EQC Project 03/490 - Understanding local source tsunami: 1820s Southland tsunami

Client Report
2005/153

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EXECUTIVE SUMMARY

Southland lies adjacent to an offshore section of the plate boundary between the Australian and Pacific tectonic plates. The plate boundary zone consists of numerous active faults that could potentially rupture the seafloor in large earthquakes and trigger tsunami with devastating consequences for the Fiordland and Southland coasts. However little is known of the frequency and size of earthquakes likely to be generated in this southern part of New Zealand's plate boundary, and even less is known of the potential for tsunami to be triggered in this area. The historical record provides some indication that such events are a real threat for Southland and Fiordland. Historical reports of a tsunami killing a group of Māori on the beach near Orepuki in the 1820s and large earthquakes and a possible tsunami occurring in Fiordland in 1826 are investigated in this report.

Further assessment of historical records and collection of additional reports suggests the 1820s Southland tsunami and the 1826 Fiordland earthquake are separate events so the exact date of the 1820s Southland tsunami remains unknown. The source of this tsunami also remains unknown but modelling indicates it is more likely to have been a Puysegur subduction zone earthquake than a Fiordland subduction zone or Alpine Fault earthquake. The effects of the 1826 earthquake are considered consistent with a large Fiordland subduction zone event that caused uplift of the coast near Doubtful Sound, extensive landsliding and possibly a local tsunami.

Physical evidence of these events was searched for at three sites on the Fiordland coast. At Cascada Bay near the mouth of Doubtful Sound, a series of raised bedrock platforms and recent revegetation of an 8 m high surface could be the result of tectonic uplift occurring in previous large earthquakes and vegetation stripping in a past tsunami. However further investigation of these features and confirmation of their age would be required before attributing them to the known historical events. Evidence of older earthquakes and / or tsunami appears to be preserved in sedimentary sequences at Martins Bay in northern Fiordland and at Goose Cove in southern Fiordland. Therefore, although the steep topography, high rainfall and strong wind and wave regime of Fiordland make it a challenging place for any evidence of past earthquakes and tsunami to be preserved, there is potential onshore to derive further information about behaviour of the offshore faults.

Three sections of the plate boundary considered most likely to be tsunamigenic were modelled: the offshore section of the Alpine Fault, the Fiordland subduction zone and the Puysegur subduction zone. Models were developed for the predicted deformation of the sea floor in a large earthquake on each of these structures (elastic dislocation models). Tsunami propagation models were used to predict the behaviour of the resultant tsunami from generation to impact. The offshore section of the Alpine Fault is considered capable of rupturing in a magnitude 7.8 earthquake, which would trigger a tsunami with water elevations at the shore of at least 4 m and maximum impact between Milford Sound and West Cape. The Fiordland coast also endures the greatest impact (water elevations[1] at the shore of 4 m) from a tsunami triggered by a Fiordland subuction zone earthquake of magnitude 7.7-7.9. The Southland coast is most vulnerable to earthquakes in the Puysegur subduction zone where an earthquake of magnitude 8.5-8.6 is considered possible and would trigger a tsunami with water elevations at the shore of 4 m at numerous points along the Southland coast.

1. [1] WATER ELEVATION at the shoreline is used here because tsunami run-up is only approximated in this application of the tsunami model. TSUNAMI RUN-UP (m), a measure much used in tsunami-hazard assessment, is the vertical height the waves reach above the instantaneous sea level at the time of impact at the farthest inland limit of inundation. This measure has a drawback in that its relationship with the water elevation at the shore depends markedly on the characteristics of waves and on the local slopes, vegetation, and buildings on the beach and foreshore areas, so it is highly site-specific. While many models can properly treat this variability in overland conditions (such as the model used in this study), the necessary input data describing the topography and vegetation is usually nonexistent. While using a "vertical wall" at the shoreline edge of the model grid represents steep cliffs and dunes rather well, it is only a crude approximation for run-up over a more general low-lying topography.

1.0 INTRODUCTION

Māori history as recorded and interpreted by 20th century European historians refers to an apparent tsunami in Southland in the 1820s, in which a large group of Māori were killed while walking along the coast near Orepuki. In unrelated documents, sealers in Dusky Sound, Fiordland, record a large earthquake in 1826 (with aftershocks extending into 1827) that caused a probable tsunami, many landslides in the mountains, and possibly coastal deformation. The purpose of this project was to investigate whether this earthquake could be the source of the Southland tsunami, and more generally, whether large Fiordland or Puysegur Trench earthquakes pose a significant tsunami threat to Southland, as well as to Fiordland and whether there is evidence of past tsunami having occurred along the Fiordland coast.

We have addressed these questions by:

- Collating known historical evidence of both events.
- Using expert knowledge of the geology and near-shore topography of the Fiordland coast to identify two sites (Martins Bay and Goose Cove) judged suitable for preserving evidence of past tsunami inundation.
- Identifying a site referred to in historical records (Cascada Bay) that may preserve evidence of uplift associated with a large earthquake, and might also preserve evidence of past tsunami.
- Visiting these three sites and examining them for evidence of earthquakes and tsunami in the form of "anomalous" sedimentary deposits and changes to coastal landforms and vegetation. Radiocarbon dating of appropriate organic material has been used to estimate the age of such events.
- Selecting tsunamigenic earthquake sources in Fiordland using existing seismological and offshore fault information as well as information on potential earthquake uplift obtained during the site investigations at Cascada Bay, Doubtful Sound.
- Using numerical modelling of the earthquake sources and the resultant tsunami to investigate the potential of credible large Fiordland and Puysegur Trench earthquake sources to cause moderate to large tsunami in coastal Southland and Fiordland.

This report documents the research that was undertaken over the last two years towards these objectives, and summarises the results that have been found.

Figure 1 Location map showing important sites mentioned in the text. 1:250,000 topographic map courtesy of Land Information New Zealand (Crown Copyright Reserved).

2.0 TECTONIC SETTING OF THE FIORDLAND REGION

2.1 The Plate Boundary

The active boundary between the Pacific and Australian tectonic plates traverses the length of New Zealand with a NE-SW orientation and in the south-western South Island it exists offshore of the Fiordland coast (Fig. 2A). The plate boundary is complex in this region, accommodating oblique convergence of c. 35 mm/yr, partitioned onto a variety of structures (Lamarche and Lebrun, 2000). Here we describe three major components of the plate boundary that are likely to be tsunamigenic (Fig. 2B): the Alpine Fault, the Fiordland subduction zone and the Puysegur subduction zone. We acknowledge that there are many other onshore and offshore faults about which little is known but which may contribute significantly to the tsunami and earthquake hazard of the region.

The Alpine Fault is an 850 km long, predominantly dextral strike-slip structure with the southernmost 230 km existing offshore west of the Fiordland coast (Barnes et al., 2002; 2005). In the central South Island the fault accommodates at least 70-75 % of the Pacific–Australian relative plate motion (e.g., Norris and Cooper, 2000), and a component of reverse slip is responsible for the growth of the Southern Alps. In Fiordland the Alpine Fault becomes near vertical and takes on almost pure strike-slip motion (e.g., Berryman et al., 1992; Sutherland and Norris, 1995). The fault heads offshore towards the southwest near the mouth of Milford Sound where it accommodates 26 ± 10 mm/yr of dextral strike-slip motion (Barnes et al., 2005); similar to rates observed at the southern end of the onshore portion of the fault (26 ± 7 mm/yr, Hull and Berryman, 1986; Sutherland and Norris, 1995). The predominantly strike-slip nature of the fault in the offshore region is not believed to be conducive to tsunami generation. However identification of the complex segmentation of the fault offshore and existence of associated pull-apart and releasing-bend basins (Fig 2B) (Barnes et al., 2001; 2005) suggests that fault rupture would lead to substantial changes in bathymetry potentially capable of triggering a tsunami (refer to section 5.0).

The Fiordland subduction zone accommodates part of the Pacific–Australian plate convergence offshore west of Fiordland where oceanic Australian plate is subducted beneath continental Pacific plate (Eberhart-Phillips and Reyners, 2001; Barnes et al., 2002). While several conceptual models for the three-dimensional configuration of this part of the plate boundary have been suggested (eg, Lebrun et al., 2000; Reyners and Webb, 2002), it is arguably the least well-understood part of the New Zealand plate boundary. Barnes et al. (2002) describe a 240 by 25 km zone of thrust faulting, west of the offshore strand of the Alpine Fault, which is estimated to accommodate 1-5 mm/yr of shortening perpendicular to the margin. It is unknown what length of subduction thrust is likely to rupture in a single earthquake, but this 240 km long zone may be close to providing a maximum likely scenario for tsunami generation.

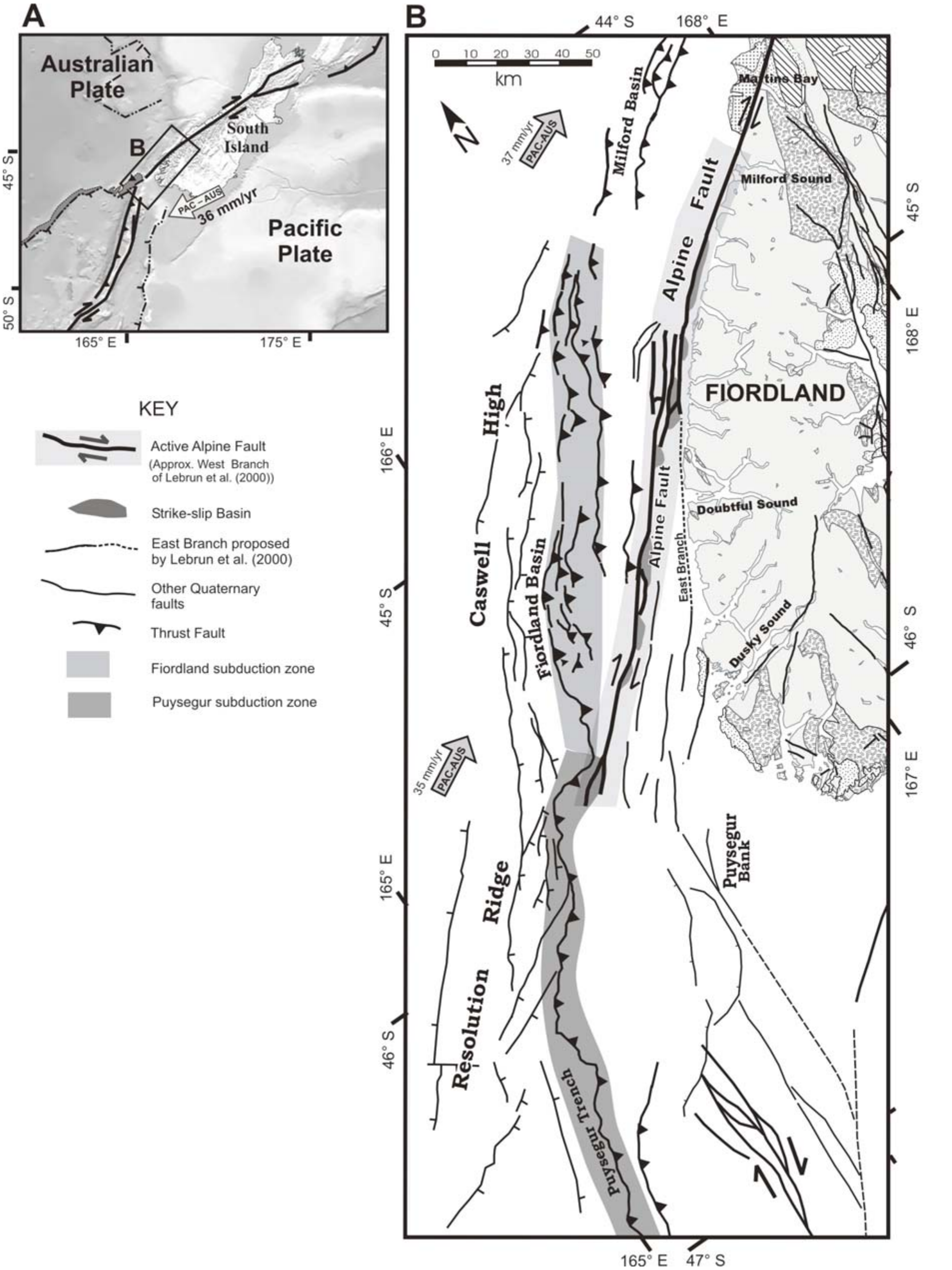


Figure 2 A: Tectonic setting of the South Island on the boundary between the Pacific and Australian tectonic plates. Macquarie Ridge (not labelled) lies west of the plate boundary south of 50ºS. B: Simplified tectonic setting of Fiordland illustrating three main components of the plate boundary in this region (modified from Barnes et al., 2005).

The Puysegur subduction zone extends c. 150 km southwest from the southern end of the west branch of the Alpine Fault beneath the ocean southwest of Fiordland (Lamarche and Lebrun, 2000). The Puysegur Trench marks the position of the subduction thrust at the surface of the seafloor – a clearly defined feature compared with that further north at the Fiordland subduction zone. Again, rupture of the whole length of this subduction zone is taken as a maximum scenario for tsunami generation.

2.2 Previous Earthquakes

2.2.1 Modern (post-1960) Earthquakes

Since initiation of this project, three of New Zealand's largest recorded earthquakes have occurred in the study area offshore of Fiordland and Southland. In December 2004 an M_w 8.0 earthquake occurred near Macquarie Ridge (Fig. 2; south of Puysegur Trench) and in November 2004 an M_w 7.0 earthquake occurred at the Puysegur Trench. Data from these events and their aftershock sequences are currently being analysed. The third earthquake, M_w 7.2 in August 2003, is the largest earthquake ever recorded instrumentally on the Fiordland subduction zone (Reyners et al., 2003). It occurred on a shallow part of the interface between the subducting Australian Plate and the overlying Pacific Plate beneath the coast near Secretary Island. The GPS-derived fault model predicted an area of uplift of 45 cm immediately offshore of Secretary Island (Reyners et al., 2003) and the model is supported by observations derived from intertidal organisms of at least 20 cm of uplift at Open Cove inside Thompson Sound (Cochran, unpublished data). The earthquake caused enough deformation of the seafloor to trigger a small tsunami (see tsunami section below).

A number of modern earthquakes have occurred in all three of these general areas – Macquarie Ridge, Puysegur Trench and the Fiordland subduction zone. In particular, there has been a series of earthquakes in the Doubtful Sound area over the last two decades: the M_w 6.7 Te Anau earthquake in 1988; M_w 6.4 Doubtful Sound in 1989; M_w 6.8 Secretary Island in 1993; and M_w 6.1 Thompson Sound in 2000. Of these events only the 1989 and 1993 events were subduction interface earthquakes, both with rupture planes overlapping that of the 2003 earthquake. The other earthquakes represent complex deformation within either the subducted or overlying plates (Reyners and Webb, 2002; Robinson et al., 2003).

2.2.2 Historical Earthquakes

The record of large (>M6.5) pre-1960 earthquakes in Fiordland is without doubt incomplete and the locations and magnitudes of known earthquakes are poorly determined. Of the six events in the catalogue (National Earthquake Information Database, GNS), one is the 1826 earthquake that initiated this study; another is a doubtful event in 1817, with the other four being post-1900 (1938, 1939, and two in 1943). Investigation of likely source parameters for these post-1900 events indicates they are related to deformation in the Fiordland subduction zone (Doser et al., 1999). Known historical accounts suggest that other large Fiordland events probably occurred between 1855 and 1900, but this part of the catalogue of earthquakes has not been comprehensively investigated and no magnitudes, locations or depths have been assigned to the events. It is highly likely that they cannot be reliably assigned. Therefore this era provides little insight into past seismic activity, or the frequency and magnitude of large Fiordland earthquakes.

Historical earthquakes in the Puysegur region (1918 and 1945) appear to be related to strike-slip movement near the Puysegur Bank rather than subduction thrust motion at Puysegur Trench (Doser et al., 1999). However locations and solutions for these events are also uncertain.

