the pre-Taupo event that is marked by a unique suite of diagnostic criteria indicative of tsunami inundation. Similarly, it is quite possible that this correlates with the dated event from Hopping Farm. The main focus of this report is on events that have occurred within the last 4000 years, but brief mention is made of two earlier events about 6300 years BP, and another estimated at 7000–7500 years BP. The latter may merely reflect increased storminess within the Bay of Plenty. Geochemistry indicates that the period prior to 6300 years BP is marked by increased marine influence and probably reflects an absence of the spit seaward of the site.

[4] The assumption that Loisels Pumice took some time to arrive on the shore presumes that all the pumice was produced in one event, however pumice could equally have been produced in an earlier (smaller?) event or events and then washed up as the result of a later eruption from the same source – hence the Healy/Loisels age)

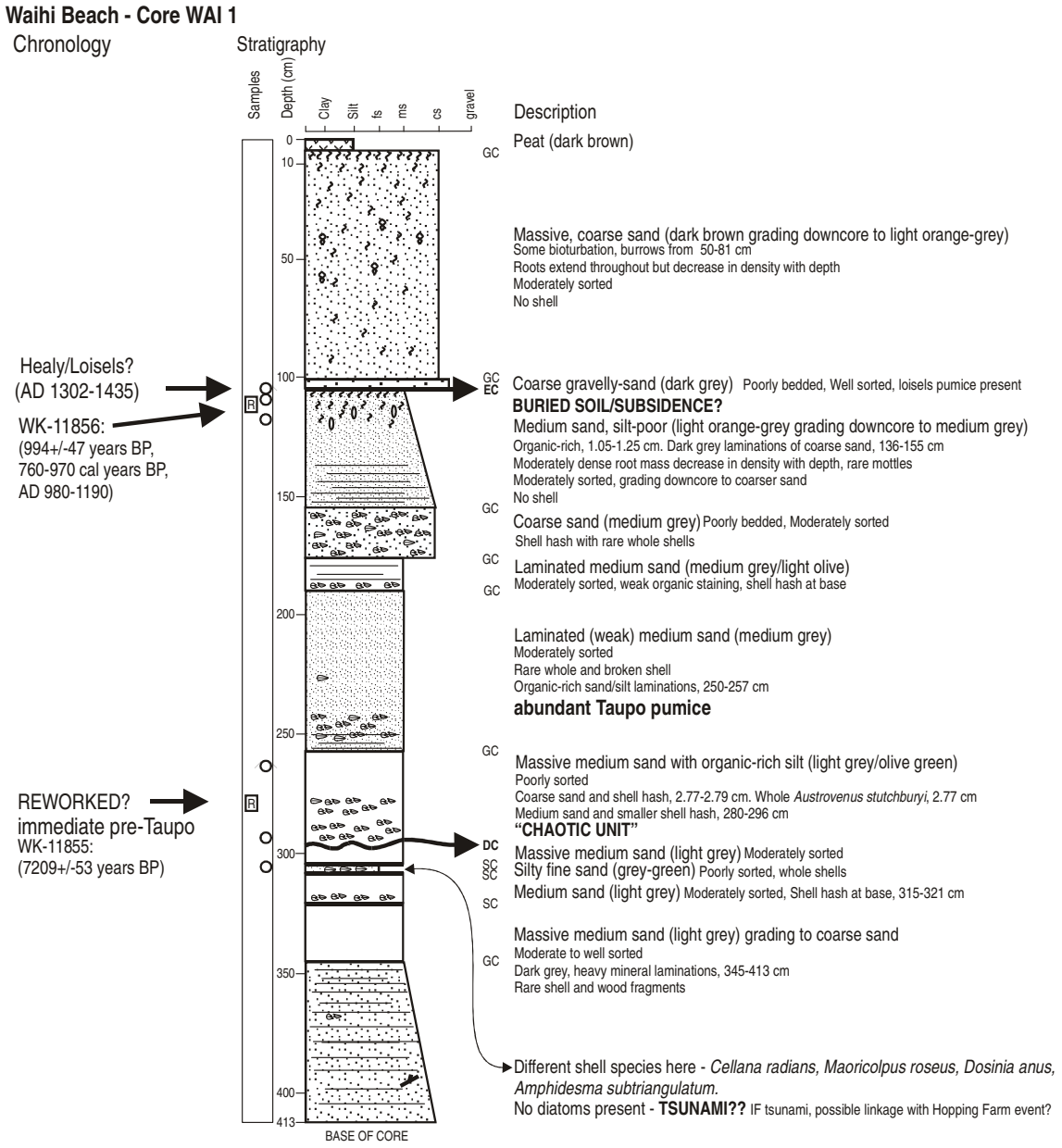


Figure 7: Waihi Beach—Stratigraphy and chronology of Core WAI 1

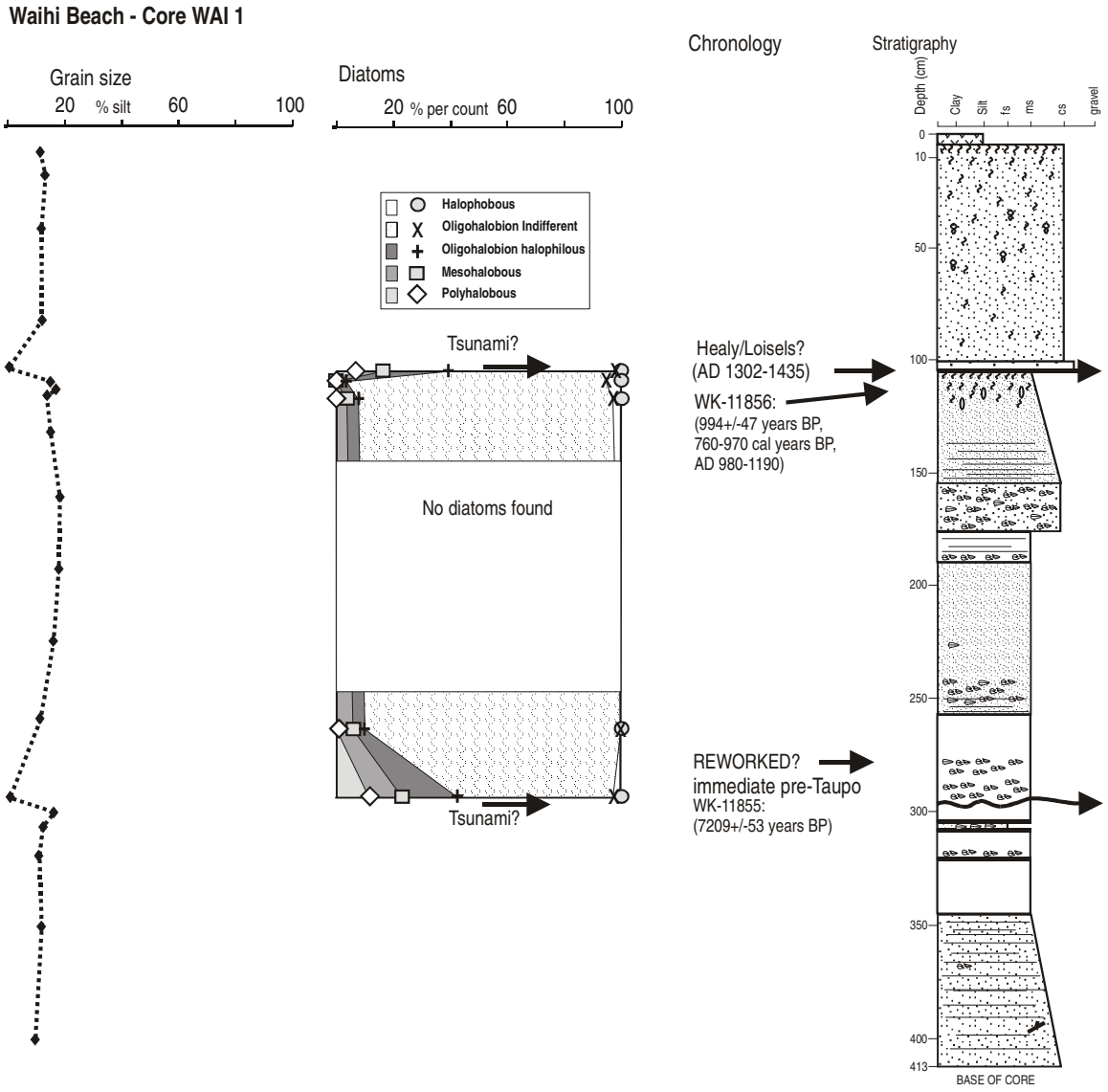


Figure 8: Waihi Beach—WAI 1: Stratigraphy, sediment grain size, diatoms, and chronology (refer to text for interpretation of possible tsunamis).

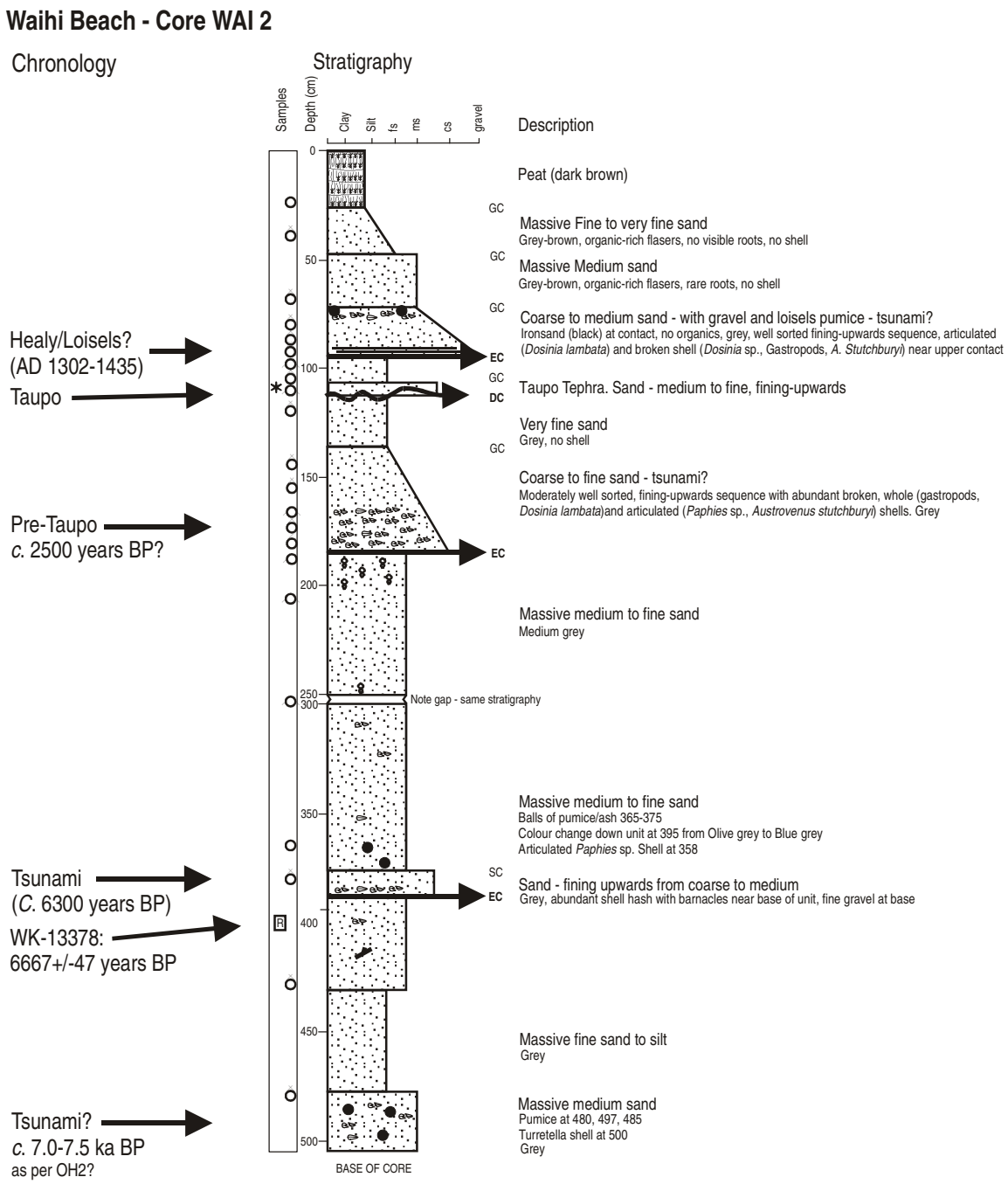


Figure 9: Waihi Beach—Stratigraphy and chronology of Core WAI 2.

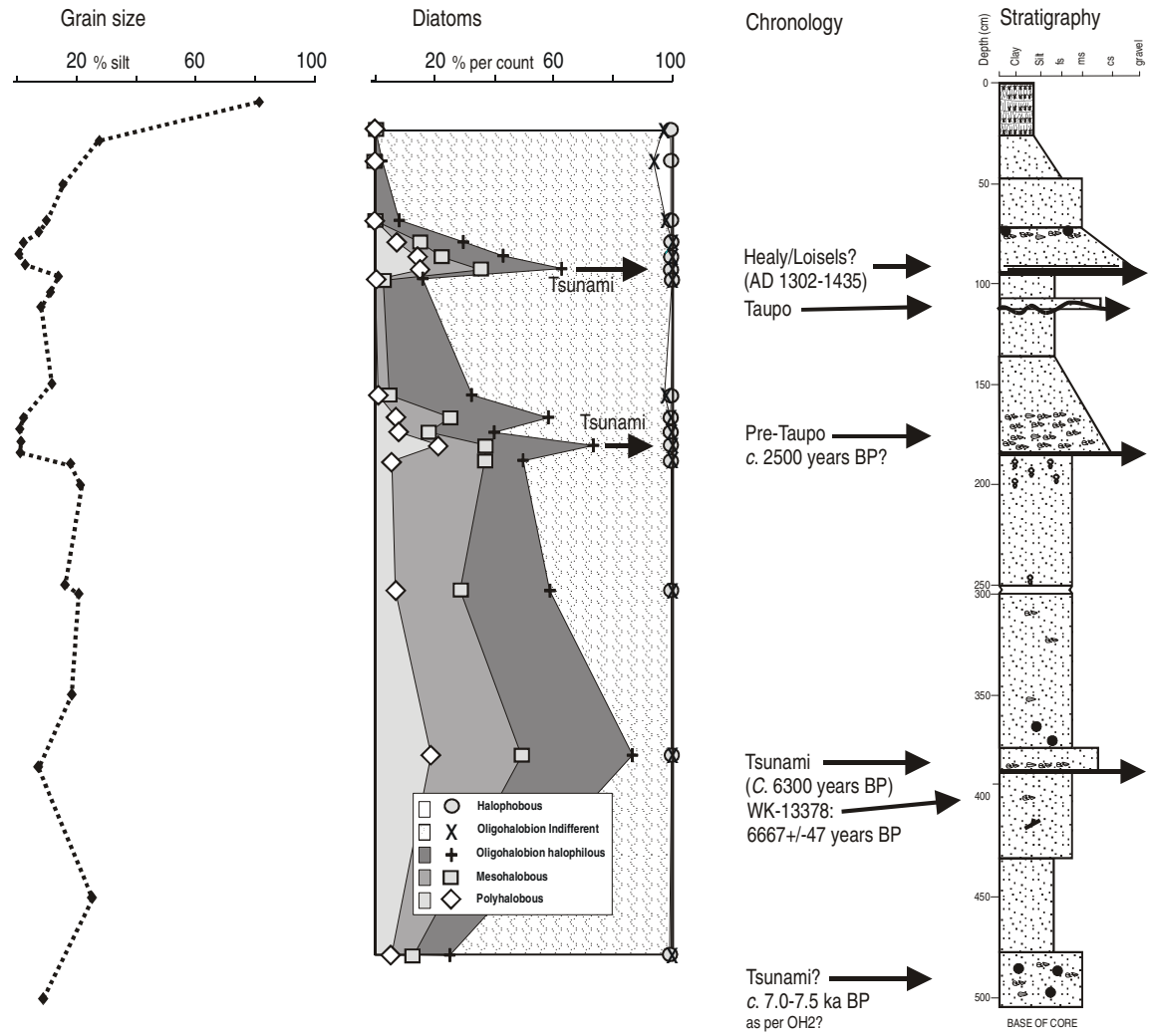


Figure 10: Waihi Beach—WAI 2: Stratigraphy, sediment grain size, diatoms, and chronology (refer to text for interpretation of possible tsunamis).

