2. Tsunami source details

The Northland area faces a range of potential tsunamigenic sources that include several local and distant fault systems, and underwater landslides. A NIWA study (Goff et al. 2006) identified South America as a potential distant source, Tonga-Kermadec Trench and Solomon Sea/New Hebrides Trench as regional sources, and several potential submarine landslide sites off shore of the far North coast. For the purpose of this study, we focus on South America as a distant source, and the Tonga-Kermadec Trench as a regional source. The rationale behind this is as follows: the plate boundary along the west coast of South America has a relatively high rate of convergence and is seismically very active. Large subduction earthquakes in this region have return periods of around 50-100 years. In New Zealand, prior to 2010, there were three moderately large tsunamis from South America: in 1868, 1877, and 1960. In South America however, there have been an additional 11 historical tsunamis propagated by M_w 8+ events between 1562 and 1859. Tsunamis originating from South America thus make up the most probable tsunami threat facing the Northland region and can be expected every 50-100 years.

Large subduction zone earthquakes in the Tonga Kermadec Trench (of which the Hikurangi Trench is the southern extension) occur far less frequently, with return periods of 500 to 2000 years (Goff et al. 2006). The impact of tsunamis generated by these events on the New Zealand coastline is far more extreme, possibly evidenced by the presence of likely paleo-tsunami deposits discovered 30-40 m above sea level at Henderson Bay and Great Barrier Island (Nichol 2003, 2004). Thus, these scenarios represent a likely worst-case scenario for the whole Northland region.

Solomon Sea and New Hebrides earthquakes are unlikely to produce significant tsunamis on the Northland coastline. For example, the April 1 2007 Solomon Island earthquake (M_w 8.1) produced a tsunami with a maximum amplitude in New Zealand of 50 cm at Charleston and considerably smaller in other places. Submarine landslides can cause large tsunamis; however, their effect is generally more localised than for tsunamis generated by large earthquakes. In what follows we characterise the source details for the most important events.

2.1. Distant, eastern source: South America (Chile/Peru): source and wave propagation description

The subduction zone running along the west coast of South America has frequent (approximately every 50 years), large ($>=M_w$ 8.5) earthquakes. The largest recorded earthquake (M_w9.5) occurred in Chile in 1960. The resulting tsunami and two other South American tsunamis (1868, 1877) inundated the Northland coast during the last 150 years (Chagué-Goff and Goff 2006). The only other significant tsunami reported in Northland was in 1883 from the Krakatoa eruption. Tsunamis from the north (e.g., Solomon Islands/Fiji Basin or Indonesia) have not caused significant impact on the Northland coastline in recorded history. Thus, South America represents our most probable source of far-field tsunamis.

Three primary factors determine how New Zealand is affected by tsunamis that travel across the Pacific Ocean: source location and geometry, wave transformations that occur when the tsunami crosses the ocean, and the effects of the bathymetry and geometry of the recipient continental shelf and coastal region.

The size of a tsunami generated by a subduction zone earthquake depends on several factors, including the magnitude, source geometry and location of the event. The magnitude and source geometry of the earthquake determine the surface deformation, which, in turn, determines the overall size and length scale of the tsunami. Representative source geometry can be calculated via an empirical formula that takes fault length, width, slip and moment magnitude into account. Plate boundaries and convergence rates can be found in Bird (2003), and Pacheco et al. (1993) gives information on the fault geometry. A typical fault could be 100 km wide along the dip and extend to 45 km below the surface. The location of the earthquake also determines the primary direction that the tsunami will propagate and, hence, whether it will hit or miss New Zealand. A good example of this is a comparison of the tsunamis caused by the 1868 and 1960 Chilean earthquakes. The 1868 earthquake was smaller in magnitude and caused a smaller tsunami in general, however, its location meant that its waves and energy were aligned to focus on the New Zealand coastline. Although the 1960 earthquake was larger, its source location directed the ensuing tsunami mostly to the north of New Zealand, so that less of its energy reached New Zealand. Thus, the net effects of tsunamis from these very different events on the north of New Zealand were similar.

Tsunamis generally follow large circlular routes across the ocean, but this path is modified by reflection and refraction where the water depth changes. Waves may also be diffracted as they pass through island chains, but there are no significant islands between South America and New Zealand. Usually, trans-Pacific tsunami propagation is modelled using nonlinear shallow water theory with dissipation of short wavelengths, or by linear shallow water theory where all waves travel at the same wave speed. This latter approach is not strictly correct because, over this great distance, the waves are dispersive (long waves travel faster than short waves) and generate a train of wave rather than a single wave. This behaviour is important when the run-up is considered because multiple waves can cause resonance effects.