2.2.3 Pre-historical Earthquakes

Paleoseismological studies aimed at retrieving pre-historic records of subduction zone earthquakes generally require greater research effort than those for onshore faults and have only recently begun for the Hikurangi subduction zone offshore of the North Island of New Zealand. Such work for the Fiordland and Puysegur subduction zones is yet to be initiated. However identification and dating of past earthquakes on the onshore portion of the Alpine Fault has received attention for several decades (eg, Adams, 1980; Cooper and Norris, 1990; Bull, 1996) and ages of the last three earthquakes on the central Alpine Fault are now reasonably well defined at c.1460 AD, 1610-1620 AD and 1717 AD (Wells et al., 1999). Several studies underway will extend the record of Alpine Fault earthquakes back in time enabling better assessment of the fault's behaviour.

Identification of episodes of forest disturbance related to Alpine Fault earthquakes in Westland has provided an additional technique with which to assess the age and extent of pre-historic fault ruptures (eg, Wells et al., 1999). In the course of that work, forest disturbance coinciding with the age of the Fiordland earthquake at 1826 AD has been identified (Cullen et al., 2003). This disturbance is unlikely to relate to an Alpine Fault earthquake as no such earthquake has been reported in the historical record and the distribution of disturbance differs from those of previous Alpine Fault earthquakes. The episode of synchronous tree establishment (occurring after the inferred disturbance) is dated at 1800-1850 and extends between the Ohinemaka and Waiho floodplains in southern Westland (Cullen et al., 2003). An historical account of the 1826 AD Fiordland earthquake suggests extensive landslide damage as far north as Cascade Point (refer to section 3). However, if the forest disturbance is also a result of this earthquake, the inferred zone of strong shaking would need to be extended by over 100 km to the north.

2.3 Previous Tsunami

The remoteness of Fiordland means that its historical record of tsunami (Downes, unpublished data) is deficient. The small tsunami resulting from the August 2003 M_w 7.2 Fiordland and M_w 8.0 December 2004 Macquarie Ridge earthquakes were the first earthquake-generated events recognised for this area since the 1826 tsunami described in section 3.0. Rupture in the August 2003 earthquake is thought to have caused 45 cm of uplift over a wide area of seafloor thereby triggering a small tsunami – the most proximal measurement of which was at Jacksons Bay 180 km NE of the epicentre, where the peak-to-trough amplitude was c. 30 cm (Power et al., 2005). The tsunami was also recorded at several locations in Australia with amplitudes about half those at Jackson Bay. Considering the distance travelled to Australia and the consequent dissipation of energy along the way, this response is quite strong although still small. This is in part due to local resonance effects, but also to the strong directivity that characterises tsunami from large earthquakes. Specifically, tsunami wave heights are greatest perpendicular to the strike of the fault. Hence the tsunami was strongest towards the Tasman Sea and Australia, and weak along the strike of fault (up the West Coast).

Post-earthquake landslide reconnaissance also revealed that a landslide into Charles Sound caused a 4-5 m tsunami that was localised to within a few hundred metres of coastline and damaged a nearby jetty. A landslide into Deep Cove in May 1987, which did not occur in association with an earthquake, also caused very localised 3 m waves.

There is some indication that a tsunami occurred in association with the 1826 Fiordland earthquake. Written reports refer to “the flux and reflux of the sea” being violent but it is hard to differentiate between seiching and tsunami from these reports alone (refer to section 3.0). At Okarito Lagoon, over 300 km NE along the coast from Doubtful Sound, physical evidence for tsunami inundation of the lagoon and surrounding shoreline at about this time has been documented (Goff et al., 2004). If this tsunami damage was the result of the 1826 Fiordland earthquake, a much greater offset of the seafloor (and greater earthquake magnitude) would be required for that earthquake than for the August 2003 Fiordland earthquake in which tsunami wave height at Okarito Lagoon would have been no more than 30 cm. There has not been any paleotsunami research previously undertaken in Fiordland but research is currently underway in surrounding regions such as Southland and Westland. Previous tsunami known to have inundated Southland are outlined in section 3.0.

3.0 HISTORICAL RECORD

3.1 1820s Southland tsunami

The account of a possible tsunami in Southland which initiated this study appeared in a regular column by Millie Saunders in the Southland Daily News in about 1936:

> According to Māori tradition, 1820 or thereabouts, was a year of great disturbances, earthquakes in the north and tidal waves in the south … the Maories of Murihiku, both on the west and south-east coasts affirm that a great tidal wave swept over the area where Orepuki now is, and all along the western coast, and thereby many hundreds of the Ngati Mamoe tribe met their deaths by drowning. It appears that it was the usual custom for tribes to meet together at the Waiau mouth in the autumn to provide themselves with enough fish to carry them over the long winter…. They were travelling along the beach track when great tidal waves rose from the ocean, and carried them out to sea, and it is believed no survivors remained to tell the story; a similar adventure is also recorded from an area beyond Fortrose. (Saunders, M ca. 1936).

In the course of this project, several other accounts of what seem to be the same event have been found (see Appendix 1). The dates differ, but are within the first three decades of the 19th century. The source of the story, however, has not been identified, and as indicated in the account cited above, it is probable that it is a Māori oral tradition recorded and interpreted by Europeans. The accuracy of any of the accounts is unknown. Written records of the sealers and whalers have been extensively investigated by historians, and as far as is known no accounts of this event have been found among them. However, it was beyond the scope of the study to investigate all potential archival and published material.

The account given above is the most informative of any of the accounts, particularly in its reference to earthquakes in the north, the time of year (autumn), the number of Māori drowned, and to the extent of the event. The descriptions are consistent with a tsunami rather than a storm. Further, the stories are quite different from a relatively well recorded event (by Europeans at the time) in June 1823, in which 40 or so Māori, including the chief, were drowned when their canoe foundered on a food-gathering trip to Ruapuke Island, in Foveaux Strait (See Appendix 1).

The tsunami seems to have affected the coast from the west to the south-east assuming that the event at Fortrose occurred at the same time (it is not clear in the account). Because of the cliffs at Orepuki (Fig. 3), which are noted in historical accounts to have retreated in the last 150 years (Smith, 2003), only a moderate tsunami, perhaps 2-4 m high, would be needed to cause the deaths of all walking along the beach at the base of the cliffs.

Figure 3 View of the Orepuki coastline looking south from Gemstone Beach.

As far as the reference to earthquakes in the "north", few pre-European settlement (in 1840) earthquakes are known and well-identified, the exception being the 1826 Fiordland earthquake described in the next section. There is no suggestion in any account that the Māori felt an earthquake immediately prior to the tsunami. Nevertheless, Fiordland and Puysegur earthquakes are obvious choices as local tsunami sources that might be consistent with the extent of the tsunami, and several scenarios are investigated later in this report. A further source, not investigated here, could be a large South American event. However, the tsunami would need to be larger than those experienced in Southland in 1868, 1877 and 1960 (Downes, unpublished database), and larger than the South American events known to have occurred in the period 1800-1840 (ITDB/PAC 2004).

3.1.1 Timing

Accounts of a possible tsunami relating to a much earlier period (possibly around 1400, in the early period of Māori settlement in Southland) also have been found (McFadgen pers. comm.), and are given in Appendix 1. While several of the stories are clearly mythical and difficult to interpret, and others could relate to flooding in storms, one has strong similarities to the 1800s tsunami story, casting doubt on the reliability of the 1800s date. It is possible that 20th century oral historians have associated the story with the wrong period. Alternatively, an additional event could have occurred in the 14-15th centuries. Despite the uncertainty, we will continue to refer to the tsunami as occurring in the 1820s until such time as a more definitive date or era is determined.

3.2 1826 Fiordland earthquakes and tsunami

The accounts of the 1826 Fiordland earthquakes come from the publication and notes of one person only – Rev. Richard Taylor, naturalist, amateur geologist and missionary who in 1854-5 resided in Wanganui. His publication, *Te Ika a Maui* (1855), records a number of earthquakes and their effects, prominent among them, the 1855 Wairarapa earthquake. He collected his information from a number of sources, including newspapers and personal contacts. In the case of the 1826 earthquake, the information seems to have come from Edward Meurant, who started as a sealer on the Fiordland coast and later became a Māori interpreter.

At the initiation of this study, the only account known was that in *Te Ika a Maui* (Taylor, 1855):

> It appears that from 1826 to 1827 there was an almost constant succession of earthquakes, some of which were sufficiently violent to throw men down. At times he [a person formerly engaged in sealing in Dusky Bay] and his party, who then resided on a small island, were so alarmed lest it should be submerged, that they put out to sea; there, however, they found no safety, for such was the flux and reflux of the ocean, that they were in the greatest danger of being swamped, and were thankful to get on shore again. The sealers were accustomed to visit a small cove called the jail, which was a most suitable place for anchorage, being well sheltered with lofty cliffs on every side; and having deep water in it close to the shore , so that they could step out on to the rocks from their boats. It is situated about 80 miles to the north of Dusky Bay. After the earthquakes the locality was so completely altered; the sea had so entirely retired from the cove, that it was dry land. Beyond Cascade Point the whole coast presented a most shattered appearance, so much so that its former state could scarcely be recognized. Large masses of the mountains had fallen, and in many places the trees might be seen, under water. (Taylor, 1855; p 235)

In the last two years, Taylor's notes in preparation for his book have been found (Rodgers, pers. comm.) and these provide critical detail, on when the largest earthquakes occurred, and the location of The Jail:

> In 1826 the greatest shocks were felt, they were sufficiently severe as to throw a man down. They continued five or six months to the beginning of 1827, seldom 2 days elapsed without one occurring. The narrator who was engaged in sealing there in Dusky Bay stated that some times they were so alarmed lest the island sh^{d} [should] sink that they used to push off to sea & there the flux and reflux of the sea was so violent that they were afraid of being swamped at sea & had to return again. [The narrator also stated] That there was a small cove which was shut in on every side with high land and thence named the Jail in which they were accustomed to anchor their sealboat as there was deep water in it and they could jump out of their boat on [to] the rocks. After these shocks, the sea completely retired from the cove so that it ceased to be any longer a harbour, as the boat could only come to its mouth. This [information] was furnished by Mr Meurant the native interpreter who has been many y[ear]rs in the middle island Oct 5 1847.

The jail cove is in Grono's Harbour about 80 miles N of Dusky Bay. Beyond Cascade Point the whole coast presented a most shattered appearance. After the shocks large masses of the mountains fell, so that in many places trees might be seen under the water. [Here is inserted a sketch of the location of The Jail, shown in Figure 4]. (Taylor, ca. 1854). [The full account without edits and with explanatory notes is given in Appendix 2.]

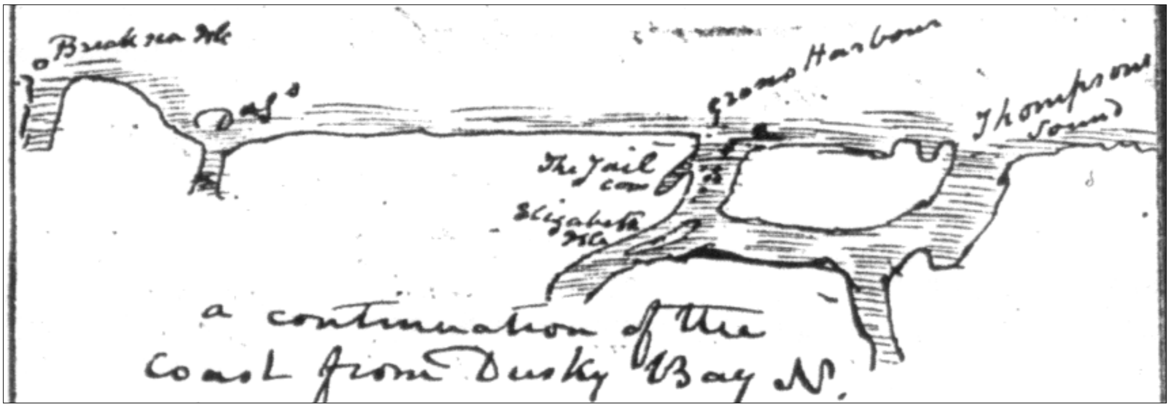


Figure 4 Sketch map from Taylor (ca. 1854) showing the location of "The Jail" within Grono's Harbour, which is now known as Doubtful Sound.

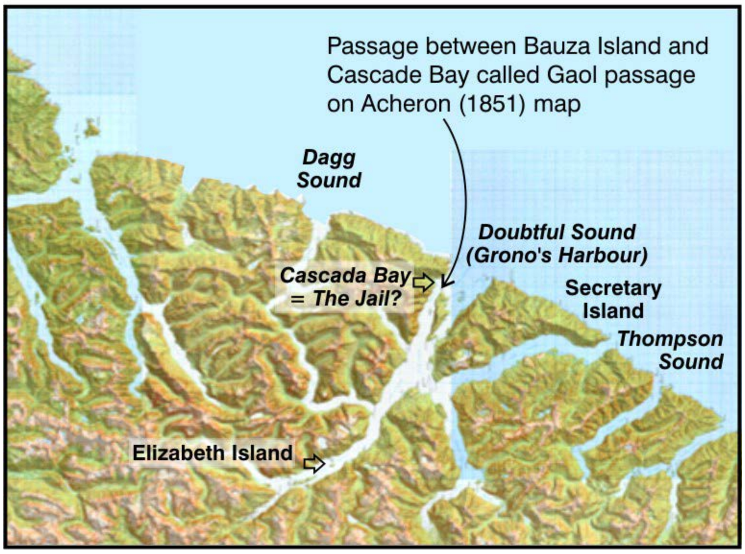


Figure 5 Map of the Doubtful Sound area of Fiordland for comparison with Figure 4 showing the sketch map from Taylor (ca. 1854).

Both documents clearly describe a sequence of large earthquakes with strong shaking (≥MM7), possible uplift, widespread occurrence of landslides in the mountains and along the Fiordland coast (≥MM8), and tsunami or seiches in Dusky Sound.

3.2.1 Uplift

The changes as described at The Jail cove suggest that significant uplift may have occurred. Figures 4 and 5 indicate that this cove is near the entrance to Doubtful Sound, originally called Grono's Harbour after a sealing ship's captain of that name (Hall-Jones 1984). Cascada Bay most aptly fits the description as a small cove, and its location is consistent with the name of the passage between Cascada Bay and Bauza Island, now named Patea Passage but previously named on 1″/mile topographic maps and on the 1851 Acheron map as Gaol Passage (the Acheron map is reproduced in Hall-Jones (1984)). The reason for this name being given on the 1851 Acheron map has not previously been recognised (see Hall-Jones' (1984) comment on the name).

Fortuitously, photographs of Cascada Bay were taken when a seismograph was installed in a post-earthquake survey after the August 2003 M_w 7.2 earthquake, which occurred on the plate interface beneath Doubtful and Thompson Sounds. The photos showed what appeared to be raised wave-cut platforms, consistent with a series of uplifts in past large plate interface earthquakes. The lowest platforms suggested uplift considerably greater at the coast than that in the August 2003 event (Reyners et al., 2003), and hence, if confirmed, were probably caused by much larger earthquakes centred nearby. For this reason, this site was included as one of the three sites to be investigated for evidence of tsunami deposits in this project. Section 4 reports on the results of this reconnaissance.
In relation to the uplift, the 1793 chart of Gaol Passage and Doubtful Sound made by Bauza of the Spanish ship, *Descubierta*, has also been checked but does not show enough detail of either the shoreline or water depths to compare with present day charts.

3.2.2 Landslides and rockfalls

An earthquake of much larger magnitude than the August 2003 M_w 7.2 earthquake is also suggested by the apparently greater extent and severity of ground damage and landslides in the 1826 earthquake. The main area of landsliding in the August 2003 event extended over 3000 km^2 (Reyners et al., 2003) and landslides in the mountains would have been observable along about 80 km of coastline. The implication of the last paragraph of Taylor (ca. 1854) is that significant landslides/rock falls in the mountains and along the coast were visible along a large part of the Fiordland coastline, possibly 180 km or more, south of Cascade Point (near Jacksons Bay) (Fig. 1). It is unlikely that the name Cascade Point could refer to any other location than that which is recognised today. Although there are several locations with the word Cascade in their name (including a cove in Dusky Sound), Cascade Point is one of the earliest names given, appearing on James Cook's 1770 chart (Hall-Jones 1984).