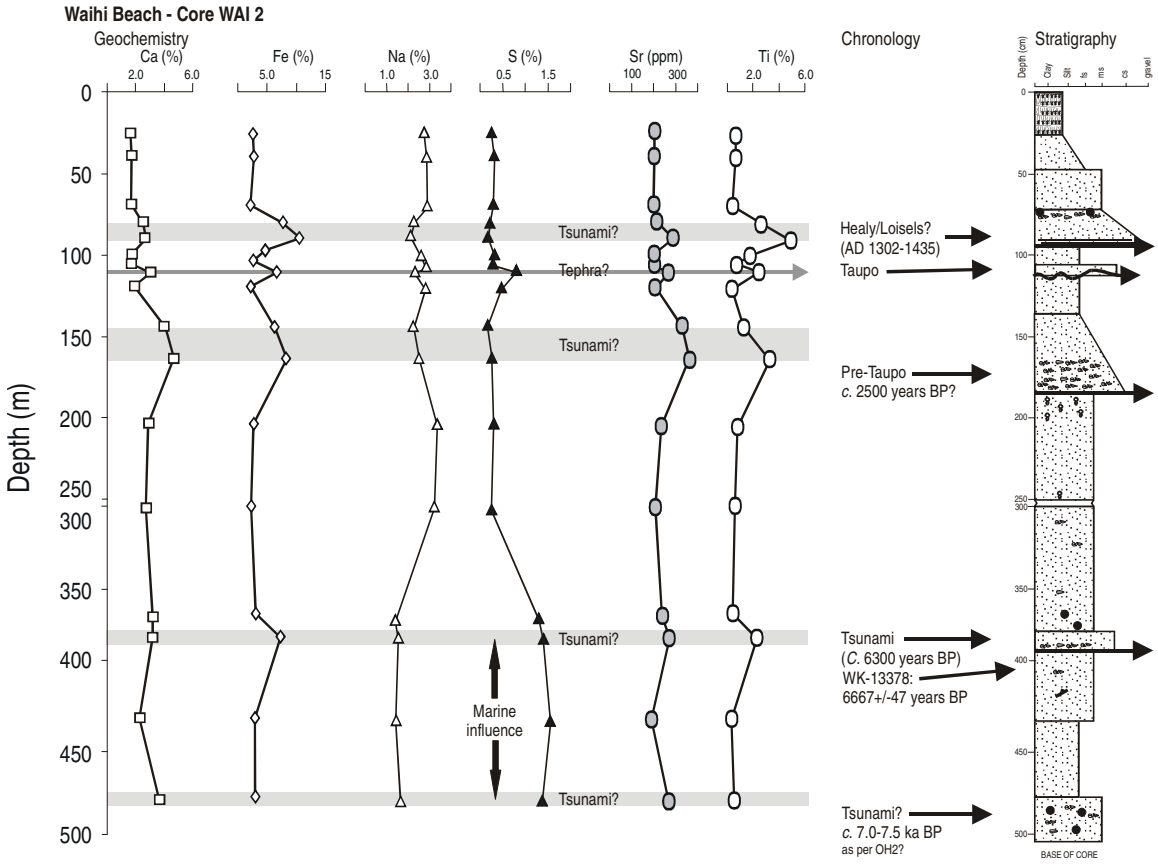


Figure 11: Waihi Beach—WAI 2: Stratigraphy, geochemistry and chronology (refer to text for interpretation of possible tsunamis).

4.3. Ohiwa Harbour (Figures 12-13)

Core OH2 was taken from a small tidal wetland estuary with a catchment less than 1 km^2. It was sampled for tephra, radiocarbon, grain size and micropalaeontology, and examined in detail.

Five tephras were identified, the Kaharoa, Taupo, Whakatane, Mamaku and Rotoma. Two "chaotic units" similar to the one reported in WAI 2 were identified. While good chronological control has been established, the record of past tsunamis is less well defined. What appears to be a fairly ubiquitous event around AD1302–AD1435 is evident in this core as well. It is marked by a combination of: a) fining-upwards coarse sand unit (not what is typically found for large storms); b) significant numbers of marine diatoms, and c) the presence of Loisels Pumice. This is almost immediately underlain by the Kaharoa Tephra, which is associated with a series of sandy interbeds that may be indicative of a locally-generated tsunami or increased storminess associated with the eruption. The most notable event is a 2.0 m subsidence that occurred between the Taupo and Whakatane eruptions (initial analyses inferred this to

have taken place between the Kaharoa and Taupo eruptions).[5] A deformed layer consisting of sand, silt, coarse organics, shell hash, and a significant number of marine diatoms overlies the subsided soil and is interpreted as a tsunami deposit. The clear subsidence signal is unique among all the cores studied, but whether tsunami inundation is localised or not is difficult to determine. It is tempting to link this with the pre-Taupo events that have been recorded at both Waihi sites and Hopping Farm. The chronological resolution we have from all these sites means that these events could be interpreted in several ways, as either one event, separate events, or some combination of them all. This will be discussed more in the summary below. Two further possible tsunami inundations are recorded prior to the pre-Taupo tsunami. These are significantly earlier than the 4000 year BP limit, the most recent probably coinciding with the chaotic unit of WAI 2 about 7000–7500 years BP, and an earlier one around 9500 years BP. Both require further study.

[5] This conforms with other evidence for subsidence in Ohiwa harbour within a similar timeframe (Hayward et al., 2004).

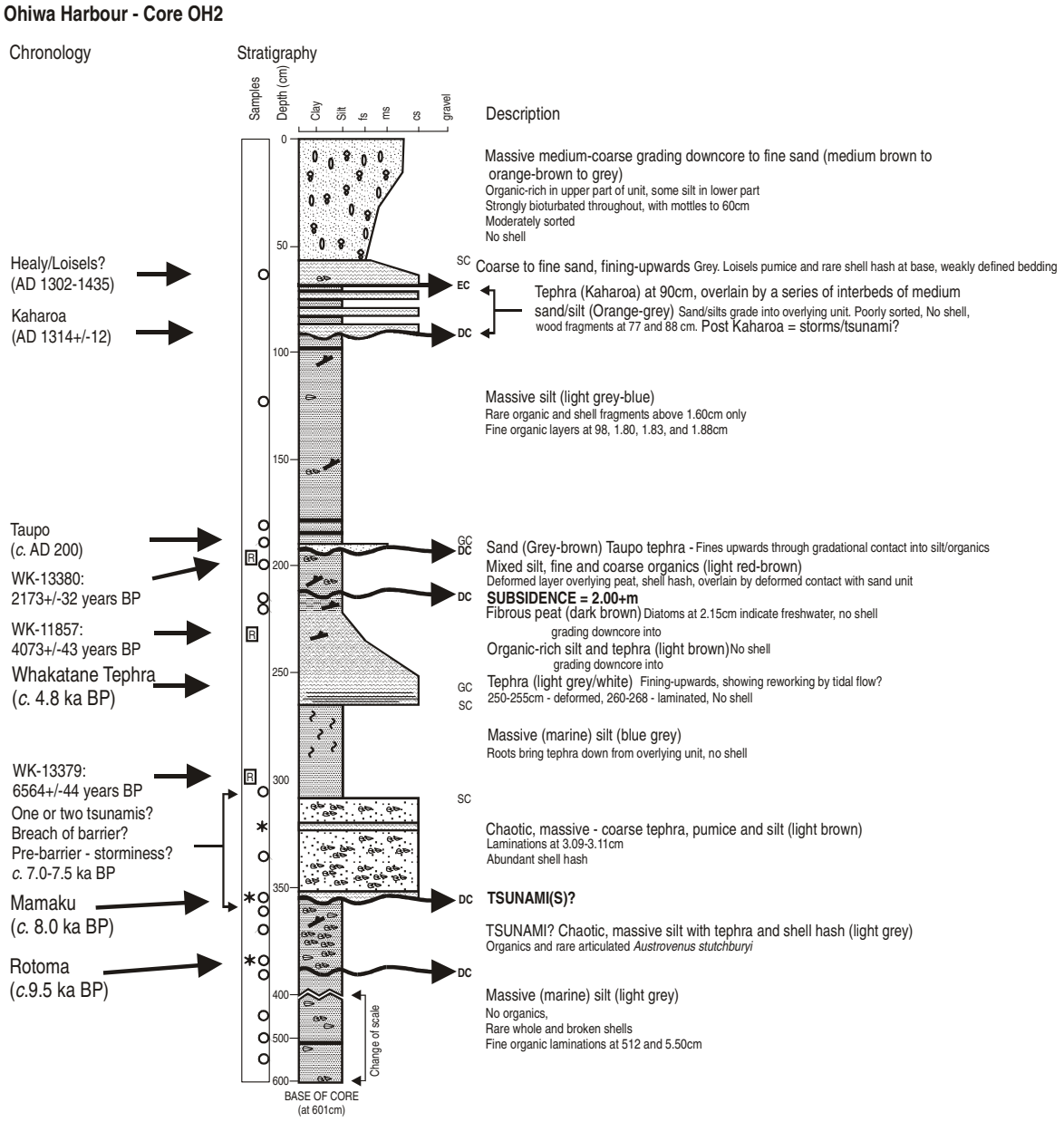


Figure 12: Ohiwa Harbour—Stratigraphy and chronology of Core OH2.

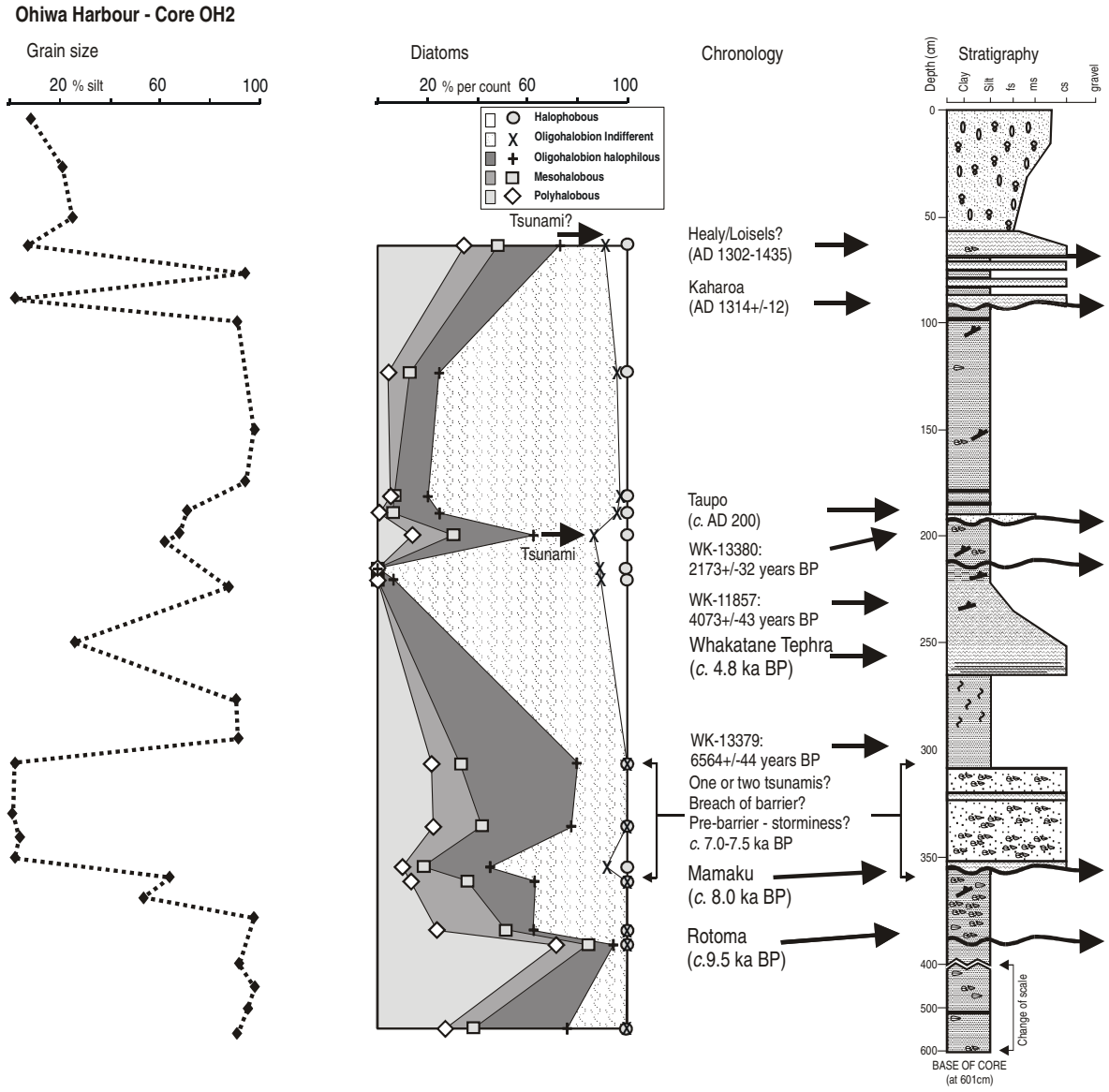


Figure 13: Ohiwa Harbour—OH2: Stratigraphy, sediment grain size, diatoms, and chronology (refer to text for interpretation of possible tsunamis).

4.4. Jacobs Creek (Figures 14-16)

Jacobs Creek is a small, narrow valley (about 4 km long and 0.5 km wide) about 1 km east of Waiotahi Estuary. The valley sides are composed of uplifted Quaternary marine sediments and, as opposed to the adjacent Waiotahi and Ohiwa basins, the valley appears to have been relatively tectonically stable during the Holocene. Core JC1 was taken from a drained wetland about 250 m inland from high water mark and behind a 3 m high coastal barrier. The core is primarily composed of peat and organic-rich silts (gyttya) interspersed with rare tephra units and associated sediments. Four

tephra units were identified; Kaharoa, Taupo, Mapara, and Whakatane. The environment appears to have been remarkably sheltered from the sea during the past 5000 years.