For the purposes of this report, we have chosen to specify an incident tsunami at the eastern edge of the model mesh (210 degrees east longitude) that represents a "direct hit" scenario similar to the 1868 South American event. We have used the concept of inverse modelling to ground truth this scenario using the more reliable historical data for the 1868 event and to a lesser degree the 1960 event. In effect, the model fills in the missing information along the coast using the historical data as a rough guide.

2.2. Regional/Local – Eastern source: Tonga-Kermadec Trench

Subduction zone earthquakes occur along the Tonga-Kermadec Trench associated with the Pacific/Australian plate boundary. The Hikurangi Trough (HT in Figure 1) is a southern extension of the Tonga-Kermadec Trench. The TKSZ is located beneath the eastern margin of the continental shelf and the Kermadec Ridge from the east coast of the North Island to Tonga, where the Pacific Plate under-thrusts (subducts) to the west. Historic earthquakes of magnitude M_w 8.0 to 8.3 have occurred along the Kermadec boundary (ITDB/PAC 2004) in the early 1900's.

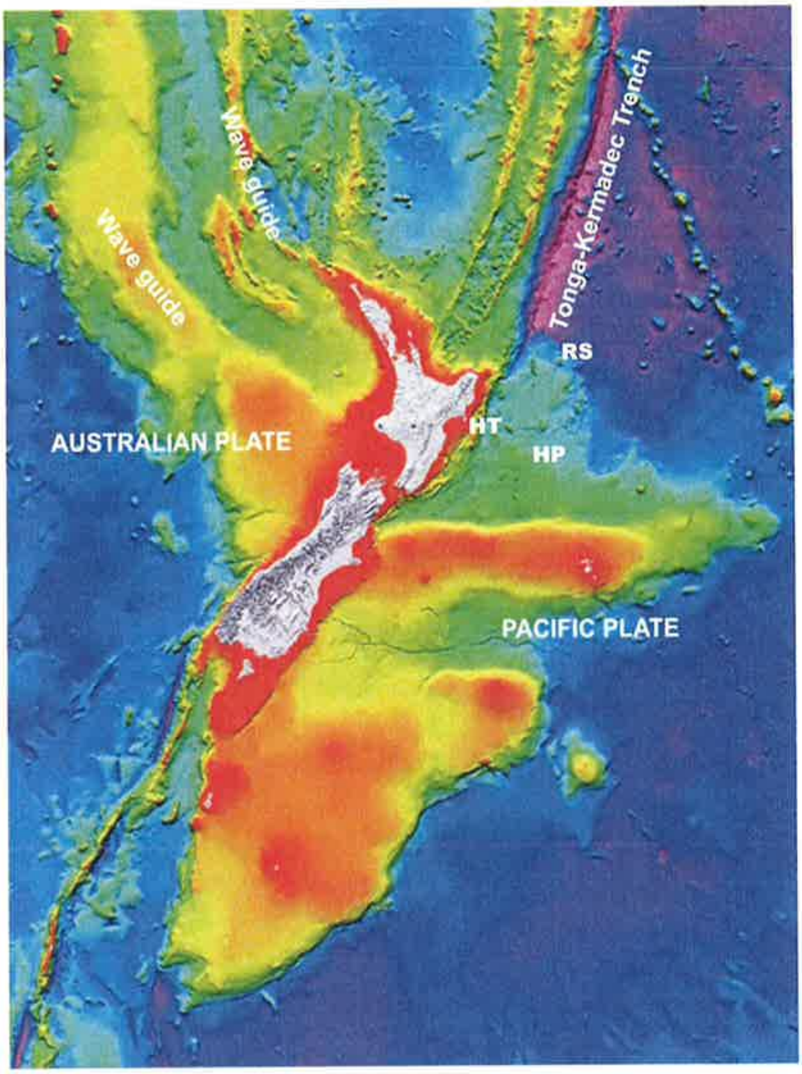

Figure 1: Geophysical setting of New Zealand showing the Tonga-Kermadec Trench (taken from Goff et al. 2006).