Using the attenuation relationships of Dowrick & Rhoades (1999) to predict isoseismals and the Modified Mercalli descriptors for ground damage of Hancox et al. (1997) (ie significant landsliding occurs at ≥MM7), this large extent of ground damage is consistent with a very large earthquake of about magnitude 7.6-7.8 (see isoseismal maps, Section 6).

3.2.3 Tsunami and seiching

The earthquakes also seem to have been accompanied by tsunami or seiching within Dusky Sound, where the sealers were stationed. Seiches, that is, oscillations in large bodies of water in response to the passage of earthquake surface waves, often occur at large distances from large earthquakes, particularly those greater than M_w 7.5. For a tsunami large enough to cause concern to the sealers, there needs to be significant uplift over a large area of the sea-floor, or large sub-aerial or submarine landslides nearby to their location. As noted in Section 2, the August 2003 tsunami was relatively small. A plate interface earthquake of the magnitude suggested by the uplift and the extent of landsliding would be consistent with the generation of a significant tsunami.

3.2.4 Timing

Taylor (ca. 1854) specifies that the largest of the series of earthquakes occurred in the second half of 1826. This is not consistent with the timing of the 1820s Southland tsunami that reportedly occurred in autumn (as discussed in the previous section). However, there is some uncertainty in the accuracy of the 1826 date.

Although an extensive search for archival material relating to the earthquakes was outside the scope of this project, the records of Meurant (Meurant 1843; 1840s diaries and letters), James Boultbee (Starke, 1986; 1826-7 journal), Captain John Grono (Grono-Books Association, 1984) and Carrick (1903) are in the National Library of New Zealand. Grono, a sealing captain, who had a sealing station in Doubtful Sound, was not in Fiordland in 1826 and nothing is mentioned of any earthquakes in the book his descendants have written from family records. Boultbee's journal (Starke, 1986) is especially important as he was sealing on the Fiordland coast in 1826-27 and according to Starke (1986) mentions no earthquake. Further, he refers to feeling his *first* earthquake at a much later time (1833) in Indonesia (Starke, 1986). Starke notes in her introduction (p. xlv) that Boultbee's story in his 1826-7 journal of well-established sealers' base camps is inconsistent with the changes purported to have occurred in the 1826 earthquakes, and that the changes must have occurred at a later time.

As mentioned previously, there are few accounts of earthquakes for this time (Grouden, 1966), but one account that may be relevant to the 1826 earthquake is that in Carrick's 1903 collection of historical records of pre-1840 events (reprinted in Carrick (2005)). His source was reportedly an 1820 Tasmanian newspaper, but it is uncertain whether Carrick is quoting an article or interpreting it. The article relates to changes observed by a sealing captain called Edwardson in the coastal margin and river systems at the end of Edwardson Sound between visits in 1814 and 1818, and again in 1820 (Fig. 1). In 1814 a river about 4 km long exited from a lake and dropped into the sound over a precipice. By 1818, Edwardson observes that the river and waterfall had disappeared and the sound terminated in a shallow tidal lagoon with only a small channel leading from it to the sound. By 1820, the small lagoon had opened out to a large lagoon/harbour with a deep water entrance channel, the lagoon being fed by two rivers. Edwardson attributed the changes to changes in the river system from the lake and Edwardson (or Carrick, interpreting Edwardson's account) makes a further comment: "Evidently the convulsion which wrought so much havoc amongst the seal islands [in Dusky Sound] must have been severely felt at this place." It is possible that this refers to the 1826 event in Taylor, and it could be speculated that the 1826 date could be 1814-1818. The question of whether the effects in Edwardson Sound could be the result of an earthquake cannot be speculated on here as it would require a reconnaissance of the area.

The reliability of Carrick's account or the source is also suspect. A facsimile edition published by Cadstonbury Publications in 2005 (Carrick, 2005) includes an uncomplimentary review of Carrick's (1903) publication in an excerpt from Hansard, the record of New Zealand Parliamentary debates, in 1905. The review suggests that there were many inaccuracies and errors in Carrick's transcriptions and sources. The original source information should be verified before reaching any conclusions as to whether Edwardson's account is trustworthy.

While there is some uncertainty in the date, the description of the earthquake effects is so plausible that the occurrence of a large earthquake at the time when the sealers were working the coast (i.e from 1792), is credible. We will continue to refer to the earthquake as 1826 until such time as a more definitive date is determined.

3.2.5 One mainshock or several large shocks?

Taylor's (ca. 1854, 1855) accounts suggest a sequence of earthquakes, with no suggestion of one event being much greater than the others. While it is possible that the landslides were caused by a number of large events, a more likely scenario, to account for the uplift, is that the sequence of shocks represents a mainshock-aftershock sequence. The extended sequence over 6-7 months and the clear suggestion of quite large aftershocks is consistent with the M7.6-8 magnitude indicated by the extent of landslides and uplift.

In summary, the effects of the earthquake in about 1826 are consistent with it being a very large plate interface earthquake. As a consequence, two large plate interface earthquake scenarios, one with, and one without, surface rupture, are used in the development of dislocation and tsunami propagation models (Section 5.0). If uplift is not considered, then an Alpine fault earthquake is a potential source of the shaking and tsunami. Dislocation and tsunami propagation models are also developed for an Alpine Fault source (Section 5.0). Other sources, such as earthquakes on other offshore faults shown in figure 2, of the occurrence of landslides in association with the scenario earthquakes considered here, are possible.

4.0 PALEOTSUNAMI RECONNAISSANCE

OBJECTIVE: To investigate possible sources of the 1820s Southland tsunami and the occurrence of similar tsunami in the pre-historic geological record by identifying sites for future in-depth investigation and by coring up to three sites on the Fiordland / southern Westland coast judged to be suitable for preserving deposits of tsunami, including dating of up to three samples.

4.1 Introduction

The purpose of paleotsunami research is to find physical evidence for past large tsunami. Such evidence can help confirm early historical reports and provide indications of the source, size, extent of inundation, and frequency of past tsunami at a particular site. Currently no physical evidence of the 1820s Southland tsunami has been documented but research is underway at several sites along the Southland coast. Assuming the source of the 1820s Southland tsunami was a local earthquake, it was most likely generated to the west or southwest of Southland somewhere along the Alpine Fault, the Fiordland subduction zone or the Puysegur subduction zone. The Fiordland coast would be the first to be inundated by tsunami triggered in at least two of the above scenarios (refer to section 5.0). Although sparsely populated, the proximity to potential tsunami sources makes Fiordland worthy of paleotsunami reconnaissance. Any findings regarding tsunami magnitude and frequency in Fiordland are likely to be of relevance to the more highly populated Southland coast in cases where large earthquakes are the triggering mechanism. Relevance to populated localities in Fiordland, such as the tourist destination of Milford, would need to be assessed on an individual basis with detailed propagation modelling to determine how tsunami waves travel into and within the fiords.

4.2 Methods

The investigation was restricted to three sites because of the logistics involved. In addition to the remoteness, the very steep topography and high-energy wind and wave regimes make Fiordland a challenging place to search for paleotsunami. The general requirements for a good paleotsunami site include:

- Potentially exposed to inundation by tsunami
- Sheltered from storms
- Possessing a seaward source of sediment that could be entrained by a tsunami
- Possessing a sedimentary record into which sediment could be transported and preserved
- Able to be investigated with manual sampling techniques (ie, sites requiring drilling through a deep water column were beyond the scope of this project)

Much effort was put into selecting promising sites before visiting Fiordland. Sites between Te Waewae Bay in the south and Barn Bay in the north were rated for their potential to preserve tsunami deposits based on aerial photos and consultation with people familiar with the sites (Appendix 3). Out of over 30 sites, only five were rated highly. The top sites were reduced to three by selecting sites that would provide coverage over northern, central and southern Fiordland, and a good range of environments, aspect and tectonic setting. Initially Martins Bay, Sutherland Sound and Goose Cove were chosen for site visits. However as the historical work progressed, Cascada Bay near the mouth of Doubtful Sound was identified as potentially being the cove where sealers had reported uplift in the 1826 earthquake (refer to section 3.0). Therefore it was considered worthwhile to visit Cascada Bay in the search for evidence of uplift (and any resulting tsunami). This site replaced Sutherland Sound.

The project involved reconnaissance-level, rather than detailed, paleotsunami investigation techniques. Each site was initially viewed on aerial photos to select potential features of interest to be visited on the ground. Mapping of coastal features was undertaken during site visits using Real Time Kinematic Global Positioning Systems (RTK-GPS) and / or simple levelling equipment. Mean sea level was estimated by comparing observed tide levels with tide tables from Land Information New Zealand and NIWA’s tide forecaster. Where relevant sedimentary sections were exposed as outcrops, these were logged and sampled. A hand-augur was used to sample sediment beneath shallow lagoons and estuaries. Shell identification was used to provide paleoenvironmental information where possible. Two samples from each site were radiocarbon dated to provide an indication of the age of various events and landforms. Samples were collected with permission from the Department of Conservation (permit # SO/009/Res).

Field observations for each locality were made on three short site visits in early 2004 as listed below:

25 February, 0800: Flew Te Anau – Cascada Bay with Southwest Helicopters
26 February, 1500: Flew Cascada Bay – Te Anau with Southwest Helicopters

30 March, 1000: Flew Hollyford airstrip – Martins Bay Lodge with Air Fiordland
1 April, 1600: Flew Martins Bay Lodge – Hollyford airstrip with Milford Helicopters

21 April, 0830: Flew Clifden – Goose Cove with Southwest Helicopters
22 April, 1630: Flew Goose Cove – Clifden with Southwest Helicopters

4.3 Field Observations and Discussion

4.3.1 Martins Bay

Martins Bay is a large bay (>5 km wide) in northern Fiordland at the mouth of a valley containing Lake McKerrow and Hollyford River. The Alpine Fault crosses near the mouth of Lake McKerrow (Fig. 6) and runs offshore c. 20 km south of Martins Bay. The bay is exposed to the open sea to the west and would be vulnerable to tsunami triggered by faults offshore in the Milford Basin (see Figure 2B) or those travelling along the coast from further south. There are several potential sites on the coastal plain for recording recent and older large tsunami. Three main localities within the bay were investigated and sampled; these are described below.

Figure 6 Locality map of Martins Bay in northern Fiordland showing sites investigated for this project and the position of the Alpine Fault. 1:50,000 topographic map 260-D39 courtesy of Land Information New Zealand (Crown Copyright Reserved).

1. Boulder-cobble surface

At the southern end of Martins Bay (grid reference: D39 108353) a set of high back-beach sand dunes are sparsely vegetated with pingao and flax. A landward set of dunes have an active seaward face but are densely vegetated with shrubs and trees at their crest. Between these sets of dunes a relatively flat boulder-cobble surface extends about 100 m in a shore-normal direction and almost a kilometre parallel to the shore (Fig. 7). The surface is covered in sub-rounded to sub-angular boulders and cobbles of a variety of lithologies. Clasts appear to be lying flat on the surface without any imbrication. Coarse sand, small pebbles and shells also occur on the surface and all are fine enough to be wind-blown. Some shells have definitely been wind-blown because living moss was found underneath them. Pieces of wood up to about 50 cm long and occasional larger logs are scattered across the surface. There is some relief to the surface with cusp-shaped ridges extending parallel to the shore.

Figure 7 Boulder-cobble surface at Martins Bay looking southwest with seaward dunes at far right and landward dunes at far left. Low boulder ridges are visible centre-left.

The boulder-cobble surface is underlain by medium-grained, well-sorted grey sand. A pit dug beneath the surface revealed three boulder-cobble units that may represent similar concentrations of large clasts as seen at the surface currently (Fig. 8).

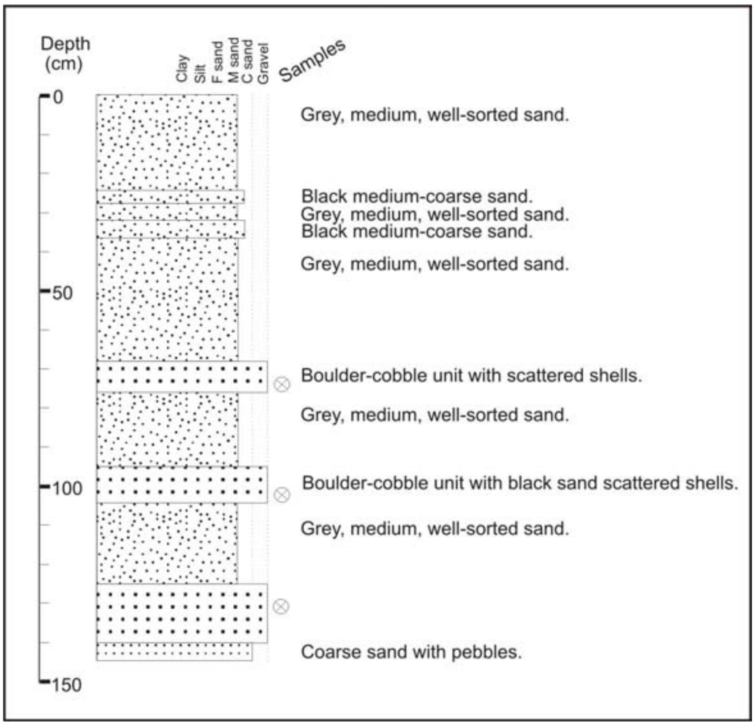


Figure 8 Stratigraphy observed in pit dug beneath the boulder-cobble surface.

Although boulder-cobble layers on sand have been interpreted as tsunami deposits at other sites in New Zealand (eg, Nichol et al., 2003), this deposit is not very high above sea level and is most simply explained as a deflation surface. Such a surface forms through concentration of the large clasts in a storm beach by gradual removal of fine clasts. The height and extent of surrounding dunes attest to the presence of winds strong enough to carry out such a process. A modern storm beach surface south and seaward of the boulder-cobble surface indicates there is a source of boulders and cobbles that get incorporated into the modern beach (Fig. 9).

Figure 9 Boulders and cobbles on the modern storm beach at the southern end of Martins Bay.

2. Wetland-lagoon sequence

The coastal plain south of the lower Hollyford River includes several wetland and lagoonal areas oriented parallel to the shore that may be old courses of the lower Hollyford River (Wellman and Wilson, 1964) or swales between successive beach ridges. At present they are low-energy environments accumulating peat that could be good repositories of high-energy tsunami deposits. In the wetland at the southern end of McKenzie Lagoon (grid reference: D39 112354), auger holes revealed half to one metre of peat at the surface underlain by at least two metres of sand that could not be penetrated. There was no evidence of high-energy deposition within the peat and the significance of the basal sand could not be established. In places (eg, D39 111353), a very hard sand unit near the surface, possibly an iron-pan, prevented auguring in the wetland.

3. Shellbed Section

In the west bank of the lower Hollyford River, downstream of the mouth of Lake McKerrow, sections of sand and gravel are exposed and are thought to represent beach and barrier sediments built up during rising sea level in the early Holocene (Wellman and Wilson, 1964; Hull and Berryman, 1986). At the landward extent of these exposures, immediately downstream of the mouth of Lake McKerrow, estuarine sediments also exist with *Paphies australis* (pipis) in growth position. One sand unit was previously recognised as containing a rich molluscan fauna with species transported from deep water (Wellman and Wilson, 1964; Hull and Berryman, 1986). This was considered worthy of re-sampling to determine whether a tsunami could have been the mechanism for transport of the deep-water shells.

The section was re-located (updated grid reference: D39 133348), described and sampled (Fig. 10). The unit of interest is about a metre below current river level so it was difficult to sample and describe in detail in situ. It is a unit c. 30 cm thick of poorly sorted pebbles, sand and silt. The occurrence of a diverse molluscan assemblage was confirmed with identification of over 30 different species (Table 1). The presence of shells from various environments was also confirmed with identification of rocky shore species, offshore bay, and estuarine tidal channel species (see notes included in Table 1). The mix of grain sizes and species present in a single unit indicates a chaotic agent of deposition.