Two probable tsunami inundations are recorded, the most recent coinciding with similar events reported elsewhere around AD1302–AD1435, the earlier one is both pre-Taupo and pre-Mapara, placing it at an age of > ~2200 years BP. It is difficult to determine the precise timing of the pre-Mapara event, although given the number of times that only one pre-Taupo event has been reported in the cores discussed above, it seems highly likely that this is a similar inundation. Whether it links with that of the Hopping Farm tsunami is unclear.

Both events are marked by distinct peaks in marine diatoms, and also by unique geochemical signatures. Interestingly, geochemical data indicate a period of increased marine influence between these two tsunami inundations. It is possible that the barrier was reduced in height following the pre-Mapara event, although it may be indicative of poorly constrained evidence for additional tsunamis not recognisable in the sedimentary record.

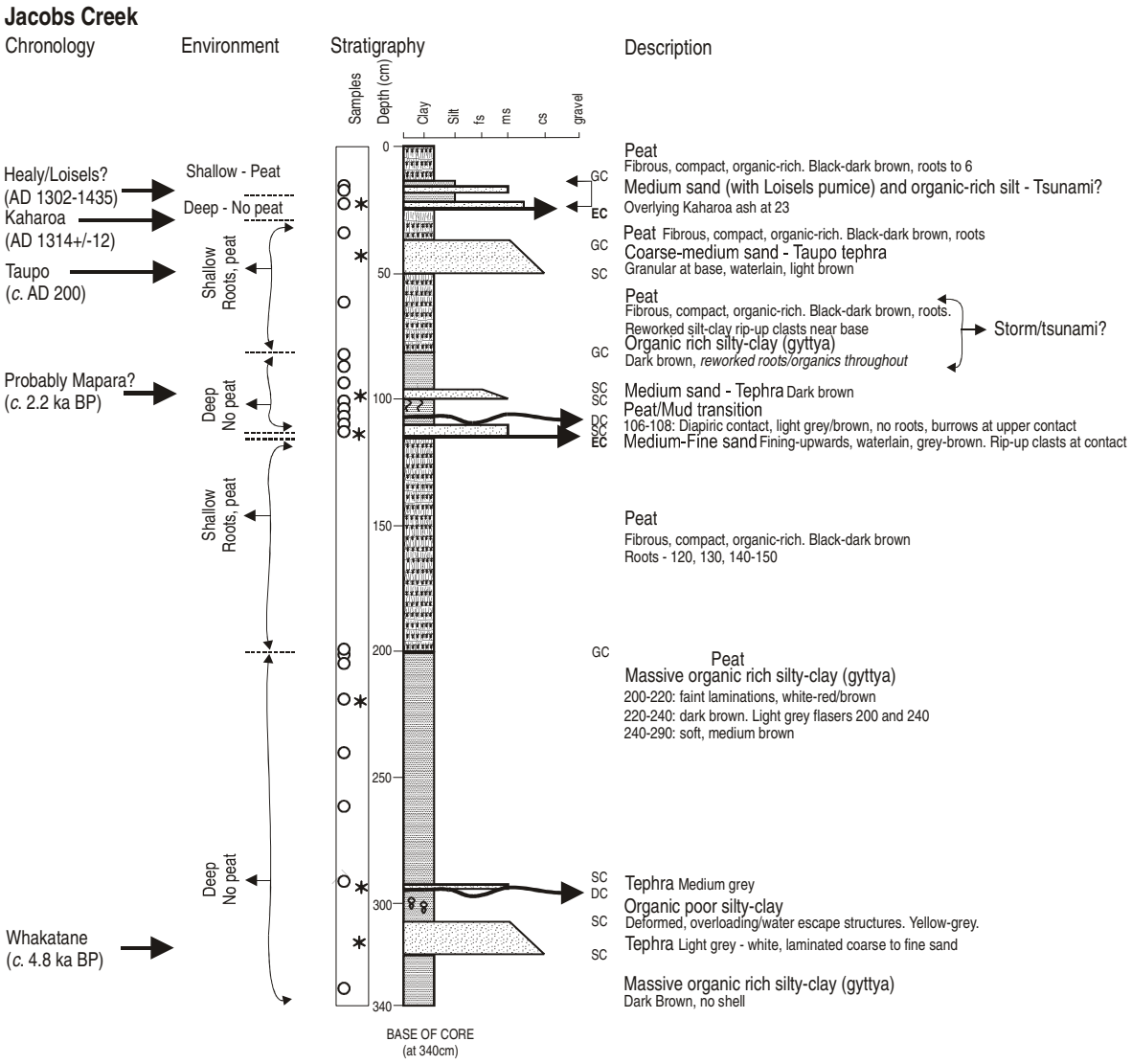


Figure 14: Jacobs Creek—Stratigraphy and chronology of Core JC1.

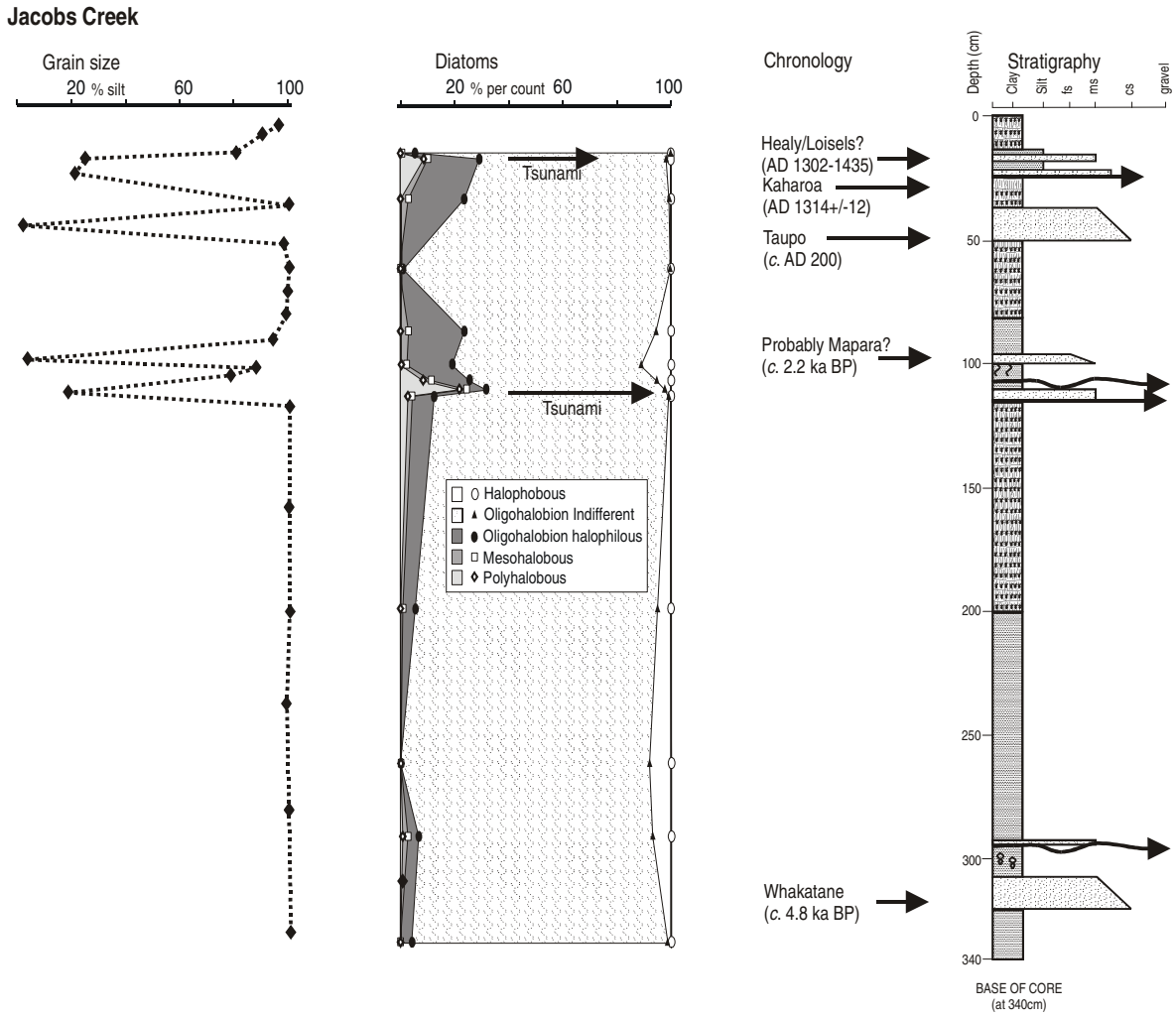


Figure 15: Jacobs Creek—JC1: Stratigraphy, sediment grain size, diatoms, and chronology (refer to text for interpretation of possible tsunamis).

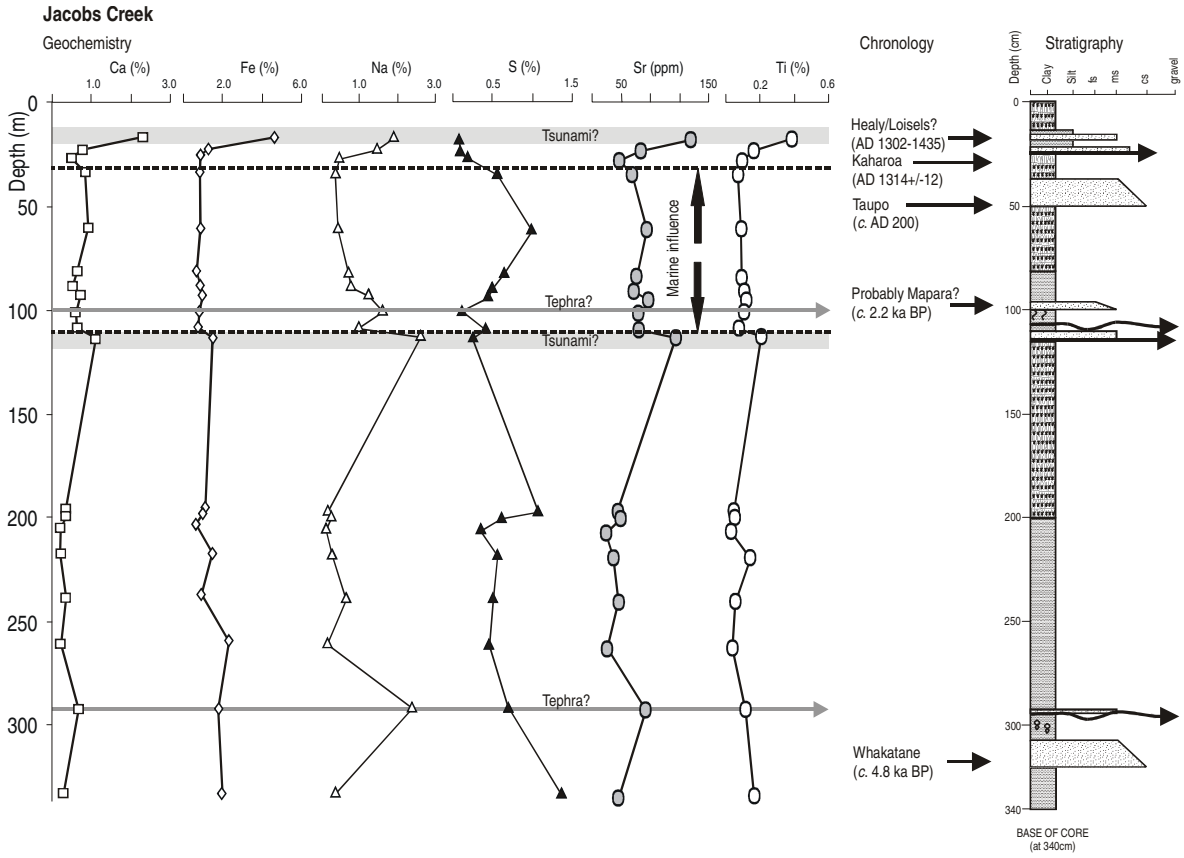


Figure 16: Jacobs Creek—JC1: Stratigraphy, geochemistry and chronology (refer to text for interpretation of possible tsunamis).

4.5. Summary (Figure 17)

The results of core data, ancillary tsunami research in the Coromandel (Goff, 2002a) and Bay of Plenty (Goff, 2002b), and an in-depth literature review have been used to produce the summary diagram shown in Figure 17. This work has served to supplement the record from cores.

These data are fragmentary at best, but we have attempted to ascertain the likely spatial extent of coastline affected by individual events using these results. The most important assumption made here is that it is assumed that if an event dates to within a specific 100 year time period it occurred at the same time as other events within the same period. It should be noted that the sedimentary record deteriorates over time and as such, what are inferred to be older, local source/local damage events may indeed be larger, regional source/regional damage events.

Core data have been used to identify up to six events dating back to 9500 years BP. However, only three have occurred within the last 4000 years. These are:

- AD1302–AD1435, an event dated by the occurrence of Loisels Pumice. This is believed to have been of region-wide significance and possibly happened around the time period of the Healy eruption and subsequent caldera collapse, because of the presence of Loisels Pumice.[6] This is later than a possible event recorded at Ohiwa immediately following the Kaharoa eruption, and the age range could be reduced to AD1314±12–AD1435;

- AD1314±12, a small possible tsunami (or increased storminess) recorded at Ohiwa Harbour only. This is believed to have been of local significance;

- A pre-Taupo (pre-Mapara at Jacobs Creek) event estimated to have taken place at about 2500 years BP, although it could be contemporaneous with the Hopping Farm tsunami dated to 2962±52 years BP (WK-11860). This is believed to have been regionally significant. It is tempting to group all these together and yet they may represent a series of separate local events that occurred within an approximate 500-year time frame. The only compelling reason for grouping these together is that the evidence from Hopping Farm, albeit close to sea level adjacent to a river valley, is of widespread forest destruction at least 7 km inland associated with a unit of marine sediments. We suggest that inundation to such a substantial landward extent represents a regionally-significant event and as such appears likely to be recorded at other sites.

Supplementary data (see top of Figure 17 for references), much from archaeological sources, add a further two events to this paleo record. These archaeological sources primarily include data about occupation sites overlain and/or separated by sand units associated, in many cases, with Loisels Pumice. The dating—based upon radiocarbon dates and cultural associations, elevations above sea level, and general characteristics of the deposits—have been compared with data collected during this present contract. Evidence for an event between AD1302–AD1435 is found throughout the Bay of Plenty. At Hahei, Hot Water Beach, Onemana, Whangamata, Whiritoa, Thornton, Pupuaruhi and Waiotahi there is evidence for either two events during this time period (a Loisels Pumice event overlain by another inundation) or of just the later event that can be stratigraphically separated from Loisels Pumice-related activity. There appears to be only one more recent event, and that is local subsidence recorded at Kohika.