Palaeotsunami data have been used successfully to ground-truth models of local tsunami inundation from this source for some parts of New Zealand (Walters et al., 2006a). There are considerable data available incorporating records accumulated from a series of studies (e.g., Nichol et al., 2003; 2004; Bell et al., 2004; Goff et al., 2005; Chagué-Goff and Goff, 2006). Undoubtedly, there is a large impact from this source on the eastern and north-eastern facing coasts of the North Island.

Although the Tonga-Kermadec-Hikurangi trench extending north along the east coast of the North Island to Tonga is a prominent feature in the seabed topography, there is a paucity of definitive geophysical information for the area north of East Cape. South of East Cape along the Hikurangi subduction zone, Reyners (1998) suggests that the subduction zone area decreases between Wairarapa and East Cape and events of about M_w 6.9 are estimated for the northern part of this segment.

Up to a distance of approximately 250 km north of East Cape, the Hikurangi Plateau (HP in Figure 1) is being subducted beneath the Kermadec margin. At the northern edge of the plateau is the Rapuhia Scarp (RS in Figure 1) where the Hikurangi Plateau volcanics change abruptly to oceanic crust (Collot and Davey, 1998; Davey and Collot, 2000). The Rapuhia Scarp is approximately 1 km high and results in an increase in depth of approximately 1.5 km to the north. This implies that the shelf slope is over-steepened and there is a possibility of large submarine landslides and slumps. In addition, the scarp may act as a termination point for subduction zone ruptures to the north and south. Moreover, the change in fault dip across the scarp suggests that surface deformation caused by fault rupture may vary between sections to the north and south of the scarp.

Since the magnitude and location of subduction zone events is not well defined for this area, Goff et al. (2006) explored a range of events and compared the results with elevations of palaeotsunami deposits primarily from the 15th century (e.g., Walters et al., 2006a). The first set of events were three M_w 8.5 events placed south and north of the Rapuhia Scarp, and the next northern section of the fault respectively. The southern section mainly impacts the Bay of Plenty to Great Barrier Island, the next northern section mainly impacts Northland, and the far north section has minimal impact because the tsunami is directed north of the North Island. Events farther north have even less effect. An M_w 8.0 to 8.7 would be considered a reasonable magnitude event for the TKSZ. However, the 2004 Boxing Day tsunami in Indonesia serves notice that these are probably under-estimates for the maximum magnitude event that could occur. Hence, Goff et al. (2006) also considered an M_w 9.0 event, which is twice as long and has twice the slip of the M_w 8.5 event. This fault extends across the Rapuhia Scarp.

From these earlier results, we have chosen the two events that have the most impact on Northland. The first of these is a dislocation placed north of the Rapuhia Scarp (M_w

8.5). This was the M_w 8.5 event that had the greatest overall impact on the Northland coastline. A rupture south of the Rapuhia Scarp also generates a significant tsunami that hits Northland and has higher wave heights at the coast in the Whangarei/ Bream Bay area; however, the differences in this area are not large and the tsunami we chose has higher wave heights for the remaining coast. The second tsunami-generating earthquake has a longer dislocation extending north from East Cape (M_w 9.0). This generates a probable worst-case scenario. Pacheco et al's (1993) data provides an estimate of dislocation width (87.7 km) and depth (40 km). The dislocation length is 250 km for the smaller event, and 500 km for the larger event. For a dislocation of 10 m at the fault surface, the resultant moment magnitudes are 8.5 and 9.0. Using this fault geometry, we used Okada's (1985) dislocation model to calculate seabed displacement.

The best estimates of return periods for large local/regional tsunamis come from palaeotsunami records. This evidence is limited and only the largest events (with run-up greater than 5 m) leave records of their passing. Chagué-Goff and Goff (2006) give evidence for a large palaeotsunami, which occurred approximately 600 years ago. More recent evidence from the wetlands at Mimiwhangata (Pearce 2006) shows three such large events in the last 6000 years, giving a return period of around 2000 years for the very biggest tsunamis (which we model here as a M_w 9.0 subduction zone event in the Tonga-Kermadec Trench).

Events such as the Tonga-Kermadec M_w 8.5 are expected to occur more frequently than the larger events with a return period on the order of 500 years. This is not to say that the specific event that we modelled will occur every 500 years, but an event which may cause comparable damage is likely to occur on average with this frequency. In other words, if it were possible to collect a long enough record of observations, say over a period of tens of thousands of years, then one would expect that the *average* of the time intervals that separated events causing this level of damage (or worse) would be approximately 500 years. This probably represents an upper estimate of the return period for these events, since they may have occurred with greater frequency, but geologic evidence has so far escaped detection.