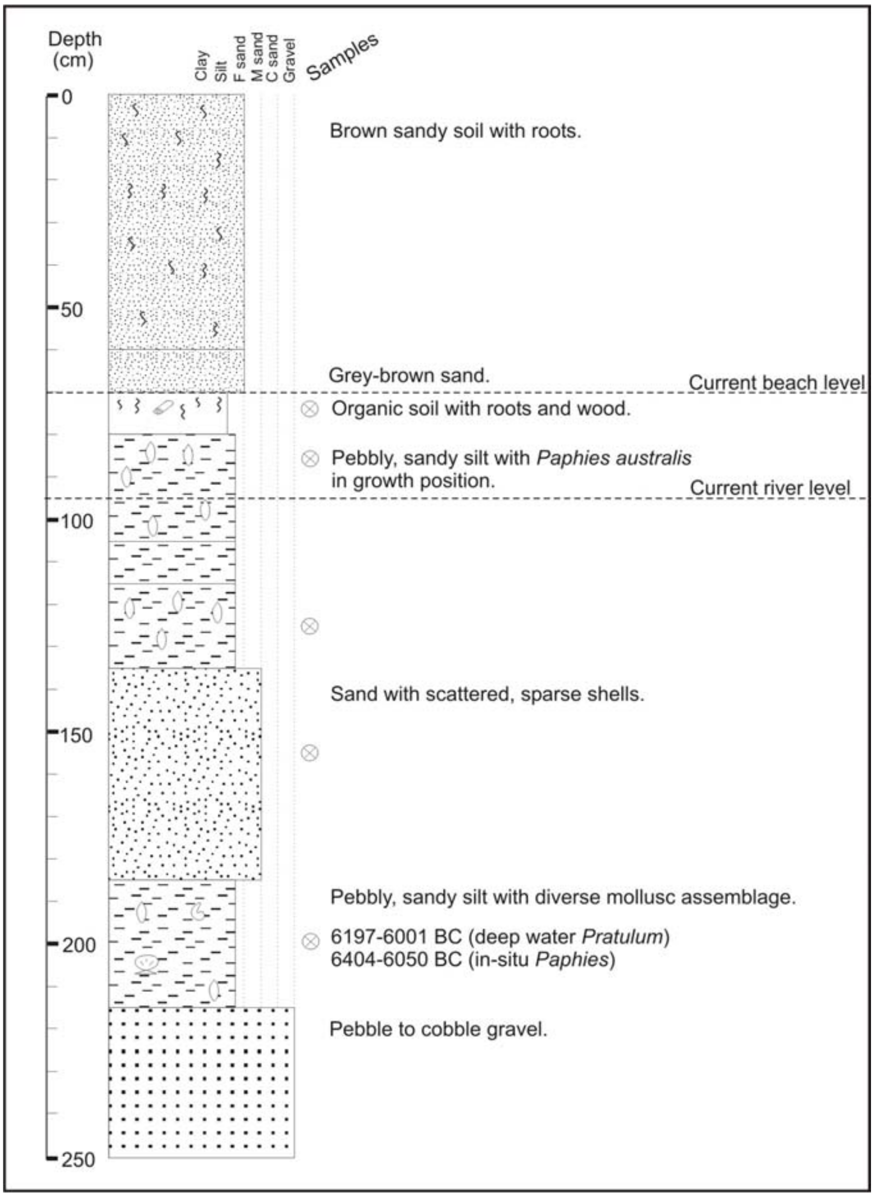


Figure 10 Stratigraphic section in the west bank of the lower Hollyford River near the mouth of Lake McKerrow (grid reference: D39 133348). The shellbed of interest lies between 185 and 215 cm on this stratigraphic log. NB: River level and river bank beach level appeared average at time of fieldwork.

Table 1 Molluscan assemblage from a pebbly, sandy silt unit 90-120 cm below river level near the mouth of Lake McKerrow. Re-collected from Wellman's locality with identification and ecological notes by Alan Beu.

Bivalvia:
Paphies australis
Tawera spissa
Pecten novaezelandiae
Mytilus galloprovincialis
Ruditapes largillierti
Diplodonta (Zemysina) globus
Nucula ?*dunedinensis*
Leptomya retiaria
Pratulum pulchellum
Pleuromeris zelandiae
Pecten novaezelandiae
Mytilus galloprovincialis
Modiolus areolatus
Borniola reniformis

Gastropoda:
Maoricolpus roseus
Trochus (Thorista) virids
Notoacmea helmsi
Trochus (Caoelotrochus) tiaratus
Diloma sp.
Micrelenchus cf. *tenebrosus*
Micrelenchus ?*huttoni*
Micrelenchus caelatus
Potamopyrgus antipodarum
Trichosirius cavatocarinatus
Ranella australasia
Xymene plebeius
Xymene aucklandicus
Neoguraleus sp.

Polyplacophora ("chitons"):
Rhyssoplax aerea

Brachiopoda:
Magasella sanguinea

Crustacea:
Barnacle plate

Polychaeta:
Smooth "worm" tubes

Ecology:
No specimens of *Austrovenus* are present. While this fauna is certainly dominated by abraded, broken and disarticulated "pipi", which probably lived in a semi-estuarine tidal flow channel in the lake entrance (ie, more-or-less at the deposition site), the only other taxa likely to have lived in this same environment are *Ruditapes*, *Nucula, Xymene plebeius, Neoguraleus,* and perhaps *Diloma* (depending on which species is represented). All the other taxa present are likely to have been transported into the deposition site from a more exposed environment, and in many cases from a little further offshore than the immediately adjacent beach (*Diplodonta, Pecten*, some *Tawera, Maoricolpus, Pratulum, Pleuromeris, Trichosirius, Magasella*). Exposed sandy beach taxa are scarcely present – either in this or in Wellman's previous collection. *Mytilus, Modiolus, Borniola, Rhyssoplax, Ranella,* the *Micrelenchus* species (very common), *Trochus viridis, Xymene aucklandicus,* the barnacle, and perhaps *Diloma* are rocky shore taxa, transported a little way along from the rocks near the channel entrance. *Potamopyrgus* is a freshwater gastropod, presumably transported down-stream from the lake.

While a much larger sample would be required to duplicate Wellman's fauna (S105/f504, = D39/f7504), the collection listed here leads to the same conclusion as originally reached – this is NOT the estuarine, *Austrovenus*-dominated fauna that would be expected to live in the tidally-influenced part of Lake McKerrow. It is the fauna of a tidal channel and of the beach, rocky shore, and slightly offshore from them, just outside the entrance channel; ie, perhaps deposited in a bay-bar, or perhaps all the non-estuarine taxa have been transported over the bar into the channel. The lack of obligate sandy beach taxa is possibly explained by the mildly sheltered nature of Martin's Bay; it may well have such taxa as *Tawera, Ruditapes* and *Leptomya* living just off the beach at present, rather than obligate sandy beach taxa such as *Mactra, Crassula, Paphies donacina* ("tuatua"), and *Dosinia*. However, that is unlikely; perhaps the large, thin, sandy beach bivalves just get broken on the bar.

Shells from a semi-estuarine tidal channel environment (*Paphies australis*) and an offshore bay setting (*Pratulum pulchellum*) were selected for separate radiocarbon analyses to determine whether either component had been reworked from older deposits. The age results (Table 2) show that, if anything, the estuarine shell is reworked, as it is slightly older than the open bay shell. However the ages are similar and it is statistically possible, within the 95% confidence interval, that they are the same (Rodger Sparks, pers. comm. 2005). If many shells from this deposit have similar radiocarbon ages it becomes more likely that they all died at the same time and so it also becomes more likely that the agent that brought them together from different environments also killed them. For example a large storm or tsunami could have stripped living shells from rocks around the headlands and from the floor of the bay and transported them into a semi-estuarine, tidal channel environment. The ages of both shells indicate that the unit was deposited before sea level had reached its current elevation. Therefore greater knowledge of the paleogeography of Martins Bay at this time would be required to fully interpret the significance of this deposit. Numerous sections along the lower Hollyford River provide an opportunity to reconstruct the area between Lake McKerrow and the sea and potentially provide a record of catastrophic events between c. 6500 and 4000 BC.

Table 2 Radiocarbon age results for Martins Bay samples.

Sample Depth (cm)	Sample Material	Dating Technique	$\delta^{13}C$ (‰)	Radiocarbon Age (radiocarbon years BP)	Calibrated Age 1 sigma (cal. years BP)	Calibrated Age 2 sigma (cal. years BP)	Lab Number
185-215	Shell *Paphies australis*	Standard	1.6	7691 +/- 76	6343-6050 BC	6404-6050 BC	WK 16452
185-215	Shell *Pratulum pulchellum*	AMS	0.2	7554 +/- 30	6148-6039 BC	6197-6001 BC	NZA 21843

4.3.2 Cascada Bay

Cascada Bay is a small bay (<100m wide) on the south side of Doubtful Sound not far from the mouth of the sound (grid reference: B43 295282). There are local faults in the basement rock that lie along the lineation of the valley behind Cascada Bay (Oliver, 1980) but their age and activity is unknown. The major sources of earthquakes in the area are more likely to be the thrust faults of the Fiordland subduction zone and the Alpine Fault that lie immediately offshore of Doubtful Sound. Cascada Bay was visited to look for evidence of uplift in the 1826 or other earthquakes and any associated evidence of tsunami inundation because it is a possible candidate for the small cove referred to as 'the jail' in historical reports (refer to section 3.0). The bay is sheltered from the south and west and is deep enough for a small boat to approach close to the shore without grounding. A small waterfall on a tributary stream may be the source of the name 'Cascada' but this is not visible from the shore. A stream flows into Doubtful Sound in the centre of the bay. On the right bank near the stream mouth a series of elevated bedrock platforms exist and on the left bank, gravel terraces and beaches have formed (Fig. 11). These features are described below.

Figure 11 View of Cascada Bay looking NE towards Doubtful Sound (Secretary Island in the background). Bedrock platforms can be seen on the true right bank of the stream in the centre of the photo and gravel storm beaches on the left bank in the centre left of photo. Landslides visible on Secretary Island were caused by the August 2003 M_W7.2 earthquake. *Photo: Graham Hancox.*

1. Bedrock platforms

On the eastern side of Cascada Bay near the stream mouth, a series of narrow steps in the gneiss bedrock is preserved (Fig. 12 i). There are five platforms (labelled A to E) stepping from sea level up to 8.5 m above mean sea level (amsl). Platform A is the lowest at c. -0.5-0 m amsl, it is very narrow and completely submerged at high tide; platform B at c. 0.75 m amsl is partially submerged at high tide (Fig. 12 ii and iii). The most extensive platform (C at c. 2-3.5 m amsl) consists of an elevated flat area with a large raised rock pool and resembles a wave-cut platform (Fig. 13). The greatest elevation difference is c. 4.75 m between platform C and the 6-7.5 m high platform D (Fig. 14). The highest platform (E at c. 7.5-8.5 m amsl) is only preserved as a small remnant at one point within the bay (Fig. 13). All levels contain a fair amount of relief in part due to differential weathering of veins / beds in the bedrock. Therefore only approximate step heights are reported from this initial survey: A to B = 1 m; B to C = 2 m; C to D = 4.75 m; and D to E =1.25 m.

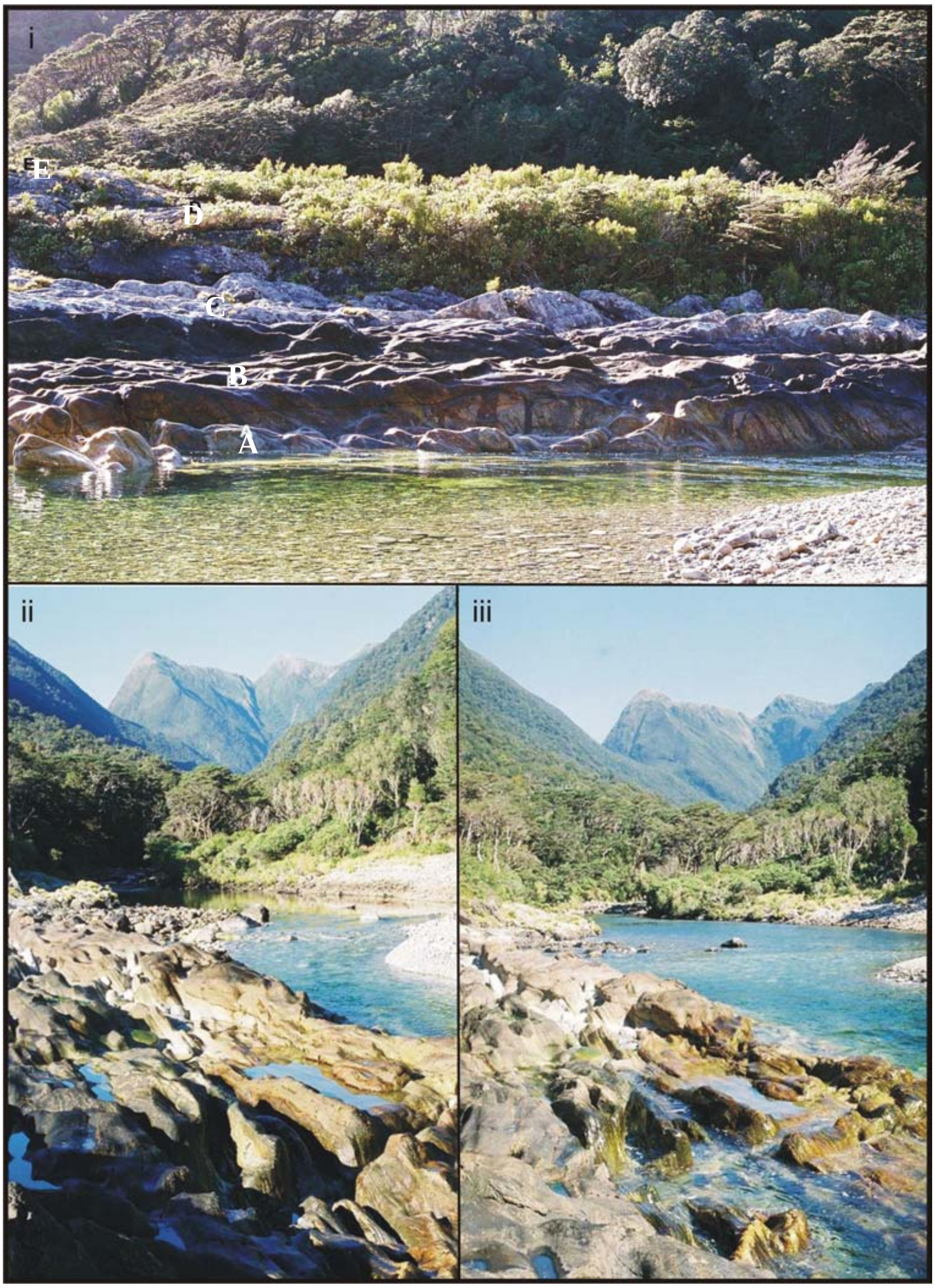


Figure 12 i) Raised bedrock platforms on the right bank near the mouth of the stream in Cascada Bay. Five levels are recognised (labelled A-E). ii) View upstream from platform B at low tide. iii) View upstream from platform B just after high tide.