The record is more confusing prior to Kaharoa, but there is a possible local subsidence event around 1200–1300 AD reported at Waiotahi, Maniatutu Road (near Paengaroa), and Parton Road near Papamoa East (Goff, 2002b). The 2002 report tentatively suggests that tsunami inundation was associated with this subsidence. However, poor

[6] Further ongoing research aims to provide more confirmatory evidence of the association between Healy or other Kermadec Ridge volcanology events and Loisels Pumice using geochemical and geophysical analysis (see Section 5.2).

dating control for the Maniatutu Road event (800–1800 years BP) means that it may equally have taken place around 1600–1700 years BP in conjunction with possible events recorded at Parton Road and Poplar Lane (off SH2 between Papamoa East and Te Puke). Prior to these events there is a period between 2000 and 3000 years BP that may represent possibly two inundation events (as shown in the summary). There are additional data points for this time period from the Papamoa and Parton Road.

The summary of the paleo-tsunami record from sediment cores for the past 4000 years (excluding the post-European historical record) is as follows working back in time, with BP short for Before Present:

- AD1600–AD1700: Kohika (*local-subsidence event*).
- AD1302–AD1435: a) later (**regional-impact event**).
  b) earlier (**regional-impact event**).
- AD1200–AD1300: Waiotahi, Maniatutu Road(?), Parton Road (*local-impact event*) [Goff, 2002b].
- 1600–1700 years BP: Papamoa East area—Maniatutu Road, Parton Road, Poplar Lane (*local-impact event*).
- 2000–3000 years BP: a) One event (**regional-impact event**) at 2500–2600 years BP recorded at Waihi Beach (both sites), Ohiwa + other supplementary sites incl. Waiotahi.
  b) One event (*local-impact event*) at 2900–3000 years BP recorded at Hopping Farm [Goff, 2002b].

In summary, over the past 4000 years a total of two major regional-impact paleo-tsunami events have been recorded in sediment cores—one in AD1302–AD1435 (with some evidence for two separate major events in this period) and an earlier event at 2500–2600 years BP. There is also evidence in various sediment cores that up to four local-impact paleo-tsunami events have impacted localised areas of the Bay of Plenty in AD1600–AD1700 (local subsidence a factor), AD1200–AD1300, 1600–1700 years BP, and 2900–3000 years BP.

A key point of these paleo-tsunami investigations is that the resolution used is only capable of identifying tsunami events with run-up height larger than 5 m (Goff, 2003). This lower cut-off limit arises from several factors including: a) a paucity of undisturbed sites due to on-going coastal development; b) the resolution of the methods used to detect and confirm paleo-tsunami deposits; c) sediment core locations are behind elevated sand-spit barriers and at least 250 m inland, together with the paleo-tsunami deposits appearing to be undisturbed by storm surges and overtopping impacts. Any further confirmation of the estimated tsunami run-up heights from paleo-tsunami events would need to be confirmed using a tsunami wave model (with realistic land topography) once a credible source-generation scenario can be constructed.

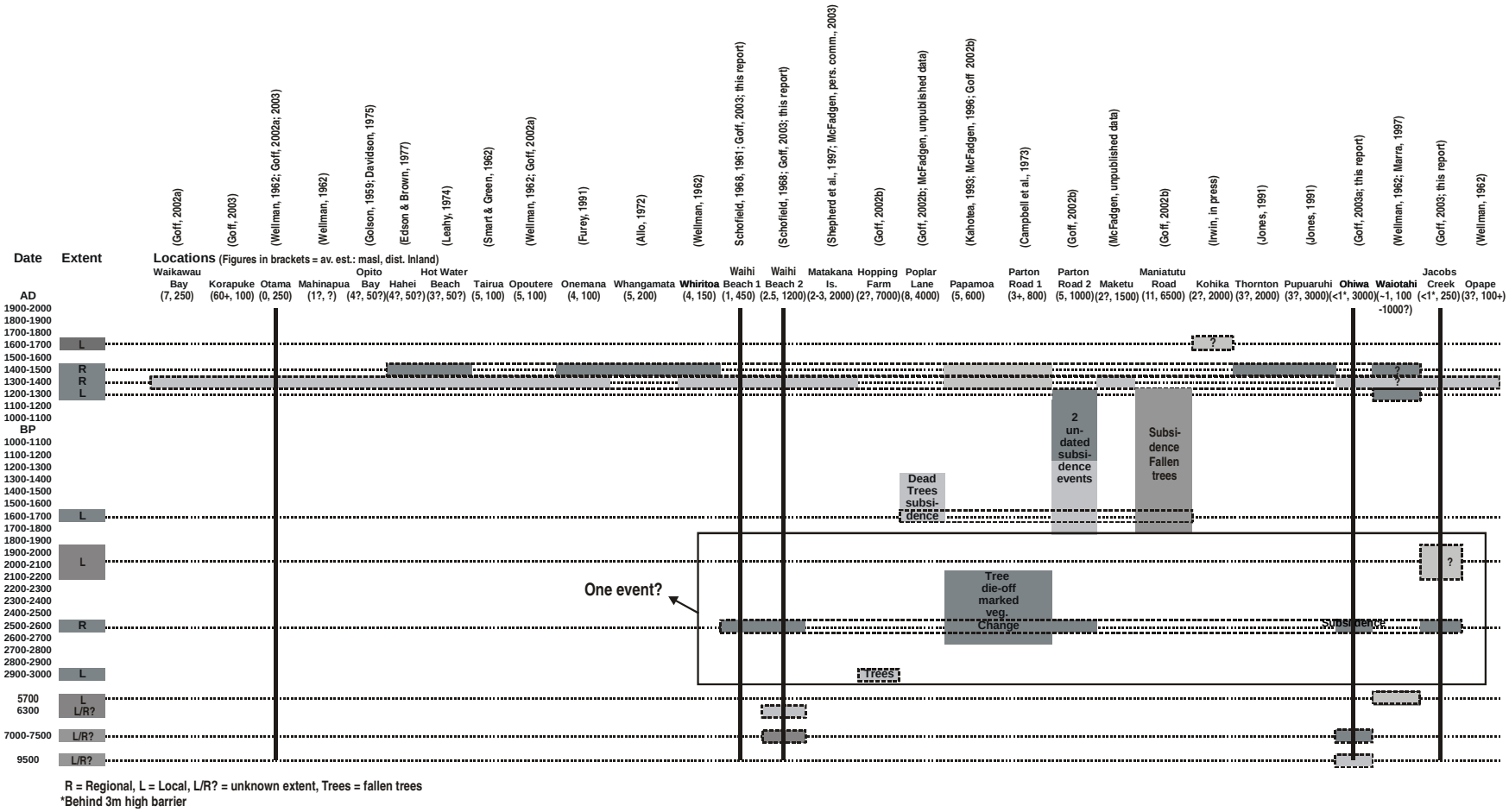


Figure 17: Summary of paleo-tsunami data discussed in Section 4 of this report (bold vertical lines indicate core sites). The paired values in brackets below the location names are average estimates (av. est.) of: (i) metres above sea level [masl] and (ii) distance inland in metres.

5. Tsunami sources (Bay of Plenty and eastern Coromandel)

How often, how big, and where from are important questions that need to be addressed in order to assess Bay of Plenty's vulnerability to tsunami hazards. Fraser (1998, published in de Lange and Fraser (1999)) has developed probabilities of occurrence of tsunamis of various wave heights at several locations, including Tauranga, using the NZ historical database and port tide gauge records (Table 2). However, a 160-year historical record is unable to reflect the full range of possible "local", "regional" and "distant" source tsunami events, the largest of which will have return periods of hundreds, possibly thousands of years. Therefore, Fraser's probabilities should be used tentatively, with the realisation that they may only represent the minimum hazard. For example, the historical record for the Bay of Plenty and eastern Coromandel contains no significant local or regional source earthquake and volcanic events, yet we know that these sources exist and paleo-tsunami results from Section 4 provide clear evidence for at least two large regionally-generated tsunami in the pre-historical era.

Table 2: Tsunami return periods (years) for the specified heights determined for a selection of New Zealand major and minor ports from historical records. The return periods were calculated using the annual exceedance probability distributions of Fraser (1998) (after de Lange and Fraser, 1999).

	Tsunami height (m)			
Location	1.0	2.5	5.0	10.0
Whangarei	179	930	14500	3510000
Auckland	85	427	6280	1360000
Tauranga	**80**	**322**	**3300**	**345000**
Gisborne	44	67	135	556
Napier	56	97	243	1540
Wanganui	79	147	414	3260
Wellington	40	119	728	27200
Lyttelton	35	52	101	376
Timaru	63	130	439	5010
Dunedin	125	1075	39000	51000000

Our understanding of the frequency/magnitude of tsunami events in the Bay of Plenty and eastern Coromandel can be extended beyond the historical record in several ways:

- Firstly, our understanding of vulnerability to tsunami can be extended by searching for, and identifying in geological records, the deposits of paleo-tsunami (pre-historical tsunami). Generally, only larger tsunamis (>5 m) leave visible deposits, with the identification of smaller events requiring higher resolution studies. Such deposits are not found at all coastal sites. However,

they are vital for not only extending the magnitude and frequency record back in time, but also for verification and augmentation of model data and for iteration with model data to ensure realistic inundation scenarios.

- Secondly, we can continue geophysical investigations of the seabed, tectonics and plate subduction processes. Geological surveys of the continental shelf off Bay of Plenty are continuing, using various techniques such as multibeam swath systems and seismic profiling through the seabed layers to determine the magnitude, approximate age, and frequency of occurrence of fault ruptures, submarine landslides, and catastrophic volcanic eruptions or caldera collapses that may potentially have caused tsunamis. Further work is also investigating the significant changes in crustal structure along the plate margin, as it is not yet certain if the entire subduction zone is a potential tsunamigenic hazard.

- Thirdly, we can identify all possible "local", "regional" and "distant" sources of tsunami, their frequency/magnitude relationships, and then model the propagation of tsunami from them. This will identify the whole range of events that can be expected, but doing such research is a long-term project that has only just begun. Geological and paleo-tsunami verification of such models is also needed.

5.1. Distant source–national impact tsunamis

Until more robust numerical and probabilistic models can be developed, we can gain some insight into the magnitude and frequency of distantly generated tsunami by looking at the longer historical record of earthquakes in remote areas that generate tsunami capable of reaching New Zealand. In particular, the history of large tsunamigenic (tsunami-generating) earthquakes along the South American coast spans hundreds of years longer than New Zealand's history, and large tsunamigenic earthquakes have been identified back to the 16th century (Table 3). Table 3 shows that there have been nine events in the last 450 years that produced near-source run-up heights near to or greater than those produced locally by the 1868 or 1877 events (highlighted in bold, Table 3), and hence probably capable of producing significant tsunamis in New Zealand. The average recurrence time (50 years) is about that experienced anywhere in New Zealand during the last 160 years. This provides a first estimate of the frequency of significant South American source tsunamis in New Zealand.

Table 3. Large South American earthquakes that have produced tsunami with maximum wave heights greater than 8 m locally (extracted from HTDB/PAC, 2001). Those events in bold are either known to have caused, or have the potential to have caused, significant impact in New Zealand comparable with the 1868, 1877 and 1960 tsunami.

Year	Month	Day	Lat. (°N)	Long. (°E)	M_b/M_s	M_w	M_t	Max. local height (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**	**Arica, 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, N. Chile/ S. Peru**	**10 (Chat) 4 (Main)**
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	8.5		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**	**~5**
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	-17.2	-73.4		8.4	8.4	8	S. Peru	0.3

Note: M_b/M_s –body wave/surface magnitude; M_w –moment magnitude; M_t –tsunami magnitude [Abbreviations: S = south; N = north; Chat = Chatham Islands, Main=S. and N. Islands].

Would these events be comparable in size to historical events? Several of the South American tsunamigenic earthquakes (for example, 1746) occurred at locations well north of the 1868, 1877 and 1960 events, some with significantly greater near-source tsunami heights than the 1868 or 1877 events. At present, it is not known whether these locations direct waves towards New Zealand more or less effectively than the 1868 event, which is generally used as a worst-case scenario for distantly generated events. However, the possibility that a location north of the 1868 event will produce a more damaging tsunami than the 1868 event cannot be excluded. Not only the source location, but details of the source (size, orientation), will determine the characteristics of the wave train that reaches New Zealand's continental shelf. Numerical models of the 2001 Peru tsunami (Figure 18) suggest that this location directs waves reasonably well towards New Zealand. The models suggest that New Zealand sites exhibited the highest tsunami response in the wider Pacific to the 2001 Peru event (particularly the Chatham Islands and Banks Peninsula). Measured tsunami wave heights (Goring, 2002) were near to predicted values (about 0.3 m amplitude). Trans-Pacific tsunami models are progressively being applied by GNS to other tsunamigenic sources in South America in order to confirm those areas that pose the highest potential danger to New Zealand.

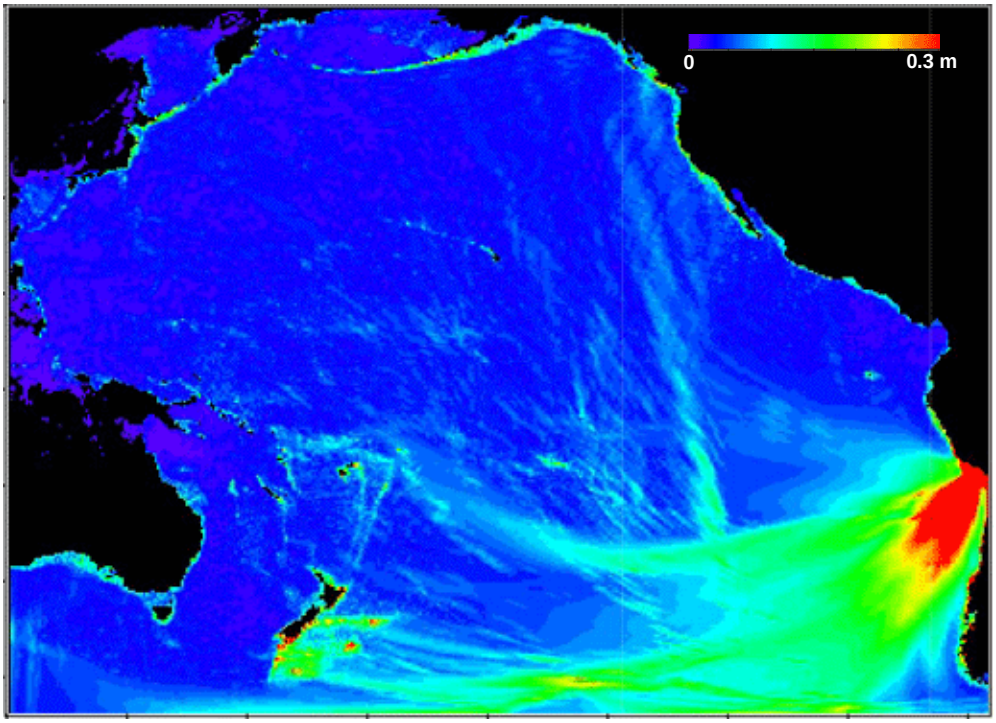


Figure 18: Indicative maximum wave height distribution for the 2001 Peru tsunami. Note that the resolution of the picture does not allow enlargement to get better detail around New Zealand. Note also that only the Pacific Ocean is represented, that is, ocean east of Central America is represented as land. (Image from www.pmel.noaa.gov/~koshi/peru/dcrd/ maximum.gif).

Another potential source of tsunamis in New Zealand are large subduction interface earthquakes along the east coast of Canada/US Pacific Northwest coast (Cascadia region). Credible source dislocation and tsunami wave models for a large earthquake (magnitude M~9) known to have caused a tsunami along the coast in 1700 (identified by paleo-tsunami evidence, as well as Japanese historical records) have recently been developed by Satake (pers. comm., 2003). Recent model simulations of the event show that the wave heights may have been substantial in parts of New Zealand, possibly over 1 m in parts of the Bay of Plenty. More precise estimates of the potential tsunami height require numerical models to be run with more accurate local bathymetry and nearshore topography around New Zealand. The latter are essential to bring numerical tsunami propagation models of Pacific–wide tsunami accurately from deep ocean bathymetry models into shallow continental shelf and coastal waters, and hence incorporate local bathymetry features that may focus incoming waves onto certain coastal areas or cause amplification in harbours and embayments (see Section 6).

The potential of other distant source zones to produce significant tsunami in New Zealand is largely unknown. However, other than some minor damage in Northland caused by the April 1 1946 tsunami from the Aleutian Islands (Alaska), none have been significant in New Zealand in the historical record. Tsunamis from most Northern Hemisphere or western Pacific source zones are likely to be scattered and attenuated by the presence of many islands along the wave propagation paths. However, this needs to be confirmed by further tsunami modelling.

5.2. Regional and local tsunami sources

Developing an understanding of the frequency/magnitude of all local and regional source tsunami events on our continental shelf areas or in the vicinity of New Zealand is a long-term project. Several local sources for the Bay of Plenty and eastern Coromandel have been investigated previously. In particular, de Lange and Healy (1986b), de Lange (1998) and de Lange and Prasetya (1999) investigated eruptive volcanic sources of potential tsunami hazards in the Bay of Plenty and eastern Coromandel. Four main volcanic tsunami source regions were identified (de Lange and Prasetya, 1999):

- Whakaari/White Island is an andesitic volcano that is currently active at the northern end of the Taupo Volcanic Zone. The main crater is breached to the southeast with the crater wall close to sea level and the crater floor beneath sea level. White island has generated small pyroclastic flows and debris avalanches in the past that may have generated tsunamis.

- Tuhua/Mayor Island volcano has a previous history of pyroclastic eruptions and caldera collapse. Although not currently active, the most recent activity was in the last 2000 years.

- A zone of hydrothermal vents between Moutohora/Whale Island and Whakaari/White Island, which could pose a hazard from phreatic (steam-blast) activity.

- Inland Taupo Volcanic Zone volcanic activity. The Okataina Volcanic Centre is the largest volcanic complex close to the Bay of Plenty coast, and has experienced five to six large eruptions in the past 400,000 years. At least one of these events caused large pyroclastic flows into the sea.

However, much of the earlier volcanic tsunami research probably needs updating with new technology (e.g., multibeam bathymetry), high-resolution tsunami models, and the additional knowledge on fault characteristics and deformation patterns. NIWA and GNS are continuing investigative sea-floor studies on the offshore volcanoes embedded in the Kermadec Ridge.

NIWA researchers have also spent the last few years investigating seafloor faulting and underwater slumping/landslides in the East Cape area, Whakatane to Whakaari/White Island area and more recently faulting zones off Tauranga. The following commentary arises out of this research work—some of it at a very preliminary field-investigation stage.

The following potential regional and local sources of tsunami have been identified (Figure 19), and are discussed in detail below.

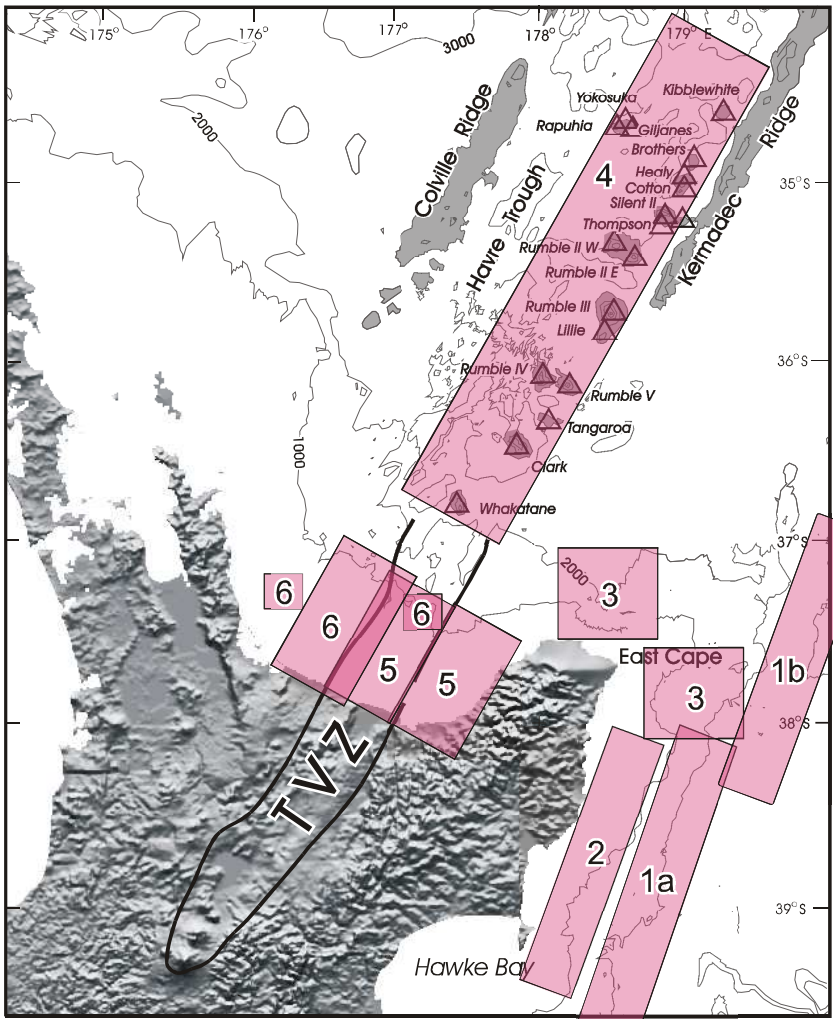


Figure 19: Location map showing the general location of types of potential sources of regional-impact and local-impact tsunamis that could impact on the BOP and eastern Coromandel coasts. Numbers [1–6] correspond to the source types discussed in the text below. Individual tsunami sources are not shown. Note: TVZ = Taupo Volcanic Zone. Figure adapted from Wright et al. (2003).

[1] Subduction interface earthquakes occur in the Tonga-Kermadec-Hikurangi region associated with the Pacific/Australian plate boundary. This source occurs beneath the eastern margin of the North Island and the Kermadec Ridge, where the Pacific Plate underthrusts (subducts) to the west. Because of significant changes in crustal structure along the margin, it is not yet certain if the entire subduction zone is a potential tsunamigenic hazard. The two 1947 M_W7.1 (March) and M_W6.9 (May) tsunamigenic earthquakes off Poverty Bay are possible small analogues for a shallow Hikurangi Margin subduction earthquake source [1a]. However, these were an unusual and relatively rare type of subduction interface earthquake, occurring at very shallow depths near the

initiation of subduction (Downes et al. 2001). The deeper part of the margin (below 12 km) is possibly capable of producing much larger earthquakes, affecting a much greater part of the margin with strong shaking and a tsunami. Cochran et al (submitted) and Chague-Goff et al. (2002) present data consistent with the occurrence of large tsunamis in association with large subduction interface earthquakes in the Hawkes Bay sector of the Hikurangi margin. However, tsunami attack in the BOP and eastern Coromandel from the Hikurangi Margin relies upon coastal-trapped wave propagation northwards along the coast and around East Cape. Historically, there is no record of shallow seismicity in the Tonga–Kermadec subduction zone [1b] producing significant tsunamis in New Zealand, but the potential for tsunami generation requires further evaluation.

[2] There are many upper-plate faults in the northern Hikurangi margin, some of which may be capable of significant tsunami generation. These include reverse faults and associated folds beneath the upper margin of Hawke Bay (e.g., Barnes et al. 2002) and Poverty Bay (Foster and Carter, 1997). As with subduction zone earthquakes in this region, these require a coastal-trapped wave to travel around East Cape to be a viable tsunami source in the Bay of Plenty. Numerical modelling is required to ascertain the possible effectiveness of such coastal-trapped waves to travel around East Cape into the Bay of Plenty.

[3] Landslide sources in the Hikurangi margin include giant complex landslides such as Matakaoa and Ruatoria which may be, but are not necessarily, triggered during large earthquakes. Collot et al. (2001) showed that the Ruatoria landslide east of East Cape was triggered some 170 ± 40 kyr[7] ago, resulting in a >3000 km^3 blocky debris avalanche and debris flow that travelled up to 100 km from its source region. The Matakaoa landslide complex north of East Cape comprises at least three large landslides (Carter and Lamarche, 2001). These include giant slabs of sedimentary strata that slid down slope semi-coherently, and debris flows that extend up to 200 km from source. Such large events could have very long return times of 10's–100's of thousands of years. By analogy with other seismic regions in New Zealand, however, more frequent but smaller landslides with volumes typically <0.5 km^3, are likely within the Matakaoa complex and in the submarine canyons of Bay of Plenty. Further mapping of the BOP continental margin is required to determine whether landslide scars are present and in what frequency, and modelling is required to determine what dimension and emplacement mechanism of a landslide would result in tsunami inundation along the BOP coast.

[7] kyr is a unit for 1,000 years

[4] Undersea volcanism in the Tonga-Kermadec system (and more distant) is another potential source of tsunamis. At least 23 submarine volcanoes of the active southern Kermadec arc (where the edifice diameter is >3 km) occur within 400 km of the Bay of Plenty coastline (Wright et al. submitted), three of which (Rumble II West, Brothers and Healy) are silicic calderas (Gamble and Wright, 1999). Of these, Healy is interpreted as having been formed by catastrophic submarine pyroclastic eruption with the destruction of 2.4-3.6 km^3 of the proto-edifice and formation of a 2–2.5 km wide, and 250–400 m deep caldera (Wright et al. 2003). Larger silicic centres such as Havre and Macauley, with 8–10 km wide and 500–700 m deep calderas are known further north along the Kermadec Ridge volcanic arc within 970 km of the coastline (Wright et al. submitted). Data assembled for this project indicate that one or possibly two paleo-tsunami events inferred for the Bay of Plenty and eastern Coromandel at around AD1302–1435 may be associated with eruption and/or collapse of the Healy caldera in the Kermadec Ridge (Wright et al, 2003). One of the paleo-tsunami events occurs in association with Loisels Pumice, interpreted to be from the Healy caldera (Section 4). This event could perhaps coincide with collapse of the Healy caldera cone. There appears to have been an earlier event in the Loisels Pumice-related period (AD1302–AD1435) that may have been associated with the initial (or subsequent?) submarine eruptions at Healy. Wright et al. (2003) note that the chemistry of only some of the Loisels Pumice types are similar to the Healy caldera, so clearly there are other Kermadec Ridge sources active in about the same interval, and likewise are potential tsunami sources. The Pacific tsunami database records 58 volcano-generated tsunami since 1700AD, but it is unclear how many of these are from submarine volcanism and generated a regional tsunami.

Seafloor multibeam mapping reveals many of the 23 southern Kermadec volcanoes undergo cycles of edifice construction and destruction. The latter largely occurs as submarine sector collapse (landsliding) of the volcano flanks. Volumes of each sector collapse for every volcano are unknown. However, potentially there is up to 4–5 km^3 for a single collapse, as shown from Rumble III volcano (Wright et al. 2004) which lies ~290 km from the Bay of Plenty coastline. It is presently unknown whether such collapses are large single catastrophic events or small repetitive movements.

[5] Regional active faults provide many candidate sources for Bay of Plenty and eastern Coromandel tsunamis. They include normal faults in the offshore Taupo Volcanic Zone, both on and off the continental shelf, and the northern offshore extensions of the North Island Dextral Fault Belt. The major zone of active rifting extends between Whakatane and Tauranga, with faults between Matata and Whakatane accommodating a significant proportion of the total crustal

extension (Wright, 1990; Lamarche et al. 2000). The larger faults with significant seafloor traces include the Whakaari/White Island and Rangitaiki Faults in the offshore Whakatane Graben. Normal faulting in the Taupo Volcanic Zone rarely exceeds 2 m single event displacement, but the larger boundary faults may be capable of larger surface ruptures. The possibility that fault rupture with modest displacement could generate destructive tsunamis is uncertain, and requires numerical modelling. Similarly, whether faulting on or off the continental shelf is more effective at generating tsunamis requires modelling to appreciate the effectiveness of these potentially numerous sources.

The North Island Dextral Fault Belt includes the Whakatane, Waimana and Waikaremoana faults, which have significant total displacement, and currently accommodate some of the regional extension. These faults extend offshore for 20-30 km. The c. 2500 yrs BP subsidence event with evidence for coeval tsunami inundation at Ohiwa and Jacob's Creek suggests a local fault source. However, the Ohiwa site is on the upthrown side of the recent trace of the Waimana fault and the Jacob's Creek site is east of and on the upthrown side of the next fault, the Waikaremoana fault. This geologic setting is thus inconsistent with the subsidence observed at both sites. In addition, it seems likely that vertical throw in an individual event would rarely exceed 2 m, and it is thus questionable that subsidence at both of these sites represents tectonic movement. Equally it is uncertain if 2 m of displacement of the seafloor on the continental shelf could generate a widespread, large, tsunami as far away as Waihi Beach, given the indication that tsunamis of at least 5 m wave height are required to leave mappable deposits. Perhaps fault rupture triggered a large landslide at the edge of the continental shelf and this was the source of the widespread tsunami. Several active faults with an approximate east-west strike enter the BOP from the coastline from the Motu River north to Te Kaha. At least one has been mapped offshore. These faults are downthrown to the southwest, but the size and recurrence of individual displacements is not known. However, their location, and sense of motion is well oriented to generate local-impact tsunami events around the BOP and Coromandel coastline.