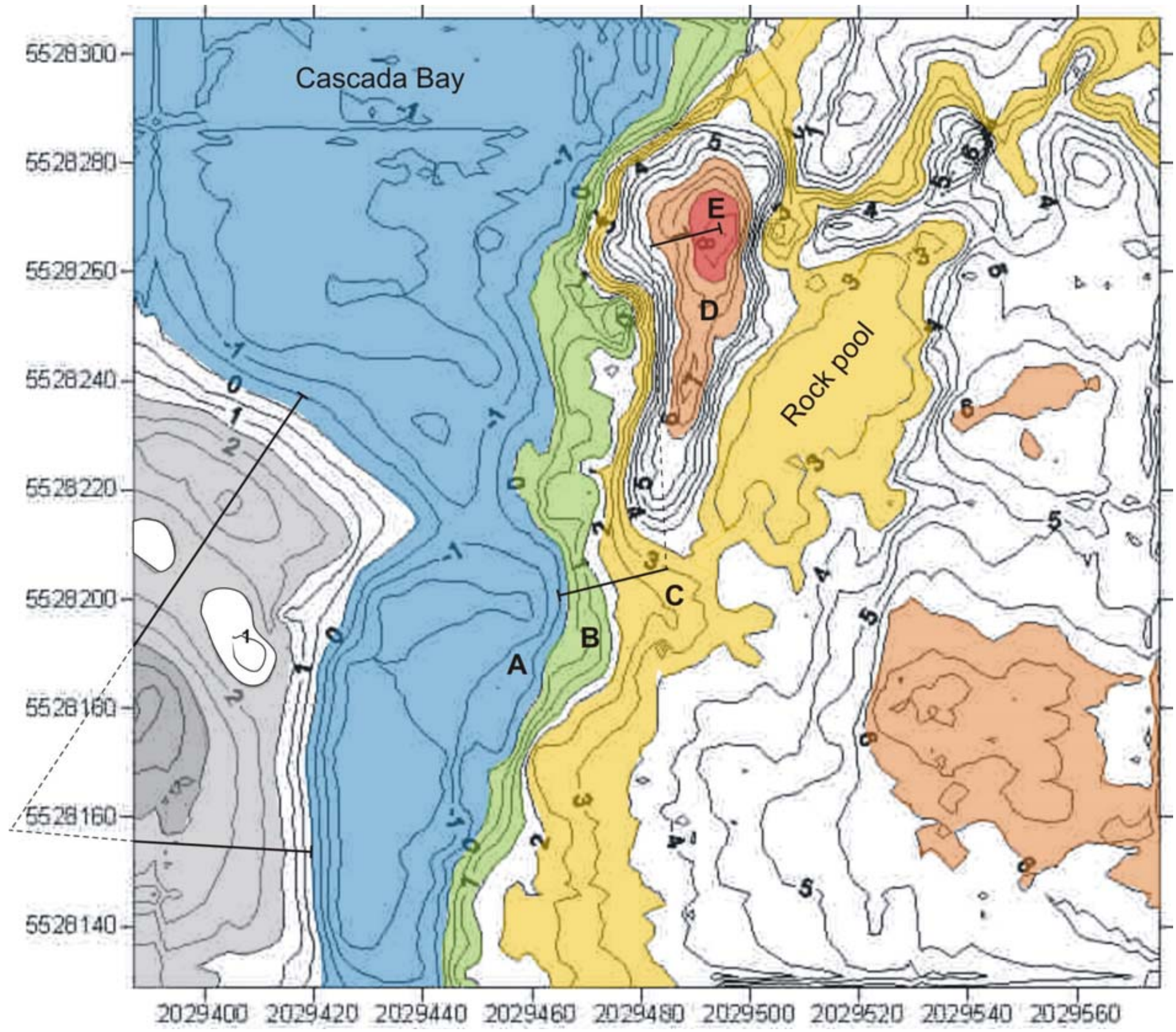


Figure 13 Topographic map of the stream mouth in Cascada Bay created using RTK-GPS. The stream and sea are in blue; other colours highlight elevations of bedrock platforms B to E; platform A is a narrow strip at water level (platform elevations were measured with a dumpy level, see Figure 14). Shades of grey highlight the beach ridges on the left bank but it was not possible to capture the full extent or topographic detail of the left bank because bush and steep hills prohibited satellite coverage. Lines mark approximate positions of the profiles shown in Figures 14 and 16.

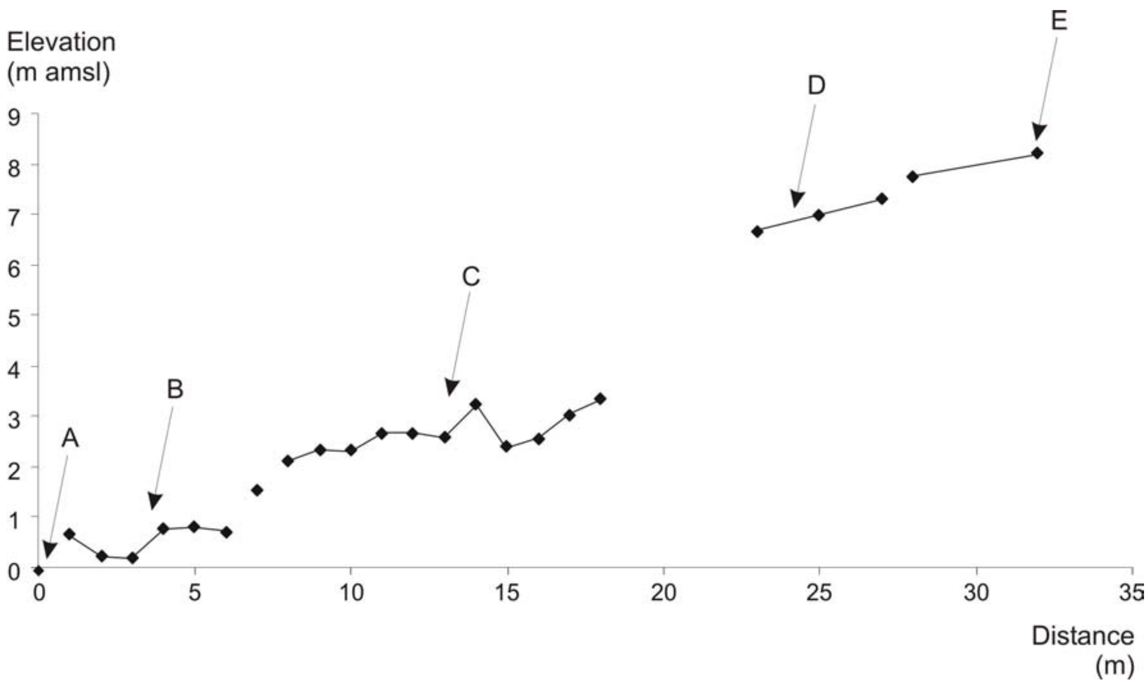


Figure 14 Profile across the stepped bedrock platforms at the approximate position of the solid lines in Figure 13. This profile was measured from a known tide position using a dumpy level. Platform elevations are reasonably consistent with the topography measured by RTK-GPS as shown in Figure 13.

The stepped nature of these benches and the relationship of platform A to current sea level suggest the platforms could be wave-cut features that have been raised above sea level suddenly in large earthquakes. One piece of evidence that could help in determining their origin would be their age. Platform C, around the western and southern margins of the rock pool, has a cover of c. 40 cm of sand on which tussock and shrubs grow. The sand is poorly sorted, medium to coarse grained with numerous angular to rounded pebbles of various rock types and up to 5 cm long. No shell material was found but chiton covering from shells was collected and radiocarbon dated. The negative radiocarbon age result for the chiton sample (Table 3) implies some contribution from ‘artificial’ atomic bomb carbon in the atmosphere and thereby a modern (post-1950) age for the chiton (Rodger Sparks, pers. comm. 2005). Modern wind-blown shells were observed on the surface of platform C so it is likely that the shells within the cover sediments were also modern and wind-blown. Dissolution of shell material (leaving chiton) obviously occurs rapidly in the high rainfall regime of Fiordland.

Table 3 Radiocarbon age results for Cascada Bay samples.

Sample Depth (cm)	Sample Material	Dating Technique	$\delta^{13}C$ (‰)	Radiocarbon Age (radiocarbon years BP)	Calibrated Age 1 sigma (cal. years BP)	Calibrated Age 2 sigma (cal. years BP)	Lab Number
	Chiton on shell	AMS	-17.8	-657 +/- 30			NZA 21842
	Peat	Standard	-27.5	90 +/- 35	1820-1822 AD plus 1898-1909 AD	1697-1725 AD plus 1808-1909 AD	WK 16377

Platforms D and E are well vegetated with tussock, flax, shrubs and a few small beech trees. On platform E, a soil pit revealed 38 cm of peaty soil on top of the platform from which a sample of the lowest 10 cm of peat was collected for radiocarbon dating. The age result (Table 3) indicates that the soil started accumulating on platform E sometime between AD 1697 and 1909 (at the 95% confidence level). This is a far more recent initiation of soil development than expected if colonisation is assumed to begin soon after the platform was raised above sea level. The platform, at c. 8 m amsl, and with four platforms below it, is likely to have been above sea level for at least a thousand years. Therefore the age of the current soil is unlikely to relate to the age of platform exposure – it merely dates the most recent colonisation of the platform by plants. This raises the question of whether previous soil and shrub coverage has been removed from the platform in extreme storms or tsunami. The existing radiocarbon date has a wide error range but it does not exclude the possibility of an early 1800s event stripping the platform bare and re-colonisation occurring subsequently as has been documented at Okarito Lagoon (Goff et al., 2004). Stripped vegetation was seen after the 2003 tsunami triggered by a landslide in Gold Arm of Charles Sound, Fiordland (Power et al., 2005). The resolution available from radiocarbon could be improved through collection of a much smaller sample at the base of the peat (instead of a 10 cm thickness) and accelerator mass spectrometry (AMS) dating instead of standard radiocarbon analysis. However the best way to improve the age control would be through tree-ring dating of the few small trees that grow on the platform.

Possible correlatives of the platforms at Cascada Bay were observed from the air on many of the islands in the entrance to Doubtful Sound (for example on the western end of Bauza Island, on the Shelter Islands and the Nee Islets) (Fig. 15). Surveying of these platforms could provide an indication of the extent of the Cascada Bay platforms and may provide further opportunity for age control by dating of trees on some of the higher, vegetated platforms. Comparison of existing islands with those on early maps may provide some insight into how recently these features have become exposed above sea level.

Figure 15 The Nee Islets at the mouth of Doubtful Sound (Secretary Island in the background) where at least two platform levels were observed from the air.

2. Beach ridges and terraces

On the western side of Cascada Bay, a large boulder beach has formed with a small low-tide beach ridge at and below msl, a high-tide beach ridge at 1.8 m amsl, and a storm beach ridge at 3-4 m amsl (Fig. 16). The beaches consist of well-rounded pebbles to boulders of several lithologies, predominantly granite. They appear to be currently active. Behind the storm beach is a relatively flat area covered in large driftwood tree trunks. Upstream of the beach ridges on the left bank there are two terraces composed of rounded pebbles and cobbles and vegetated with relatively mature trees. The lowest terrace has an elevation of c. 5-5.5 m and the higher terrace an elevation of c. 7-7.5 m amsl. These are likely to be alluvial terraces or beach ridges relating to higher relative water levels. They possibly correlate with platforms C and D respectively but this relationship has not been explored.

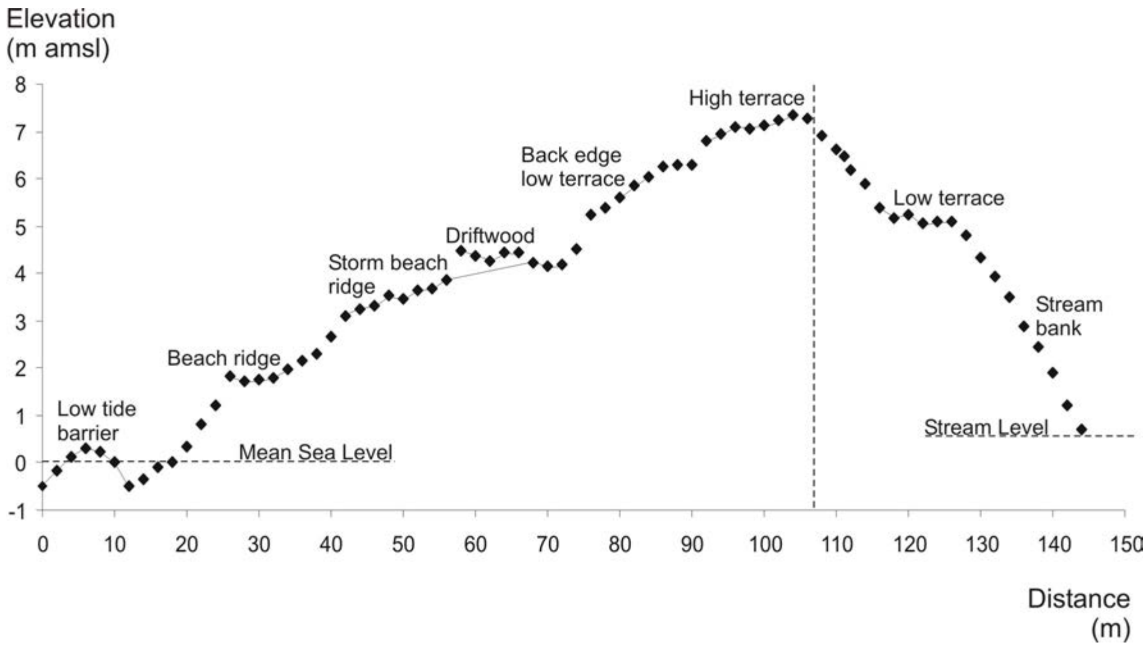


Figure 16 Profile across beach ridges and terraces on the left bank in Cascada Bay. The profile line made a sharp bend at the position of the high terrace (vertical dashed line) so the elevation of the low terrace could be measured perpendicular to the stream (approximate profile line marked by solid and dashed lines on the left bank and off the edge of the map in Figure 13). The profile was measured from a known tide position using a dumpy level. Beach elevations are thought to be more reliable than those in Figure 13 because poor sky visibility on the left bank led to difficulties in collecting RTK-GPS topographic data.

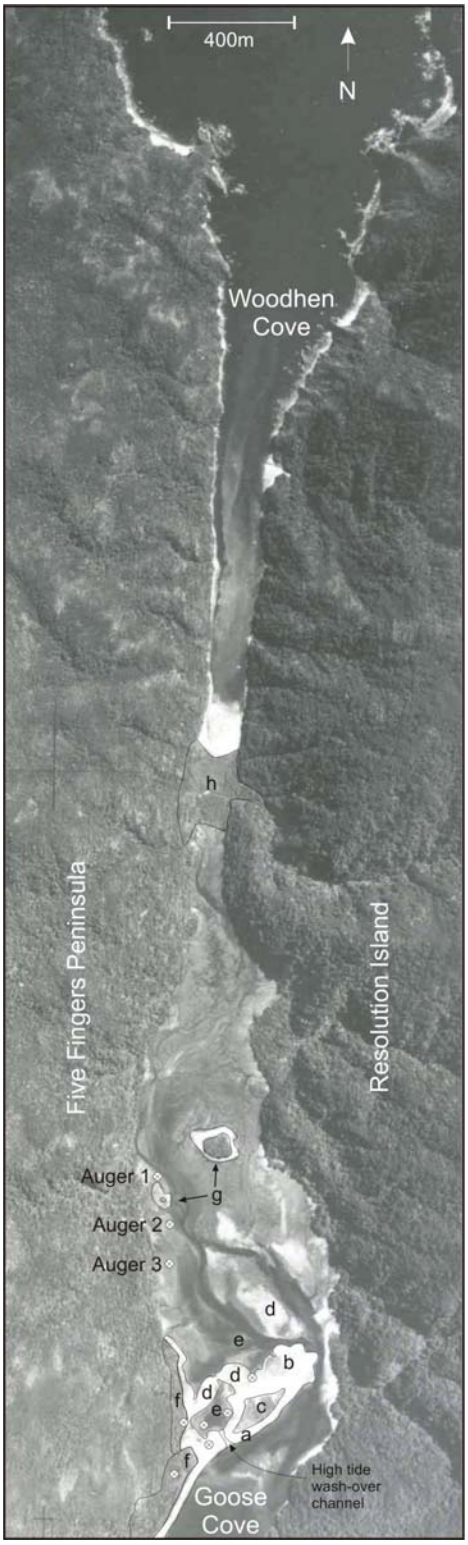
The mechanism of formation and age of platforms in Cascada Bay remain unknown but they are thought to be evidence for previous coseismic uplift. It is possible that platform B was raised 1 m in 1826 and that the lowest narrowest platform (A) has been cut in the last c. 180 years. However further investigation into rates of platform cutting and ages of the higher platforms would be required to strengthen this proposition. Also, relative uplift would not necessarily have been required for changes to the bay in 1826 to match those of the historical reports. Deposition of the large lobe of boulders where the beach ridges are at the mouth of the bay (Fig. 11) would have effectively blocked the upper reaches of the bay from boat access. Whether this supply of boulders came down the stream or from a local landslide as the result of strong shaking in an earthquake is unknown, but rounding of the boulders and formation of beach ridges suggests the lobe has spent some decades in its present position.

4.3.3 Goose Cove

Goose Cove lies between Five Fingers Peninsula and Resolution Island on the northern side of the entrance to Dusky Sound (Figure 1). It was selected for investigation because of the sheltered tidal inlet at its head. A normal fault runs along the inner edge of Five Fingers Peninsula and has the potential to generate large earthquakes. However the cove could also be exposed to tsunami triggered by earthquakes further offshore at the Alpine Fault, Fiordland or Puysegur subduction zones.

1. Goose Cove tidal inlet (grid reference: A44 080853)

Goose Cove would potentially be exposed to tsunami from the south if waves travelled north in Dusky Sound along the inner edge of Five Fingers Peninsula. Goose Cove could also be exposed to the open sea in the north if waves travelled the narrow channel of Woodhen Cove and overtopped the barrier at its head. In the middle reaches of Goose Cove a boulder bank extends northeast from Five Fingers Peninsula across the cove almost to Resolution Island. At the head of the cove an alluvial fan and / or landslide debris separates Goose Cove from Woodhen Cove to the north. These barriers enable a sheltered, tidal inlet to exist in the upper 1.5 km of Goose Cove (Fig. 17).



Key
a. Steep beach of large rounded boulders and driftwood.
b. Boulder ridges with shrubby vegetation.
c. High point on boulder bank of dune sand and established vegetation.
d. Gravel bars above high tide with *Apophlaea lyallii*.
e. Sand flats with eel grass, *Austrovenus stutchburyii,* crabs.
f. Bush-covered terraces of rounded boulders at similar height to storm beach crest. Southern terrace has c. 30 cm of well-sorted, medium sand on top.
g. Islands with vegetation. Large island appears to be of rounded boulders, small island of angular blocks - rockfall.
h. Fan / landslide debris.

Figure 17 Annotated aerial photograph showing features of the tidal inlet at the head of Goose Cove and position of the augers taken from the bed of the inlet.

The northern alluvial fan / landslide barrier was only observed from the air (Fig. 18 i and ii). It appears low enough to be overtopped by high waves but the driftwood and shrub cover may inhibit sediment transport into Goose Cove tidal inlet. Large driftwood tree trunks exist at the head of Goose Cove possibly from local tree falls or transport over the barrier or up the estuary during high water levels. The southern boulder bank is breached by a high-tide wash-over channel in the west (Fig. 18 iii) and by a deep and narrow tidal channel in the east (Fig. 18 iv). The elevation of the boulder bank is variable with a western section low enough to be overtopped by storm waves but with a high central section that has been built by wind-blown sand and vegetation (Fig. 18 v). Wash-over by storm waves is unlikely to transport sediment very far beyond the landward edge of the boulder bank. Wash-over by tsunami has the potential to entrain wind-blown sand from the boulder bank and transport it further into the tidal inlet. The lower reaches of the tidal inlet are sand-dominated making identification of transported sand difficult (Fig. 18 vi). However the middle to upper reaches of the inlet have higher silt and organic content, potentially making differentiation of transported sand easier. Sedimentary features of the boulder bank and inlet were investigated in soil pits and auger holes and these are shown in Figure 17.

Figure 18 i) View looking north from Goose Cove to the alluvial fan / landslide barrier at the head of the cove. ii) View of the alluvial fan / landslide barrier from above. iii) View of the high-tide wash-over channel on the southern boulder bank. iv) Tidal channel at the eastern end of the boulder bank. v) High central section of the boulder bank with dune sand, shrubs and small trees. vi) View of Goose Cove tidal inlet looking north from the boulder bank at low tide.

2. Sedimentary record of the inlet

Hand-augering into the bed of the inlet at low tide revealed the top c. 1.5 m of sediment at several sites (Fig. 17). At the base of all the auger holes a coarse shell hash unit was encountered that couldn't be penetrated (Figs. 19, 20, 21). Organic silty fine sand with small estuarine shells exists above the coarse unit. In auger 2 the base of this unit is highly organic and it was dated at 549-654 AD (95% confidence level) (Table 4). The top of this unit was dated at 1428-1484 AD (95% confidence level) in auger 3. Above the organic fine sand there is a sharp contact, preserved everywhere we looked in the inlet, to inorganic fine sand with large estuarine shells. This is interpreted as a sudden increase in salinity (larger shells) and a decrease in deposition of organic material, likely to be the result of greater marine influence. Generally, increased marine influence in an inlet is brought about by relative sea level rise or increased breaching of the barrier. At Goose Cove, a relative rise in sea level at this time (1428-1484 AD), is more likely to be the result of tectonic lowering of base level (eg, by an earthquake) than eustatic sea level rise because sea level was stable to within a few cm around this time (Gibb, 1986). The change to greater marine influence could also be the result of barrier dynamics at Goose Cove. The southern boulder bank could have been breached through a lack of sediment supply to maintain it and / or impact of high-energy events such as storms and tsunami. The age of the fan / landslide barrier at the head of the inlet is unknown but it is likely that emplacement of such a barrier between Woodhen and Goose Coves would have markedly changed the depositional environment in the inlet.

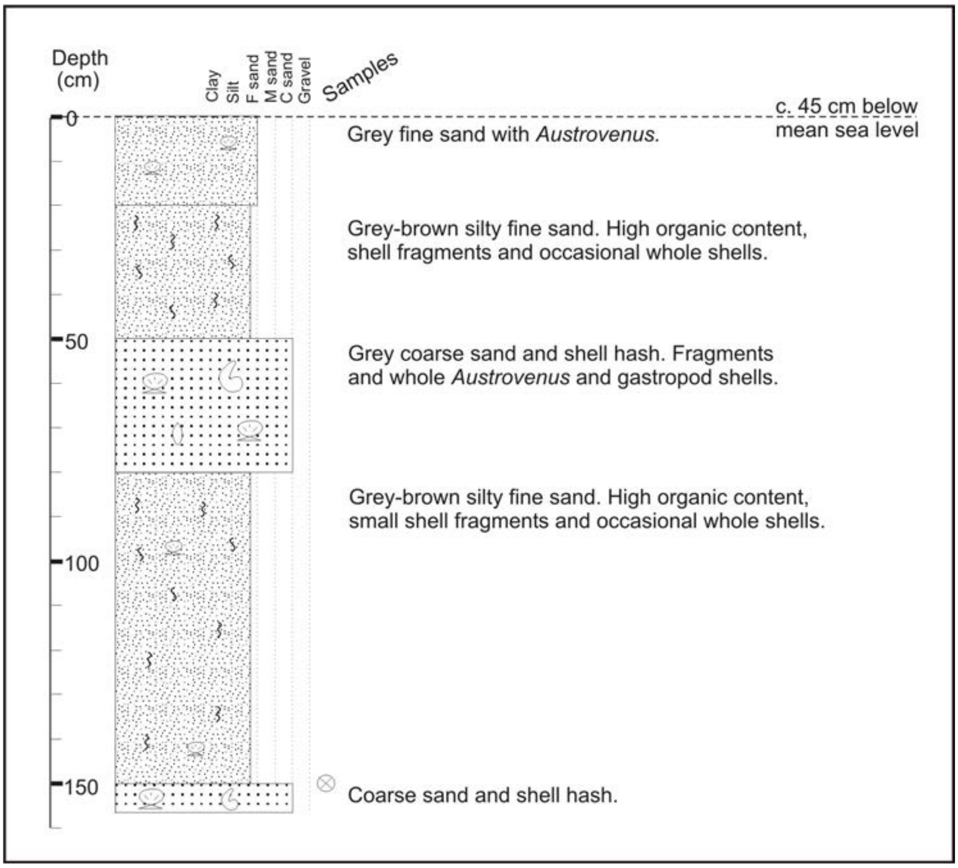


Figure 19 Stratigraphic log of Auger 1 in Goose Cove tidal inlet.

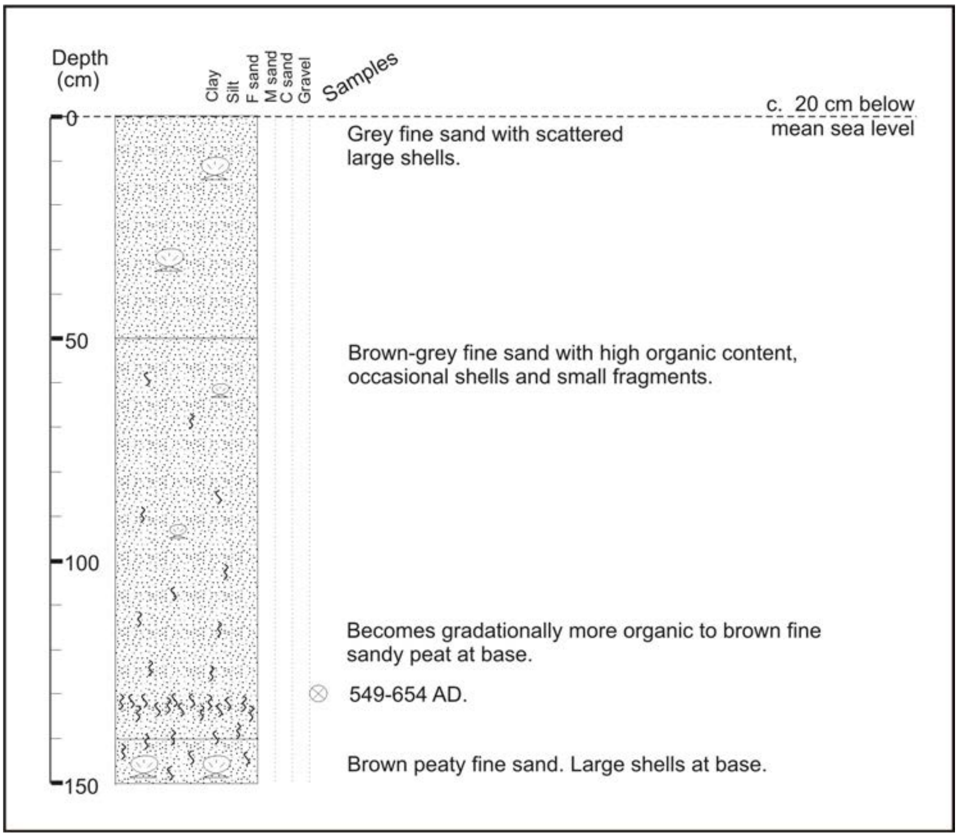


Figure 20 Stratigraphic log of Auger 2 in Goose Cove tidal inlet.

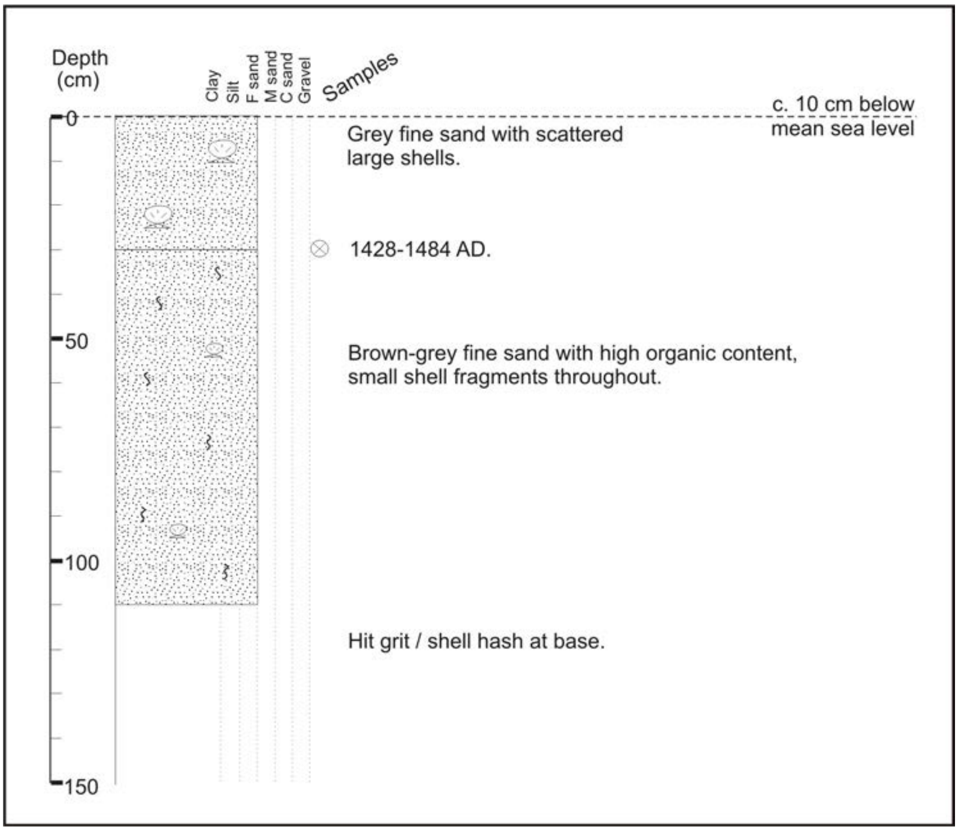


Figure 21 Stratigraphic log of Auger 3 in Goose Cove tidal inlet.

Table 4 Radiocarbon age results for Goose Cove samples.

Sample Depth (cm)	Sample Material	Dating Technique	$\delta^{13}C$ (‰)	Radiocarbon Age (radiocarbon years BP)	Calibrated Age 1 sigma (cal. years BP)	Calibrated Age 2 sigma (cal. years BP)	Lab Number
	Organic sand	AMS	-28	475 +/- 30	1441-1458 AD	1428-1484 AD	NZA 21951
	Organic sand	AMS	-30.3	1507 +/- 30	580-644 AD	549-654 AD	NZA 21844

An additional unit was found in auger 1 that was not identified in the other augers – a unit of coarse sand and shell-hash between 50 and 80 cm depth (Fig. 19). This is the only sediment indicative of high-energy conditions (apart from that not penetrated at the base of the augers). However, because of the lack of continuity between sites, this unit is likely to represent preservation of a localised tidal channel environment rather than a tsunami deposit.

The paleoenvironmental change identified in the top 50 cm of augers in Goose Cove, and dated at c. 1428-1484 AD warrants further investigation to determine its cause. It coincides with the age of an Alpine Fault earthquake in Westland but further research would be required before inferring any causal link with that earthquake. An extension of the north-south transect of augers could help determine the influence of the two barriers on the depositional environment in the cove. Dating the initial emplacement of the fan / landslide barrier at the northern end of inlet would also aid interpretation of the cove’s paleoenvironmental history. Microfossil analysis may provide estimates of change in tidal elevation, which would help identify movement of base level and consequently, the likelihood of an earthquake being the cause of change.