[6] Offshore volcanic sources in the BOP include Tuhua/Mayor Island and Whakaari/White Island. Whakaari/White Island has been considered previously and largely discounted for tsunami generation potential due to its deep-water location and any tsunami produced is likely to propagate away from the coast (de Lange, 1983; 1998). However, for Tuhua/Mayor Island, past modelling (de Lange, 1998; de Lange and Prasetya, 1999) indicates that the credible pyroclastic eruptions of a "Mt St Helens" scale (1 km^3) could produce a tsunami that impacted Tairua to Maketu, peaking at 0.5 m between Whangamata and Tauranga. An eruption ten times bigger with a pyroclastic flow of Krakatau

scale (10 km^3) would peak at around 5 m. Recent geophysical data indicates that tsunami potential could be significant. Tuhua/Mayor Island is the emergent summit of a ~15 km wide and 750 m high caldera volcano. The volcano has produced both explosive and effusive eruptions, including three phases of caldera collapse. The last of these caldera collapses, associated with the largest eruption, occurred about 6,300 years ago (Houghton et al. 1992), and included the transport of large pyroclastic flows from the sub-aerial volcano into the sea. This 6,300-year old event is presently the only recorded instance of pyroclastic flows entering the sea within the New Zealand region and is likely to have produced some form of tsunami. There is a possible causal link between Tuhua/Mayor Island pyroclastic flows entering the sea and the ~6300 yr BP event preserved in the sediment cores from Waihi Beach (Section 4.2).

Like Tuhua/Mayor Island, Whakaari/White Island is the emergent summit of a larger, mostly submarine ~17 km wide edifice. Sub-aerial eruptions have included both effusive and small explosive eruptions of mostly andesite, but including dacite, though the volcanic history of the volcano is poorly known. A small collapse of the inner west wall of the main crater in 1914 produced a debris avalanche that is interpreted to have entered the sea. The volcano sits atop a larger massif with its northern flanks extending into deep water. The volcano has an active hydrothermal system which may weaken the edifice structure and enhance sector collapse on both the outer sub-aerial and submarine flanks. Numerous smaller submarine volcanoes occur on the Bay of Plenty continental shelf and slope (Gamble et al. 1993) lying within 100–150 km of the coastline, including a number of silicic centres at least one of which is a small caldera (Mahina Knoll) with associated pumice deposits. It is highly probable that forthcoming multibeam mapping of the Bay of Plenty continental shelf and slope will “discover” other volcanoes.

[7] Tsunamis (rissaga) generated by atmospheric pressure-waves or pyroclastic flows from large onshore volcanic eruptions in the Taupo Volcanic Zone or TVZ (e.g., Okataina Volcanic Centre) or of Mt. Taranaki are other possibilities. The potential for these is little known, but the direct volcanic impact is likely to overwhelm the additional impacts and consequences of any associated tsunamis in the Bay of Plenty and eastern Coromandel.

6. Tsunami amplification and resonance along the coast

6.1. Amplification and resonance

Tsunami run-up depends on the initial or incident wave height and direction, the wave periods and wavelengths in the wave train, and how the waves interact with the coastal/shelf seabed, topography of the beach and shore-margin, and the planform (birds-eye) shape of the coast or harbour. Offshore and continental-shelf bathymetry can focus and amplify (refract) a tsunami wave-train from a "regional" or "distant" source, beyond more than just the natural increase in height when a tsunami wave slows down as it reaches shallow water. Such focusing will determine the overall tsunami height and impact on a region or country.

However another localised effect, called resonance, can substantially modify the incoming tsunami wave-train at any locality along the coast or in harbours and estuaries. Each part of the coast (bights, bays, harbours, ports or river mouths) has its own set of natural or resonant frequencies (see box below). If an arriving tsunami wave-train comprises wave periods (time between wave crests) that match the natural resonance period (or some harmonic) of the locality or region, these wave periods will be selectively "picked out" and amplified causing higher wave heights and run-up at the shoreline and continuing excitation of the water body, compared with other locations where there is no match-up with the natural resonant frequency. This explains why some areas experienced greater damage or run-up than other areas in past historical "distant" tsunami events (e.g., the South American tsunamis of 1868, 1877 and 1960) even though the same stretch of Bay of Plenty-eastern Coromandel coast was exposed to the incoming tsunami. One example of this resonance effect is aptly described in Appendix 1 for Mercury Bay during the 1960 Chilean tsunami, when wave oscillations occurred at around 40 minutes (close to the natural resonance for Whitianga/Buffalo Beach) for at least five days before they were damped out.

Definition—**resonance** is an increase (amplification) in the oscillatory energy absorbed by a system when the frequency of the oscillations matches the system's natural frequency of vibration (or its *resonant frequency*). This is most evident in an aquarium, swimming pool or bath, where cyclic movements (disturbances) of the water mass at the natural frequency of the container will quickly cause water to spill over the sides. An object or coastal water body often has several frequencies at which it will naturally resonate, especially if it is elongated in plan shape (rather than circular).

One objective of this report is to describe the resonance characteristics of the Bay of Plenty and eastern Coromandel for the range of wave periods typically spanned by tsunami waves, and what relevance they have for identifying the more vulnerable locations within the area bounded by East Cape (south) and Colville Channel (north).

In order to understand the section on resonance patterns in Bay of Plenty some general comments need to be made on the occurrence of resonance in general, and on the range of periods present in distant, regional and local tsunami.

6.2. Resonance: causes and effects

As discussed above, the occurrence of resonance is dependent on the range of wave periods present in the incident tsunami waves and on the duration or persistence of the tsunami. Within the series of waves that comprise a tsunami wave-train there will be waves spanning a range of different periods. These reflect both the large-scale features and small-scale details of the disturbance generating the tsunami and the wave dissipation and focusing processes that occur along the way. Features of very large aerial extent, such as a large earthquake with a source of width of 30 to 50 km and fault length of 100s of km, produce long periods (typically 10 to 60 minutes or more) and long wavelengths. At the other end of the scale, landslide or volcanic sources produce shorter, irregular wave periods (typically 1 to 20 minutes) because the horizontal extent (area) of the disturbance is relatively small. However, the wave height can be quite large over localised areas. Occasional very large landslides, such as the Ruatoria submarine landslide event off East Cape, have much larger dimensions and therefore generate longer periods more akin to earthquake fault ruptures.

Within any large source there will be smaller scale features that produce other tsunami waves with shorter periods and wavelengths. For example, in an earthquake-generated tsunami, the amount of movement on the causative fault may vary considerably from one part to another producing variable uplift, rupture may also occur on secondary faults, or submarine/coastal landslides or slumps may be triggered—all producing their own set of waves of different periods. Near the source, these smaller features can affect the local run-up and impact of the tsunami on the nearby shore.

Along the coast nearest a “local” tsunami source, the first few waves would almost certainly cause the greatest run-up opposite the source area. Hence there is little opportunity for resonance to amplify that run-up. However, further down- or up-coast from a local source, the formation of multiple coastal-trapped waves that travel along the coast can be substantially enhanced (amplified) if the wave periods match the natural resonant period of the locality they pass through. This is more likely to occur in more distant enclosed harbours/estuaries and pocket bays, where multiple reflections and interactions with nearby shores can increase the duration (persistence) of the tsunami. A prolonged duration at any locality due to resonance may also provide the opportunity for later, possibly smaller, waves in these areas to be more damaging because of coincidence with a higher tide. In particular, resonance may be a significant factor in locations such as Coromandel, Cook Strait, Hawke Bay or Banks Peninsula. The 1855 earthquake near Wellington, for example, produced a tsunami that caused Wellington Harbour and Cook Strait to oscillate for 8 to 12 hours, the

largest waves in parts of the harbour and Cook Strait probably occurring some hours after the first waves because of resonance (see Downes et al. 2000).

The situation is different for distant source tsunamis. Tsunamis capable of producing damaging waves a great distant from the source are almost invariably caused by very large subduction interface events involving very large source areas. The longer periods and longer wavelengths produced by these sources dominate the wave train when it arrives at distant shores. Typical periods for tsunami arriving in NZ from distant sources such as South America range from 30 minutes to a couple of hours. While run-up heights from locally or regionally generated tsunami have the potential to be much greater compared with "distant" tsunami source events (i.e., can range up to 15 m or more), run-up heights from distant source tsunamis can reach substantial levels at some locations primarily because of local resonance. Other influences that play a role are the directional focussing of the tsunami wave-train, either near the source, along the deep ocean propagation path, on the continental shelf around New Zealand or major headlands or islands around the coast. Historically, the maximum known run-ups in New Zealand from distantly-generated tsunamis are about 7–10 m on Chatham Island, and about 4–5 m at several harbour locations on Banks Peninsula (e.g., Lyttelton Harbour).

6.3. Determining areas of resonance (Bay of Plenty and eastern Coromandel)

Identification of areas in the Bay of Plenty and eastern Coromandel that have a high potential to resonate with an incoming tsunami was carried out by a combination of numerical modelling, analysis of sea-level data and observations from past events.

The primary tool was a high-resolution hydrodynamic model used to simulate the coastal effects of incoming regular waves of a constant height and a constant specified period from the eastern Pacific Ocean (Walters, 2002). Different model simulations were run covering a range of wave-train periods from 15 minutes to 5 hours.

Therefore the results from the model are most applicable to "distant source/national impact" tsunamis, because these tsunamis are more effective at inducing resonance, and the effects are more predictable than for "local source/local impact" tsunami. However, resonance analysis has some predictive value for identifying where longer-duration effects at more enclosed parts of the coast might occur for either "local-source/local-impact" or "regional source/regional impact" tsunamis. The resonance analysis has no predictive capability to determine the magnitude of the tsunami hazard of the first few waves on the coast nearest a "local" tsunami source.

For the simulations presented in the next section, the distant-source wave-train propagates in from the eastern boundary and after reflecting off the coast is allowed to radiate out again straight through all the model boundaries to avoid bouncing back off

the model "walls". A close examination of the results for the entire grid indicated that the radiation conditions used on the model's boundaries are effective in that little outgoing wave energy is re-reflected back into the modelled area, as one indeed would expect to happen in reality. The waves from the east bend around the north and south ends of New Zealand and propagate through Cook Strait. Thus this one model scenario (for a wave-train from the east) yields resonance characteristics for most areas of New Zealand's coast except parts of the west coast of both North and South Islands.

At this stage, the model is only applicable to open coastal areas, with much less resolution of the seabed bathymetry inside harbours and estuaries. Consequently, the results are not really reliable inside harbours (e.g., Ohiwa and Tauranga Harbours). [8]

In the model results presented in the next section, the pattern of amplification is of primary importance, not the exact amplitudes or wave heights. In essence, the spatial pattern shows the areas of large amplification of an incident tsunami wave of that period. These areas are then potential critical locations for tsunami resonance and are therefore more vulnerable parts of the coast. The actual amplitude that would be observed depends upon the height of the incident wave, which is in turn dependent on the details of the generation of the wave. Here, we look at the broader pattern of resonance (and by implication tsunami inundation hazard) rather than the details of individual events or simulations. In identifying the more vulnerable locations, it is important to recognise that other locations are not without hazards from distant source tsunamis. Even a tsunami that is not amplified by resonance may still be dangerous (e.g., a run-up of 1 m is considered dangerous, especially coinciding around high tide).

Other sources of resonance information were obtained from data analysis of sea-level gauge data published by Heath (1976) and Smith (1980), and recent unpublished sea-level data collected by NIWA in Mercury Bay from July to November 2002. For these modern datasets, an analysis of how the local sea level responds to other forcings (e.g., wind and storms) provides values for the various natural resonant periods of that water body.

6.4. Discussion of Bay of Plenty and eastern Coromandel resonances

Model results for wave periods of 75 and 90 minutes are the only ones presented here (Figures 20–21), being the most relevant for the Bay of Plenty and Coromandel region.

As expected, resonance patterns on the open coast show an increase in length between alternate resonant nodes (shown by coloured patches) with increasing period of the

[8] The model is undergoing continual improvement as better bathymetry for harbours and estuaries are successively added to the model grid.

incident waves from 75 to 90 minutes (Figures 20–21). As the length between resonant nodes changes, the coastal response and wave amplification then changes in predictable ways (Walters, 2002).

In general, for areas with resonant behaviour, the sequence of figures shown in Walters (2002) gives a distinct sequence of patterns. For periods larger than the largest local natural resonant period there is little amplification. As the wave period decreases to approach the natural resonant period of a particular embayment, harbour or bight, the wave height amplifies to a maximum with a spatial pattern that contains a single maximum node in that area. As smaller wave periods are considered, the wave amplification decreases until it approaches the next resonant period (first harmonic) where it again amplifies the incoming wave and shows a pattern with two maxima in the embayment or bight. This behaviour is repeated for all the remaining resonant periods (harmonics) in the range of wave periods used in the model.

There are no significant long period resonances in Bay of Plenty (from 120 to 300 minutes). At a period of about 90 minutes (Figure 20), the fundamental resonance for the Whangapoua embayment is excited and on up past Port Charles, as well as lesser amplification around Waihi-Tauranga-Papamoa and Whakatane-Opotiki areas as part of a 3-node resonance pattern between Cape Colville and East Cape (Figure 20). The geometry around Tauranga Harbour is not resolved very well, so the results inside the harbour are unreliable. At a wave period of 75 minutes (Figure 21), the wavelength is shorter and the 3 resonance nodes in Bay of Plenty are closer together. The largest amplification for 75-minute period waves is now in Mercury Bay (Whitianga) and central Bay of Plenty between Mt. Maunganui/Mauao and Maketu. For shorter periods below 75 minutes, the resonance patterns become increasingly more complicated with more amplified nodes spaced along the coast. Below 30 minutes, the patterns become less accurate because the wavelength is approaching the cut-off limit for the numerical model grid. All of the sites mentioned above feature in the visual observations of effects and impacts by eyewitnesses during historical tsunami events described in Section 3.

For Bay of Plenty, the modelled resonance patterns shown in Figures 20–21 are mainly applicable to distant tsunami because these long wave periods are more typical of large subduction zone events such as along the west coast of South America. However, harbours and estuaries in the region may resonate with local tsunami events that comprise shorter wave periods. For example, Tauranga Harbour has a natural resonance down at 20-minute periods (Heath, 1976).

The spatial patterns from the model replicate the observed patterns in historical tsunamis along the east coast of both islands (Walters and Goff, 2003). For instance, consider the 1960 Chilean tsunami, one of the most extensive tsunamis recorded in New Zealand (de Lange and Healy, 1986b; Heath, 1976). The resonance pattern

predicted by the model in Pegasus Bay and Lyttelton Harbour indicates a strong peak at around 150 minutes and the Chatham Rise tends to act as a waveguide directed at the Canterbury coast (Walters, 2002). The sea-level data also show a similar amplification of the long wave-period part of the wave energy spectrum in the 1960 tsunami (Heath, 1976) and in the smaller 2001 Peru tsunami (Goring, 2002). On the other hand there is little response around the Otago Peninsula either in the model results or sea-level data. Wellington Harbour responded in terms of its dominant resonant modes, one with a period of about 160 minutes and several with periods around 30 minutes (Heath, 1976; Abraham, 1997; Walters, 2002). The response at Hawkes Bay (de Lange and Healy, 1986b) also follows the predicted response patterns with wave periods near the resonance peaks of 150 minutes and shorter.