4.4 Summary

The steep topography and high-energy wind and wave regime of the Fiordland coastline means there are few sites likely to preserve high-resolution information about past earthquakes and tsunami. Landsliding, forest destruction and geomorphological changes to coastal waterbodies possibly provide the most robust physical evidence for past earthquakes and tsunami in Fiordland. The three sites selected for having good potential for paleotsunami or paleoseismological research, and visited as part of this project, all proved to contain features of interest that could warrant future work. The only possible evidence for the occurrence of an 1820s tsunami was the recent establishment of vegetation on a c. 8 m high platform in Cascada Bay, Doubtful Sound. Further age control for this patch of vegetation as well as vegetation at similar elevations on nearby islands and coves would be required to confirm that synchronous vegetation stripping occurred in this area in the 1820s. Cascada Bay also contained the most likely evidence of uplift in the 1826 Fiordland earthquake in the form of a narrow platform (B) raised c. 1 m above platform A. Evidence of older earthquakes and /

or tsunami appears to be preserved in the sedimentary record of the tidal inlet at the head of Goose Cove (c.1450 AD) and in the banks of the lower Hollyford River at Martins Bay (c. 6100 BC). Both of these sites would require further investigation to verify the inferred causes of change and it is likely that such work would uncover longer records of natural hazard events at both sites than the single events described here.

5.0 EARTHQUAKE DISLOCATION AND TSUNAMI PROPAGATION MODELS

OBJECTIVE: To investigate possible sources of the 1820s Southland tsunami and the occurrence of similar tsunami in the pre-historic geological record by developing elastic dislocation and tsunami propagation scenarios using computer models of one or two credible earthquake sources capable of producing tsunami and compatible with the descriptive account of the 1826-27 Fiordland earthquake.

5.1 Introduction

In the early stages of this work it was considered that the 1820s Southland tsunami may have been triggered by the 1826 Fiordland earthquake. Therefore emphasis for the modelling work was placed on developing a scenario compatible with the descriptions of the 1826 earthquake. As it became clear that the 1820s tsunami probably occurred at a different time to the 1826 earthquake, other earthquake sources were considered in an attempt to match the historical descriptions of the 1820s tsunami. No attempt has been made to model landslide or other non-seismic sources of local tsunami, or any distant sources.

There are numerous active faults offshore of Fiordland and Southland, many of which have been recently mapped, and their rupturing behaviour is as yet unknown, let alone their ability to trigger tsunami. In this study we have chosen three main components of the plate boundary that we consider capable of rupturing in large earthquakes that could trigger large tsunami: the Alpine Fault, the Fiordland subduction zone (Fiordland Basin) and the Puysegur subduction zone (Puysegur Trench) (Fig. 2). For each of these components, we have used maximum known fault lengths, thereby providing maximum earthquake magnitude scenarios. However it is as yet unknown whether the full length of these faults would rupture in a single earthquake. It is also unknown, for the subduction zones, whether the faults would rupture to the surface of the crust at the seafloor. Therefore, for each subduction zone, we have developed a surface-rupturing and non-surface-rupturing earthquake scenario. In this report five scenarios are modelled:

1. Alpine Fault rupture (Fig. 22)
2. Fiordland subduction zone event with surface rupture (Fig. 23)
3. Fiordland subduction zone event without surface rupture (Fig. 24)
4. Puysegur subduction zone with surface rupture (Fig. 25)
5. Puysegur subduction zone without surface rupture (Fig. 26)

This section of the report describes the fault dislocation models that were developed and then the tsunami model, the generation and resolution of the computational grid, the initial conditions, and results for each of the 5 scenarios.

5.2 Initial conditions: dislocation models

It is well-established, based largely on measured coseismic displacements from earthquakes around the world, that the earth's crust can be approximated as behaving elastically. Elastic dislocation theory has been developed to help calculate the surface displacements due to dislocations (or fault slip) in an elastic medium. When the ground is covered by water, such as a lake or ocean, this ground deformation can in turn be used as the initial condition for the generation of water waves. To estimate the surface displacement due to potential earthquake scenarios in the Fiordland region, we have employed an elastic, half-space dislocation modelling approach, using the equations of Okada (1985). One useful aspect of elastic dislocation modelling is that surface displacements scale linearly with fault slip, so that an event with 1 m of fault slip will cause twice as much surface displacement as an event with 0.5 m of slip on the fault. Therefore, the vertical displacements calculated using these elastic dislocation models can be scaled to consider events with larger or smaller amounts of slip than those used in the models presented here. For subduction zone earthquakes, we have estimated average slip on the fault using the scaling relationships of Abe (1975), assuming an average stress drop of 5 MPa.

5.2.1 Alpine Fault

The Alpine Fault scenario was developed from data of Barnes et al. (2005). The structure and geomorphology of the fault suggest that it ruptures as several linear sections, which are tens of km in length and have predominantly dextral strike-slip (horizontal) displacement and little vertical displacement, separated by oblique structures across "releasing bends" in the surface trace. These bends in the fault trace, which is otherwise essentially continuous, are associated with large basins. At each of the releasing bends there is increased vertical displacement, such that the sum of many fault ruptures has resulted in the observed morphology. The offshore section of the fault includes seven strike-slip basins, with step-over widths across some of the basins of up to 3 km (Fig. 2). Therefore a rupture on this fault would generate localised, elongate areas with significant vertical displacement, but little between the basins. The size of these areas of subsidence varies from c. 3 km by 8 km up to c. 5 km by 15 km. Considering the possible variations in subsurface fault geometry, simple calculations (eg, Barnes et al., 2001) indicate that for 9 m of dextral slip, the vertical displacement at the step-overs could exceed 5 m, and potentially >9 m.

From this information, we developed the displacement scenario shown in Figure 22. The modelled displacement scenario was generated assuming a pure strike-slip displacement of 9 m on the Alpine Fault over a length of c. 175 km. The model includes the three largest releasing bend basins on this section of fault. The dips used were 80° (eastward) on the strike-slip segments of the Alpine Fault, and 65° on the releasing bend segments; the model rupture extended to 20 km depth. As the displacement imposed on the fault was strike-slip, the majority of vertical displacement occurred at releasing bends along the fault (maximum subsidence, ~4.5 m). The vertical surface displacements in this model were calculated on a 1 km x 1 km grid. The resulting earthquake would be M_w 7.8.

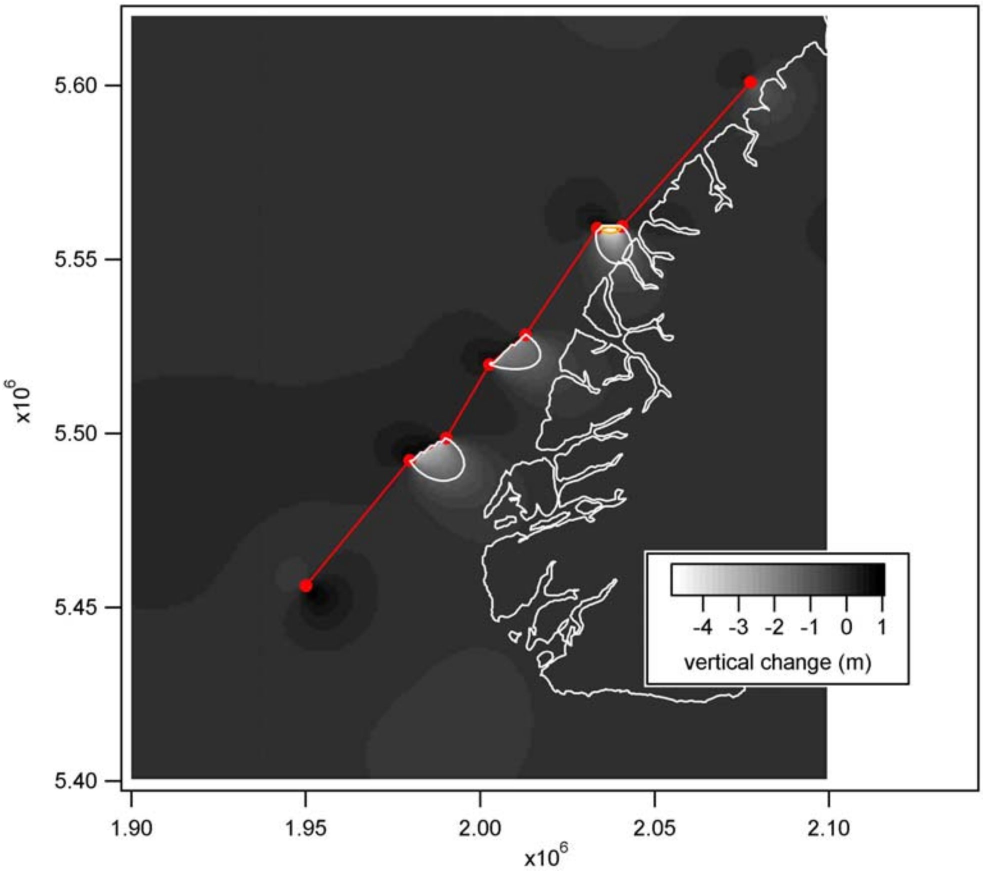


Figure 22 Vertical displacements (in m) from an Mw 7.8 Alpine Fault earthquake with rupture along several fault segments. Note the contour shown represents 1 m of subsidence, and the subsidence zone encircled by the 1 m contour ranges from 1 – 4.5 m.

5.2.2 Fiordland subduction zone

The two Fiordland subduction zone events approximate a rupture of the Fiordland subduction zone from c. 50 km north of Secretary Island down to the intersection of the Resolution Ridge with the plate boundary, one with a surface rupture (Fig. 23) and one without (Fig. 24). The event is similar to the 2003 Secretary Island event, but with increased rupture length to the south of Secretary Island. These earthquake scenarios reflect the length of the Fiordland subduction zone considered capable of producing large subduction interface earthquakes. Subduction does continue further north, but it becomes more oblique and has a narrower seismogenic width, limiting the capacity of this part of the zone to produce large earthquakes.

In the two scenarios, the total slip (all dip-slip) is 5 m on a fault plane dipping 30° towards the east-southeast. The strike of the northern end of the rupture segment is 30° (north of east), and changes to 13° at the southern end, based on seismological evidence for such a change in strike of the subduction zone (Reyners et al., 2002). Two areas of uplift can be seen in these figures. These are due to the fact that the events were modelled as two fault planes (to approximate the change in strike of the southern half of the rupture). The reason for the two humps is that the corners of the two planes meet at the downdip end, leaving a triangular gap between the planes in the up-dip direction. This gap is more pronounced in the surface rupture case, and is an artefact of the dislocation modelling. It does not reflect the observed surface thrusting. However, the models are adequate as source models for the tsunami.

The surface rupturing event (Figure 23) ruptures from the surface down to 23 km depth and would correspond to an M_w 7.9 earthquake, and has a maximum uplift of c. 2.5 m. The event with no surface rupture (Figure 24) has slip on a plane extending from 13 km to 23 km depth, and would be an M_w 7.7 event yielding a maximum uplift of c. 1.5 m. The surface displacements due to these events are calculated on a 1 km x 1 km grid.

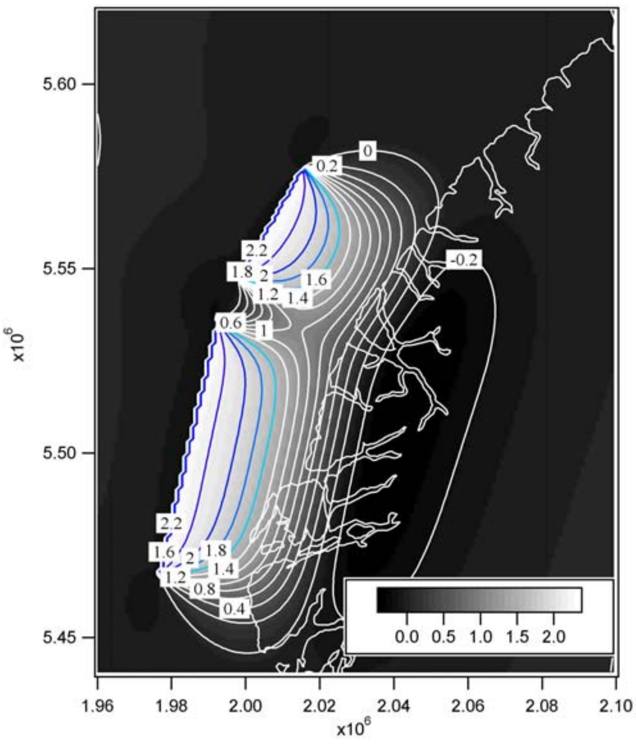


Figure 23 Vertical displacement (in m) from an M_w 7.9 Fiordland subduction zone earthquake that ruptures a large part of the subduction zone to the surface.

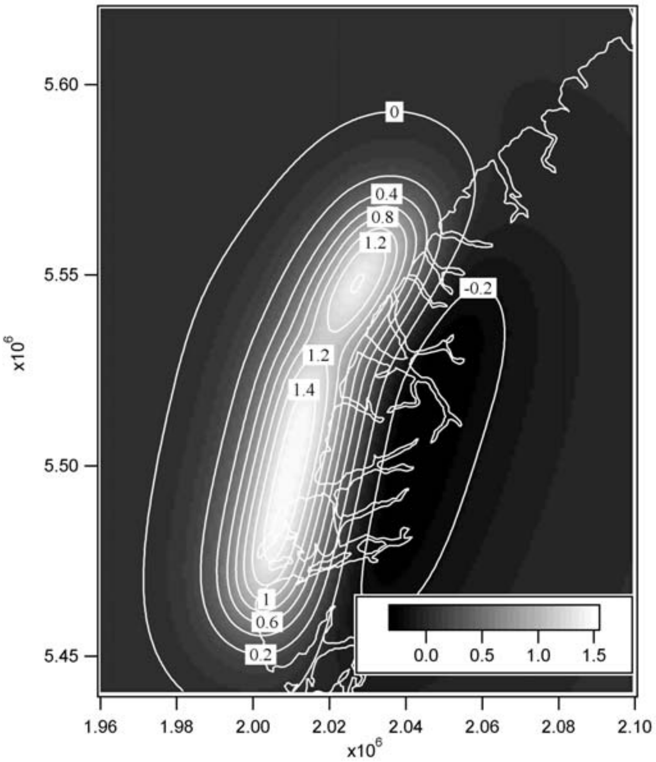


Figure 24 Vertical displacement (in m) from an M_w7.7 Fiordland subduction zone earthquake that ruptures a large part of the subduction zone without reaching the surface.

5.2.3 Puysegur subduction zone

The Puysegur subduction zone events are modelled as rupturing the total c. 250 km length of the Puysegur Trench, with 8 m of down-dip displacement on a subduction thrust dipping 10° to the east. Two scenarios are modelled: (1) a rupture from 25 km depth to the surface (Fig. 25), and (2) a rupture between 10 and 25 km depth (no surface rupture, Fig. 26).

The surface rupture case corresponds to an M_w 8.6 event and the no-surface-rupture case corresponds to an M_w 8.5 event. Surface displacements were calculated on a 1 km x 1 km grid. As these scenarios involve the whole known length of the Puysegur subduction zone and average slip on the fault determined using the scaling relationships of Abe (1975), they are likely to be close to maximum possible.

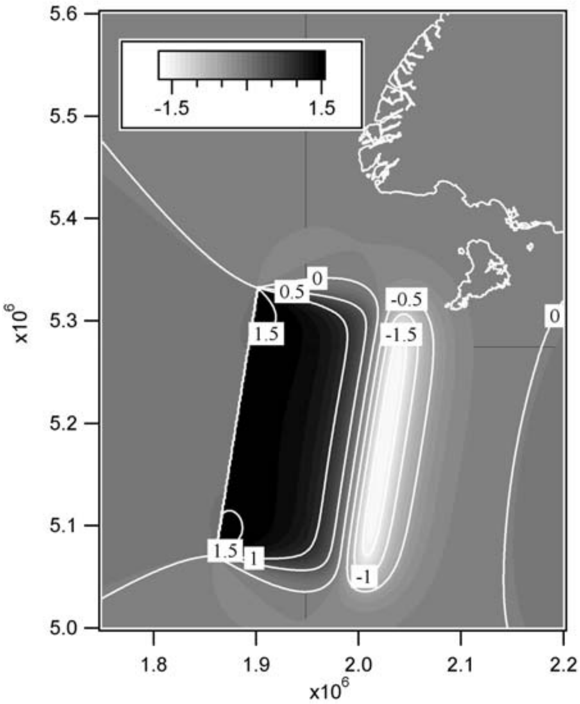


Figure 25 Vertical displacement (in m) from a M_w 8.6 Puysegur subduction zone earthquake that ruptures the length of the subduction zone to the surface.