In Bay of Plenty, the response at Tauranga (Heath, 1976) also follows the predicted response patterns with wave periods near the resonance peaks of 60 minutes and shorter. Finally, sea-level records from Mercury Bay in 2002 (unpublished NIWA data) show there are natural resonant periods of 32–37, 44, 50, 60 and 70–76 minutes, all of which cover the range likely for a distant-source tsunami event. These results from field data match the patterns from the resonance model results for 75-minute waves (Figure 21) and also the analysis of Whitianga Wharf sea-level gauge data by Smith (1980). The shorter resonant periods (32–37, 44 minutes) are only present within the inner Mercury Bay area off Buffalo Beach and the Whitianga Estuary, but decay quickly further out in Mercury Bay.

To a large degree, these resonance patterns predicted by the numerical model will reflect the type of response to a distant tsunami event. Both the sea-level gauge data and eyewitness accounts tend to confirm the resonance patterns that have been computed by the model for tsunami wave-trains from the east. This provides support for the fact that areas highlighted by the model as showing substantial amplification, are the areas most vulnerable to tsunami hazards, notwithstanding that any tsunami wave that reaches the coast with run-up heights of >1 m will be dangerous along many parts of the coast.

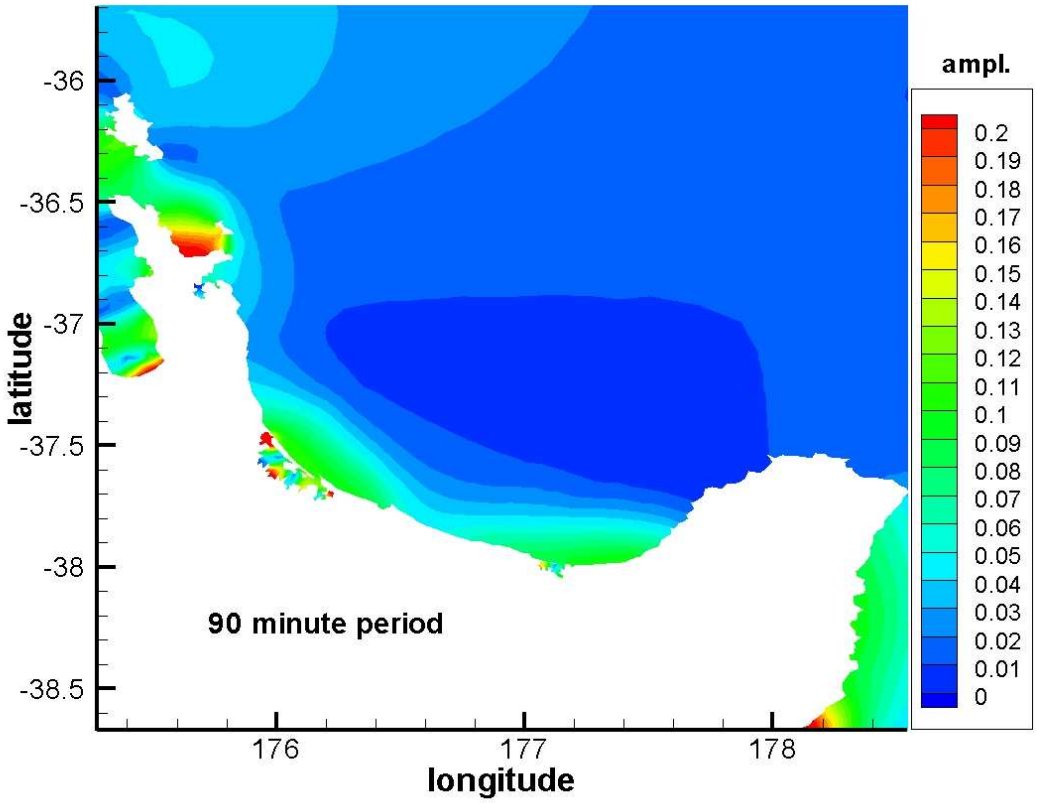


Figure 20: Amplification for an incident wave from due east with a period of 90 minutes. Substantial resonance occurs around Waihi-Tauranga-Papamoa, Mercury Bay, Whakatane-Opotiki areas and particularly Whangapoua area. *Note:* the amplification (ampl.) scale of 0.1 means, a 10-fold increase in the offshore wave height, and 0.2 is a 20-fold increase (red).

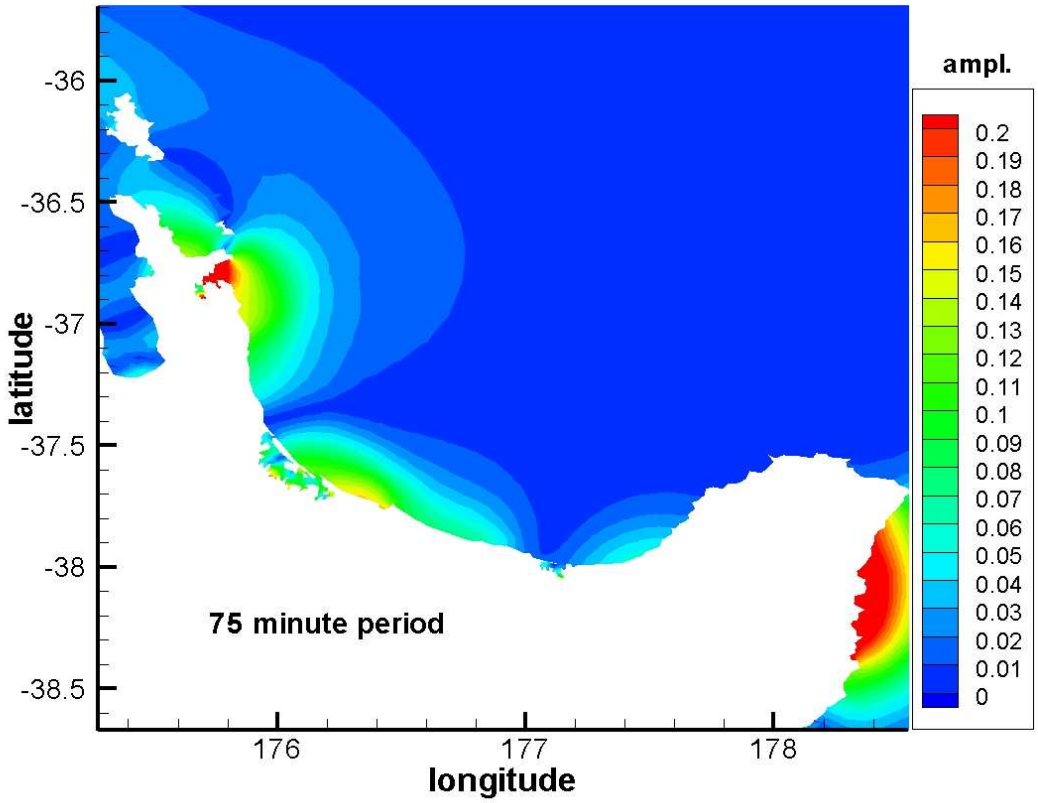


Figure 21: Amplification for an incident wave from due east with a period of 75 minutes. Substantial resonance occurs in Mercury Bay (including Whitianga), with lesser amplification in other eastern Coromandel area, Tauranga to Maketu. *Note:* the amplification (ampl.) scale of 0.1 means, a 10-fold increase in the offshore wave height, and 0.2 is a 20-fold increase (red).

6.5. Evaluation of the tsunami potential from local and regional sources

In the Bay of Plenty and eastern Coromandel, there appear to be numerous sources for local tsunami as indicated in Section 5. The more important sources appear to be those that can propagate a direct wave into this area using refraction from the local bathymetry. Coastal-trapped waves from sources in adjacent regions (north of Great Barrier Island and south of East Cape) would not appear to be significant because of wave scattering and dispersion that would take place at East Cape (northward propagating) and in Hauraki Gulf (southward propagating).

In addition, the horizontal size and directivity of the disturbance and the location of the "local" source play important roles in the ability of the wave to propagate directly onshore without dispersing substantially. For moderate size sources such as a thrust fault or submarine landslide located on the continental slope, wave dispersion effects

can become important and the wave-train will tend to separate. Directivity of the tsunami-generating disturbance from a fault rupture or pyroclastic flow will be a major factor in determining which areas are most vulnerable to different regional and local sources in the Bay of Plenty and eastern Coromandel region. For more distant "regional" sources such as underwater volcanic events in the Kermadec Ridge, the waves would propagate in a circular pattern and wave height would decrease quickly.

In the end, many factors must be considered in determining whether a particular local or regional source can produce hazardous waves at particular coastal sites. Such work has been done recently by NIWA modelling a credible fault rupture source and a submarine canyon-landslide source off Kaikoura for a tsunami hazard and risk assessment for Environment Canterbury (Walters et al. 2004).

7. Summary of the tsunami hazard for Bay of Plenty and Coromandel

7.1. Tsunami hazard study

Tsunamis can be classified into categories either by the distance from their source to the area impacted, or more relevant for emergency management purposes, the travel time to the impacted area and the length scale of impact. For this report, three categories are defined:

- local source/local impact event (within say 30 to 60 minutes travel time and affecting several 10's of km of coast);
- regional source/regional impact event (within 3 hours travel time and likely to affect most of the Bay of Plenty and eastern Coromandel);
- distant (remote) source/national impact event (longer than 3 hour travel time and likely to affect several regions).

The development of a credible tsunami hazard profile for the Bay of Plenty and eastern Coromandel covering all three categories of tsunami has been undertaken by combining data and information from distinctly different sources. These include sea-level and tsunami run-up data, eyewitness accounts, marine geophysical surveys, paleo-geological investigations of undisturbed sediment cores inland from the coast and numerical modelling of tsunami resonance behaviour in the overall region.

7.2. Paleo-tsunami record

Given that the post-European historic record is relatively short (160 years), geological field investigations and geo-chemical analysis of sediment cores opens up the possibility of detecting, interpreting and dating large paleo-tsunami events to extend the tsunami hazard record for the region. Year 1 of the Tsunami Hazard Study focused on selecting and obtaining cores from potentially undisturbed sites in the Bay of Plenty and eastern Coromandel (Goff, 2003).

Locations for the paleo-tsunami field sites investigated were: Otama Beach (near Whangapoua); Waihi Beach, Ohiwa Harbour; and Jacobs Creek (between Waiotahi Estuary and Opotiki). Evidence has also been gleaned from additional sites that have been investigated in previous paleo-tsunami studies in both the eastern Coromandel and the Bay of Plenty (Goff, 2002a, b).

By combining detailed visual, geological, geochemical and radio-carbon analyses of the sediment cores and expert interpretation of the results based on the geological context of the Bay of Plenty and Coromandel region (e.g., volcanism, tectonic uplift, Holocene sedimentation), a record of probable paleo-tsunami events has been developed.

In summary, over the past 4000 years a total of two major regional-impact paleo-tsunami events have been recorded in sediment cores—one in AD1302–AD1435 (with some evidence for two separate major events in this period) and an earlier event at 2500–2600 years BP. There is also evidence in various sediment cores that up to four local paleo-tsunami events have impacted localised areas of the Bay of Plenty in AD1600–AD1700 (local subsidence a factor?), AD1200–AD1300, 1600–1700 years BP, and 2900–3000 years BP.

A key point of these paleo-tsunami investigations is that the resolution used is only capable of identifying tsunami events with run-up height larger than 5 m (Goff, 2003).

7.3. Historical tsunami record

In historical times (since 1840), tsunamis are known to have affected places along the Bay of Plenty and eastern Coromandel coastline at least eleven times. The historical eyewitness and newspaper accounts of the behaviour and impacts of these tsunami events are detailed in Table 1 (Section 3).

Information on the historical tsunami events is based on data and information built up over the last two decades, particularly from de Lange and Healy (1986b), Fraser (1998), and data in the GNS Tsunami Database. The latter revises and updates the earlier databases with new accounts found as the result of recent comprehensive investigation of historical newspaper accounts.

The most substantial tsunamis to have affected the Bay of Plenty and eastern Coromandel areas in the last 160 years were generated by "remote" or distant sources. The largest, in 1868, 1877 and 1960, were generated by very large earthquakes in the subduction zone along the Chile and southern Peru coastlines of South America—directly opposite and facing New Zealand's eastern seaboard. A further event occurred in August 1883, probably generated by an atmospheric pressure wave from the Krakatau eruption in Indonesia. It caused run-up heights of up to 1.8 m in the Bay of Plenty–Coromandel region. In pre-European history, there are indications that a large earthquake off the Cascadia region (east coast of Canada/US Pacific Northwest coast) in 1700 could have impacted New Zealand. Recent overseas model simulations of this event (using paleo-tsunami evidence, as well as Japanese historical records) show that the wave heights may have been substantial in some regions of New Zealand, possibly over 1 m in parts of the Bay of Plenty and Coromandel. Further modelling is required to confirm better estimates of the run-up in New Zealand.

Since European settlement around 1840, no "local source/local impact" or "regional source/regional impact" events are known to have affected the Study region. However, this is not unexpected as fault ruptures tend to have return periods of 100s to 1000s of years and volcanic eruptions, return periods of 1000s to 10,000 years or more.

7.4. Regional tsunami hazard profile

Table 4 summarises the past tsunami hazard profile of the Bay of Plenty and eastern Coromandel by combining known historical tsunami events (back to 1840) and the paleo-tsunami events identified in this Study over the past 4000 years.

Table 4: Summary of the known past tsunami events across both the eastern Coromandel and Bay of Plenty region (combining the short historical record back to AD1840 and paleo-tsunami records back 4000 years). Note: BP = before present.

Tsunami run-up height (est.)	**<0.5 m***	**0.5–1 m**	**1–3 m**	**3–5 m**	**>5 m**
No. of events:	>6	1	4–5	?[#]	5 or 6
Year(s):	June 2001	Nov 1922	May 1960		*Regional*
	July 1998		Aug 1883		AD1302–1435
	Oct 1994		May 1877		2500–2600 y BP
	June 1977		Aug 1868		*Local*
	Jan 1976		1700?		AD1600–1700?
	Mar 1964				AD1200–1300
					1600–1700 y BP
					2900–3000 y BP

*Many events of <0.5 m run-up may have occurred, but gone unnoticed before instrumental sea-level records became available.

[#] No events >3 m run-up in the historical records, and paleotsunami analysis at this stage can only resolve events with >5 m run-up.

Based on the results from Table 4, it would appear that the return periods for given tsunami heights for Tauranga Harbour in Fraser (1998) and de Lange and Fraser (1999), which are also listed in Table 2, are probably too high, especially if Tauranga return periods are applied to the entire Bay of Plenty and Coromandel coast. This conclusion is based on the five or six local-impact and regional-impact events from the paleo-tsunami record that may have occurred over the past 4000 years, and a further four historical events that have produced run-up heights between 1 to 3 m since 1840, with a further event possibly reaching this run-up height in 1700. Further analysis is required to determine a realistic tsunami return period profile for various sectors of the region's coastal margin.

7.5. Local and regional sources of tsunami

While the historical record since 1840 contains no known tsunami events generated from local or regional sources, the paleo-tsunami record contains at least five or six events, most of which may have been caused by such sources.

A comprehensive summary of sources that could potentially generate a tsunami event has been complied from previous and recent geophysical investigations including seafloor mapping and seismic profiling of faulting systems, plate tectonics and

subduction, underwater volcanoes and underwater landslides. On a national scale, Bay of Plenty and eastern Coromandel face quite a diverse range of potential sources for generating a local-impact tsunami (within say 30 to 60 minutes travel time and affecting several 10's of km of coast) or a regional-impact tsunami (within 3 hours travel time and likely to affect most of the Bay of Plenty and eastern Coromandel).

Potential tsunamigenic sources (local-impact and regional-impact) are:

a) Subduction interface earthquakes occur in the Tonga-Kermadec-Hikurangi region associated with the Pacific/Australian plate boundary;

b) Many upper plate faults in the northern Hikurangi continental shelf margin from Mahia to Ruatoria, some of which may be capable of substantial tsunami generation south of East Cape. However, earthquakes in this region are unlikely to cause large tsunami impacts in the Bay of Plenty as coastal-trapped waves travelling northwards along the coast would be substantially dissipated as they moved around East Cape into the Bay of Plenty;

c) Landslide sources in the Hikurangi margin off East Cape include giant complex landslides such as Matakaoa and Ruatoria, which may be, but are not necessarily, triggered during large earthquakes;

d) Undersea volcanism in the Tonga-Kermadec system (and more distant) is another potential source of tsunamis. At least 23 submarine volcanoes of the active southern Kermadec arc occur within 400 km of the Bay of Plenty coastline, three of which (Rumble II West, Brothers and Healy) are silicic calderas;

e) Regionally active faults provide many candidate sources for Bay of Plenty and eastern Coromandel tsunamis. They include normal faults in the offshore Taupo Volcanic Zone, both on and off the continental shelf. The major zone of active rifting extends between Whakatane and Tauranga, with faults between Matata and Whakatane accommodating a significant proportion of the total crustal extension. The larger faults with significant seafloor traces include the Whakaari/White Island and Rangitaiki Faults in the offshore Whakatane Graben;

f) Offshore volcanic sources in the Bay of Plenty and south-eastern Coromandel, include Tuhua/Mayor Island and Whakaari/White Island. Whakaari/White Island has previously been discounted for tsunami generation potential due to its deep-water location and any tsunami produced is likely to propagate eastwards away from the coast. Forthcoming multibeam mapping around Whakaari/White Island will provide more updated information on the

potential tsunami hazard from this source. However, for Tuhua/Mayor Island, previous modelling studies by the University of Waikato indicate that the credible pyroclastic eruptions could produce a tsunami that impacts an area from Tairua to Maketu. There is a possible causal link between Tuhua/Mayor Island pyroclastic flows entering the sea and the ~6300 yr BP event preserved in the sediment cores from Waihi Beach (Section 4.2);

g) Tsunamis (rissaga) generated by atmospheric pressure-waves or pyroclastic flows from large onshore volcanic eruptions in the Taupo Volcanic Zone (e.g., Okataina Volcanic Centre) or of Mt Taranaki are other possibilities. The potential for these is little known, but the direct volcanic impacts are likely to overwhelm the additional impact and consequences of any associated tsunami in the Bay of Plenty and eastern Coromandel.

7.6. Locations in the region vulnerable to distant or regional source tsunami

Incoming tsunami waves from a distant or regional source can "pick-out" and excite the natural resonant period of a harbour or bay, causing the wave to amplify in height and persist longer in certain areas compared with other parts of the coast. This pattern of more vulnerable areas due to resonance effects changes with the wave periods present within any given tsunami. This means different tsunami events may preferentially impact different areas than those impacted by previous events, especially if a distant South American tsunami is compared with a regional tsunami event. However combining resonance modelling for tsunami wave periods of 75 and 90 minute oscillations with historical accounts of tsunami damage and wave observations, has highlighted some areas of the Bay of Plenty and eastern Coromandel region that are potentially more vulnerable than other areas. These are listed below, but only apply to "distant source/national impact" and "regional source/regional impact" tsunamis.

Highest vulnerability:

- Open coast from Otama Beach to Port Charles and out to Great Mercury Island (especially Whangapoua embayment and Port Charles).
- Mercury Bay (especially Whitianga).
- Open coast between Mt. Maunganui/Mauao and Maketu (especially Kaituna River and Maketu).
- Open coast between Matata and Torere (especially river entrances e.g., Opotiki, Torere).
- Papatea and Whangaparaoa Bays near Cape Runaway.

Moderate vulnerability:

- All other open coast areas.
- Tauranga Harbour?
- Ohiwa Harbour?

Further high-resolution modelling is required to ascertain the relative vulnerability of harbours, estuaries and river mouths to “distant” or “regional” tsunami sources. More geophysical information is required to rank the various possible sources of tsunami generation, and additional tsunami wave modelling is needed before relative vulnerabilities of areas in the region can be determined. However, accurate modelling of tsunami behaviour along the coastal margin, including rivers and harbours and overland flow, will depend on the acquisition of high-resolution bathymetry and land topography.

Finally, a tsunami that is not amplified substantially by resonance may still be dangerous in all parts of the coast (e.g., a run-up of 1 m is considered dangerous, especially coinciding around high tide).

8. Glossary

Atmospheric pressure-wave tsunami—generated by the explosive eruption of a volcano (e.g., Krakatau eruption in Indonesia in 1883). This type of wave is generated in the sea in response to a pressure wave in the atmosphere, which reportedly can pass many times around the earth. It can imitate a tsunami at large distances not directly affected by the conventional tsunami that may have been generated at the source.

Caldera—a large, circular depression in a volcanic terrain, typically originating in collapse, explosion, or erosion.

Coastal-trapped waves—The nearshore part of a reflected tsunami wave front that can become trapped at the coast and move up- or down-coast parallel to the shore to other localities.

Diatoms—microscopic single-celled plant that has a siliceous framework and grows in oceans and lakes.

Distant tsunami source—a distant (remote) source/national impact event generated at a site, such that a resulting tsunami takes longer than 3 hours travel time to reach the Bay of Plenty/eastern Coromandel coast, and likely to affect several regions. For example, a tsunami generated at a South American location will take at least 12 hours to reach New Zealand (i.e., providing an opportunity for longer warning times).

Fault—a fracture along which there has been significant displacement of the two sides relative to each other, parallel to the fault.

Local source tsunami—a local source/local impact event generated at a site, such that a resulting tsunami takes within say 30 to 60 minutes travel time to reach the coast, and only affects several 10's of km of coast. Consequently, there will be minimal warning time. However, by association with "natural" warnings such as extreme ground shaking or volcanic activity through education programmes, some people may have time to quickly move to safety.

Paleo-tsunami—probable events occurring prior to the historical record, that are determined by analysing depositional and erosional signatures in the coastal landscape. This work is based primarily on the collection and analysis of tsunami deposits found in coastal areas (e.g., through sediment cores), and other evidence related to the uplift or subsidence associated with nearby earthquakes. Such work may provide a significant amount of new information about past tsunamis to aid in the assessment of the tsunami hazard for any region.

Pyroclastic flow—a ground-hugging avalanche of very hot ash, air, pumice, rock fragments, and volcanic gases that rushes down the side of a volcano as fast as 150 km/hour or more, and can have temperatures greater than 500°C. In a coastal setting, such flows can cause a tsunami.

Pumice—a form of volcanic glass, usually of silicic composition, so filled with cavities that it resembles a sponge and is very light.

Regional tsunami source— regional source/regional impact event generated at a site, such that a resulting tsunami takes within 3 hours travel time to reach the coast, and is likely to affect most of the Bay of Plenty and eastern Coromandel region. Consequently, warning timeframe is still limited. However, by association with "natural" warnings such as extreme ground shaking or volcanic activity through education programmes, people may have time to move to safety.

Resonance—the natural wave period of a bay, bight, estuary, or harbour that can excite and amplify similar wave periods that may be present in an incoming tsunami wave-train. Each coastal area may have several resonant wave periods (called modes or harmonics).

Run-up height—the vertical distance from the pre-event tide level to the maximum elevation that the tsunami wave attains, regardless of how far inland.

Silicic—a type of magma in the Earth's crust. Magma can vary in chemical composition from basalt to rhyolite (silicic). There are important differences in the viscosity of the various magma compositions that have a strong influence on how the magma is extruded from the crust. The combination of high viscosity and lower temperatures (800–1000°C.) of silicic magma pre-disposes this magma type to pyroclastic eruptions (with higher potential for tsunami generation) and only rarely in lava flows.

Subduction—in plate tectonics, the process whereby one plate of the Earth's crust descends beneath another plate (underthrust).

Submarine landslide—a slide of sea-bed sediments down a continental slope that occurs under the sea, usually triggered by a seismic event.

Tephra—pyroclastic materials expelled into the air from an erupting volcano before cooling. Material ranges in size from fine dust to massive blocks.

Tsunamigenic—a geological disturbance or dislocation that has the potential to generate a tsunami.

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10. Appendix 1: Recollections of the May 1960 Chilean tsunami in Mercury Bay by Howard Pascoe (Whitianga resident).

RECOLLECTIONS OF A TIDAL WAVE by HOWARD PASCOE

The disastrous earthquakes in Chile which occurred in 1960, and the resulting sea waves spreading havoc and destruction round the Pacific, have focused the attention of the whole world on one of nature's most terrifying weapons - the tsunami. Commonly called a tidal wave, the tsunami gets its name from the Japanese. Literally, it means a "harbour wave" which appears to be not unreasonable when it is realised that these waves affect harbours and bays much more than they do ships at sea.

EXTRACT FROM NEW ZEALAND NAUTICAL ALMANAC

Whitianga east Coromandel coast

Mercury Bay is an East Coast bay shaped like a funnel about 5½ miles wide and 8½ miles deep closing down to the river mouth and is very sensitive to big easterly waves and oceanic disturbances. Such as seismic waves as were caused by the Chilian earthquake. On Monday 23rd May 1960 at 9 pm. the first of the tidal waves arrived and the water in the channel fell so people could walk outside the wharf at the north end, 20 mins later it had returned to cover the road by wharf. This it proceeded to do for the next four days. We were getting a tide every 40 mins. The fishing boats that were moored in the channel between the wharf and the hotel were boarded by their owners who walked out while they were high and dry and waited for the next outgoing rush of water and took them out to sea to safety. When I took the boat I was looking after the "Atlanta" 40ft long and powered by a 180 hp g.m. into the stream we were doing 12 knots and going backwards trying to get out of the river. This we did for 20 mins until the tide turned and swept us out through the entrance to safety. All this was going on in the early hours of Tuesday morning in the dark, a most terrifying experience.
One 28ft fishing boat the "Mermaid" broke her mooring and was swept up into the mangroves about 2 mile up the river. Another the 34ft "Marlin" tore the bollard out of the foredeck and was swept up on to the road on the wharf side of Karena Creek where she remained for over a week until she was pulled down and refloated.
On the Thursday May 26 at midday a message came through to the Whitianga Police that rocks on Norfolk Island were being swept off the cliffs 40ft up and so Sgt Mat.Andrews gave the orders to sound the siren for evacuation of the town.
Some folk going up on to the high ground on Davis point and others going to the high ground behind the aerodrome, where some remained all night and others returning home again about night fall. Just above Pah Point by the cable marker Robin Lee and I carved a mark in the sandstone cliff where the dust was washed off and a distinct line could be seen along the cliff some 4ft above the highest spring tide. (1.2m)
We estimated from that to where the tide went out was approximately 19ft, all this in 20mins. The rush of water when coming in was in excess of 12 knots, and would be half as much again going out which makes it something like 18 knots.
A big flat rock on Pah Point known as Schnapper Rock which people used to fish off and about high water mark and was about 20ft x 10ft & 6 ft deep, and weighing in excess of 30 tons was swept away and hasn't been located since. The tide was coming and going so fast it was leaving quite big fish stranded and flapping which could be heard at night and many were caught by residents in the daytime.
The sea came up Monk Street to where the power station was and up to the Dairy Coy front office steps also up to Mr. Merv.George's and Mr. Don Ross's houses also on to front lawns of houses and into the old aerodrome hanger along at the northern end of Buffalo Beach where it wet crates of corrugated iron stored on the floor, damaging them. Also the old wreck of H.M.S. Buffalo was left high & dry.
Mr. Alf Simpson hitched a bull dozer on to it and pulling pieces off, which are now in the local museum. I also watched a small trawler type of boat the "Three Kings" which was owned by Mr.Les Rydes and powered by 150 hp Hercules Diesel and capable of about 10 knots battling against a rush of water and a wave about 2'6" high and going backwards ~~sinking~~ & losing ground as she tried to get out to the open sea, which she did after about 20 mins. This was about 2 am.Tuesday morning which seemed to be the worst time. It went all day like that and on the Wednesday it seemed to be calming down which it did from then on, getting slower in its flow of water. It was Saturday before it had come back to anything like normal.

On the river that Tuesday morning were vessels:-

Caroline B. Dougherson	Norma W.M.S. Clark
Three Kings	Marlin
Ronomor	Oronga
Lady May	Tuna Eke
Atlanta	Ngaire an ex whale chaser & used for

game fishing was a very fast launch and even she could not head that lot off. There were quite a few people on the river that Tuesday morning in Dinghies and smaller boats trying to get them to safety.

I have mentioned the height the water rose each time the tide came in and another alarming thing was how far it went back each time it went out. This large movement of water travelling at such speed back and forth cut all the sandbanks and shell banks in the harbour right down to the black mud. It made quite a smell days later, the reason I think, was all the pipis and shell fish that had been disturbed and had died. Some of these shell banks were between 2 and 3 ft high and were on the flats on the western side of the harbour out from the Hotel from Kerena Creek South. The one on the edge of the channel out off the football field being over 400 yds long and about 3 ft high it was levelled right off and is still like that after 26 years. The others have reformed again in practically the same places but do not seem to be quite so high as they were before, but given time I think they will reform. The telephone cable was destroyed and dragged all over the place, some lengths of it finishing up the harbour. I can find bits of if laying in the mud even now. The 6"x6"x15ft post with a white triangle on top that marked where the cable crossed the harbour stood on one of these high shell banks near low water mark. This bank was covered in white shell and the tide covering it only on spring tides. Neap tides not covering it. This went in the waves too. Nothing being left.

On the 26th of May the Kaimarama School and Mercury Bay Schools closed because of the warning from Norfolk Island and the pupils were evacuated to the hills for two days.

I often think back about it all and thank God that this all took place in a week of very fine weather of practically no winds or swell outside in the bay and nobody was drowned.

W.R. Moyes
R. Lee
M.J. George
B. Dougherson
W.M.S. Clark
Don Ross
Trevor Stead
Eric Chaney
S. Rees
R. Blaney
Eric D. Wordy
Malcolm Nicol
Leon Munro
Ethel Hailey
T. Hailey

A.J. Munro
T.W. Lee
Amy Lee
R.W. Sewell
J.A. McLeod
A.S. McLeod
A.R.J. Bruce
E.W. Chaney
R.F. Bruce
P.A. Clark
K. Hodge
C.E. Ralph
H.W. Pearse

The above statement has been verified as a true record of the happenings of 23rd May 1960 in Whitianga River by the signatories who were residents at the time