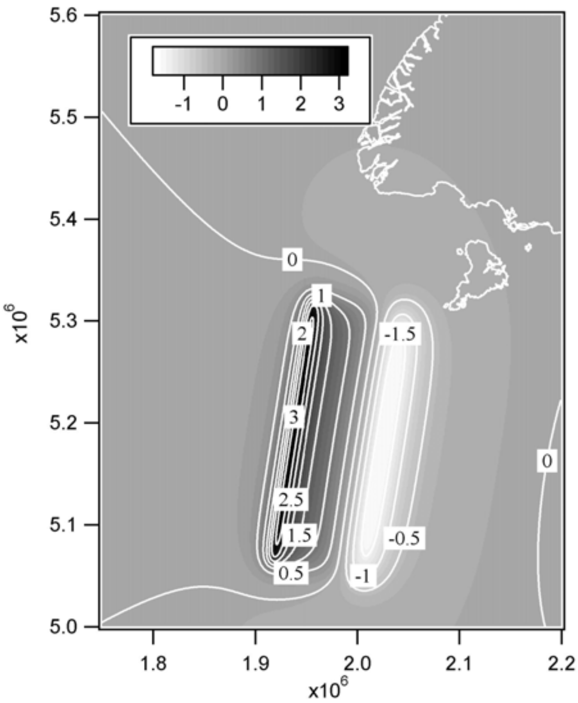


Figure 26 Vertical displacement (in m) from a M_w 8.5 Puysegur subduction zone earthquake that ruptures the length of the subduction zone without reaching the surface.

5.3 Numerical tsunami propagation model

The numerical model is a general-purpose hydrodynamics and transport model (RiCOM, River and Coastal Ocean Model). The model has been under development for several years and has been evaluated and verified continually during this process (Walters and Casulli 1998; Walters 2002, 2004, 2005a, 2005b). The hydrodynamics part of this model was used to derive the results described in this report. A more detailed description of the model can be found in Appendix 4 and in the references. A general overview is contained in the following paragraphs.

The model is based on the Reynolds-averaged Navier-Stokes equations (RANS) that are time-averaged over turbulence time scales. In addition, these equations are averaged over space to derive double-averaged equations (Finnigan 2000; Nikora et al. 2001). In this manner, sub-grid spatial effects (vegetation, bottom roughness, etc) can be included in a rigorous manner.

In order to allow flexibility in the model grid and complete tessellation of complicated coastal geometry, finite elements are used to build an unstructured grid of triangular elements of varying-size and shape. The vertical depth coordinate is mapped to a sigma or terrain-following coordinate that automatically provides enhanced vertical resolution in the near shore region (for 3-dimensional simulations). Since the interest here is on surface waves rather than internal dynamics, a vertically averaged, 2- dimensional version of the model is used in this study.

The time marching method is a semi-implicit numerical scheme that avoids stability constraints on wave propagation. The advection scheme is semi-Lagrangian, which is robust, stable, and efficient (Staniforth and Côté 1991). Wetting and drying of intertidal or flooded areas occurs naturally with this formulation and is a consequence of the finite volume form of the continuity equation and method of calculating fluxes (flows) through the triangular element faces.

At the shoreline boundaries of the grid, the natural boundary condition for this finite element implementation is that the normal flux is zero. This boundary condition can be modified through boundary integrals and allow specification of sea level, open boundary conditions, and normal flux conditions. At open (sea) boundaries, a radiation condition is enforced so that outgoing waves will not reflect back into the study area, but instead are allowed to realistically propagate through this artificial boundary and into the open sea.

The equations are solved with a conjugate-gradient iterative solver. The details of the numerical approximations that lead to the required robustness and efficiency may be found in Walters and Casulli (1998), Casulli and Walters (2000), and Walters (2002).

5.3.1 Model grid and bathymetry

The model grid has a number of requirements that ensure that the model calculations will be accurate and free of excessive numerical errors (Henry and Walters 1993). The primary requirements are that the triangular elements of the model grid are roughly equilateral in shape and their grading in size is smooth from areas of high resolution (small elements) in the coastal zone to areas of low resolution (large elements) in the offshore ocean areas. The bathymetry (sea-bed depths) for this study was obtained from three different sources (NIWA, digitised hydrographic charts and a global bathymetric database) as detailed in Walters et al. (2001).

In this study, the shoreline boundary was obtained by sub-sampling the LINZ 1:50,000 shoreline dataset. This was carried out in such a way as to achieve the minimum number of points while retaining a smoothly graded boundary and ensuring that, at any point within the fiords, the spacing between adjacent boundary nodes was less than half the width of the fiord at that point. Creating a shoreline by these restrictions ensures that each fiord has at least one line of nodes running up its centre allowing accurate interpolation of bottom topography. Note that no land topography was incorporated into the grid, basically due to time restrictions but also due to the lack of such data. Thus the edges of the model grid will act as vertical walls, and the maximum wave-height is due to a reflected wave.

Run-up on land is highly variable and depends on the characteristics of the incident wave and the details of the land topography. As a wave approaches shore and the water depth decreases, the wave speed decreases so that the wave height increases and the wavelength decreases. As the wave runs across land, there are drag effects from large objects such as boulders, vegetation and structures, and the wave is guided by the detailed variation in land topography. As a result, the spatial variability in wave run-up is dependent on the spatial variability in topography and surface roughness. While many models can properly treat this variability (such as the model used here), the necessary input data describing the topography is usually nonexistent. While using a vertical wall at the edge of the grid represents steep cliffs rather well, it is only a crude approximation for run-up over a more general topography.

The interior part of the grid was generated using the program GridGen (Henry and Walters 1993) according to the requirements described above. First a layer of elements is generated along the boundaries using a frontal marching algorithm (Sadek 1980). Then the remaining interior points are filled in using the cluster concept in Henry and Walters (1993). This grid was subsequently refined by a factor of four by subdividing each grid triangle successively into 4 new triangles using vertices at the mid-sides of the original triangle. The resulting grids contained 160880 elements and 85041 vertices for the Fiordland scenarios and 204996 elements and 107791 vertices for the Puysegur scenarios. Depths at the vertices were interpolated from high resolution bathymetry data.

5.4 Results

Traditionally, these types of tsunami have been modelled using the shallow water approximation rather than dispersive wave theory because of the additional computational overhead of the latter. Typically, models using the Boussinesq approximation for weakly dispersive waves are an order of magnitude slower and have a commensurately larger memory overhead than shallow water models. Hence, it is useful to determine the appropriate wave theory for these scenarios from the outset.

For assessing the applicable wave theory, the appropriate measure is the depth to wavelength ratio H/L. For the shallow water approximation to hold $H/L<1/20$, for intermediate waves $1/20<H/L<1/2$, and for deep water waves $H/L>1/2$. Scaling the Alpine Fault deformation, $H/L=0.2$, approximately, within the intermediate wave classification. For the subduction zone events, $H/L=0.1$ for the long wave and this ratio has larger values for the short wave components in the surface rupture case. For the Puysegur events, $H/L=0.04$ for the long wave and the ratio has larger values for the short wave components in the surface rupture case. Hence, for all these cases, dispersive waves are a significant part of the dynamics. The behaviour was verified during model simulations and is particularly evident in the Alpine Fault simulations. For the results presented here, the dispersive wave version of the numerical model was used (see Appendix). This particular model is more efficient than Boussinesq models, and is about three times slower than the shallow water version for this particular problem.

In addition, the extremely narrow continental shelf adjacent to Fiordland (Figure 27) had important consequences in the results. First, all the tsunami that were generated in deep water had large amplitude increases when they shoaled on the continental shelf. Combined with the additional amplification on run-up, all scenarios were capable of generating run-up heights of at least 4 m. Second, the narrow shelf does not allow significant coastally trapped waves. Hence, there were insignificant wave effects outside the immediate generation area as will be pointed out in the sections that follow.

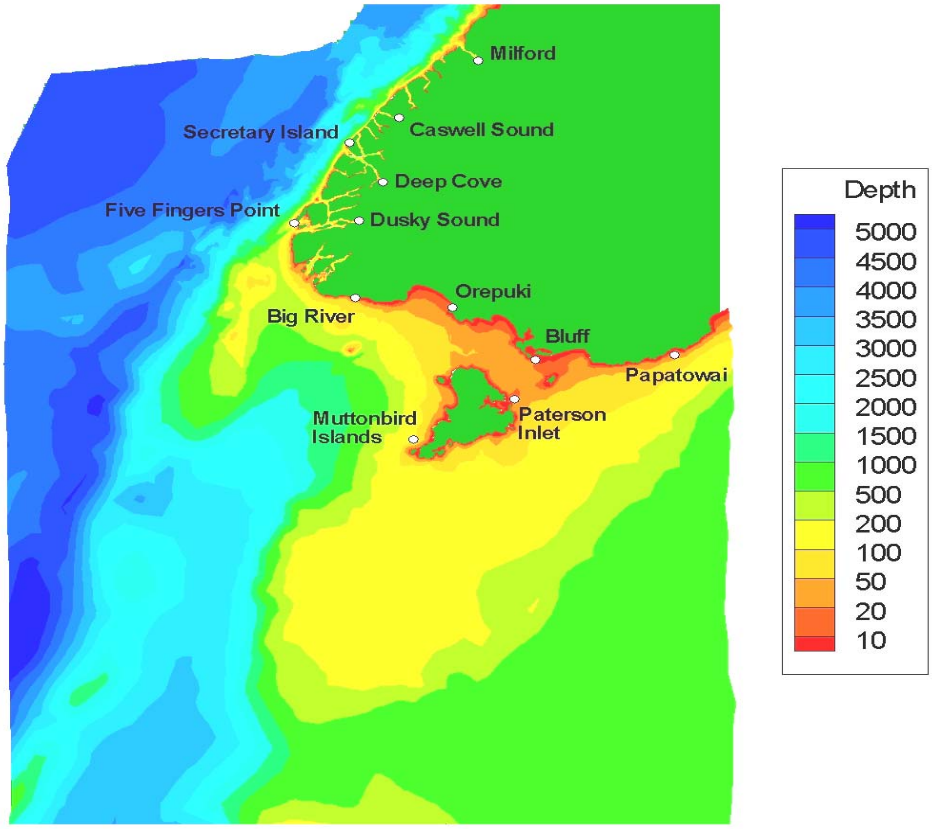


Figure 27 Map of the shoreline and bottom topography (in m) showing the locations (denoted by white dots) where sea level was recorded. Note the narrow continental shelf adjacent to Fiordland, the shallow ridge south of Fiordland, and the shelf south of Stewart Island. These features are all important in modifying the propagation of the tsunami.

5.4.1 Alpine Fault

For the Alpine Fault scenario, the shoreward propagating tsunami initially evolved into a conical-shaped wave similar in appearance to a solitary wave. There were five such waves generated with the centre three being the largest. These waves passed onto the continental shelf where the wave height increased three-fold because of shoaling transformations. The wave height increased another two-fold when the waves were reflected by land. The run-up effects were very local to the area shoreward of the generation area ie, the main impact of the tsunami was along the Fiordland coast with relatively minor perturbations of sea level along the Southland coast. The response of the fiords was variable and depended on a number of factors including the extent of constriction at the sill, the resonance characteristics of the fiord, and the location of the entrance with respect to the generation area. The highest water elevations occurred in Bligh (7.5 m), Nancy (5.1 m) and Dagg (3.6 m) Sounds [2]. However,

[2] Bligh Sound lies between Milford Sound and Caswell Sound; Nancy Sound, between Caswell and Sound and Secretary Island; Dagg Sound, between Secretary Island and Five Fingers Point (Figure 27).

more detailed modelling that includes coastal topography would be required to assess the effects of such a tsunami on the coastline within each fiord. The envelope of maximum water elevations is shown in Figure 28. The sea level response at various sites (Figure 27) is shown in Figure 29.

Due to the location of the rupture, the time for the first wave to reach the nearest open coastline is short (less than 2 minutes). At more distant locations, such as within the fiords the first wave arrives later. For example, inundation within Nancy Sound occurs within ten minutes of the earthquake.

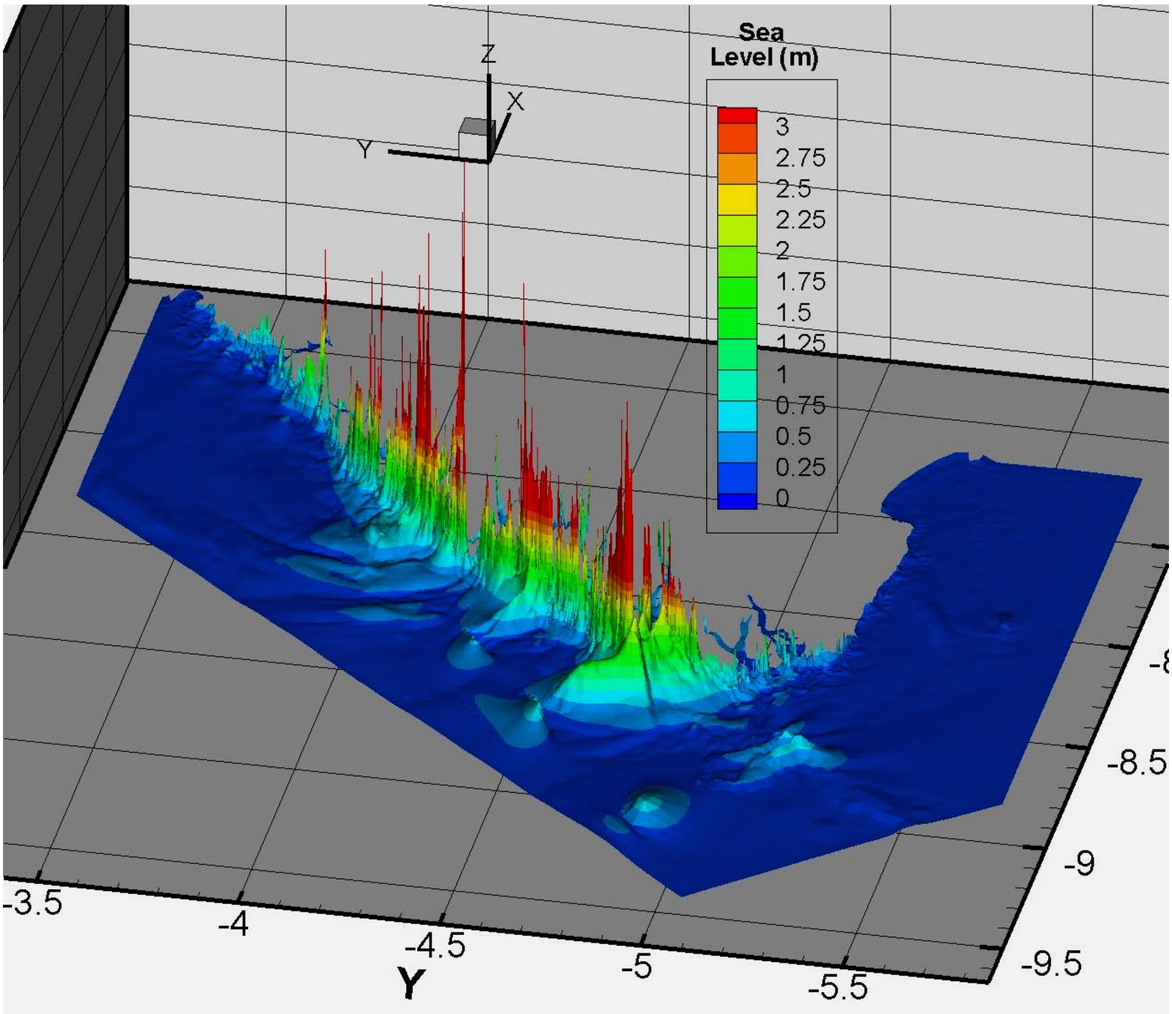


Figure 28 Envelope of maximum water levels during the Alpine Fault earthquake scenario.

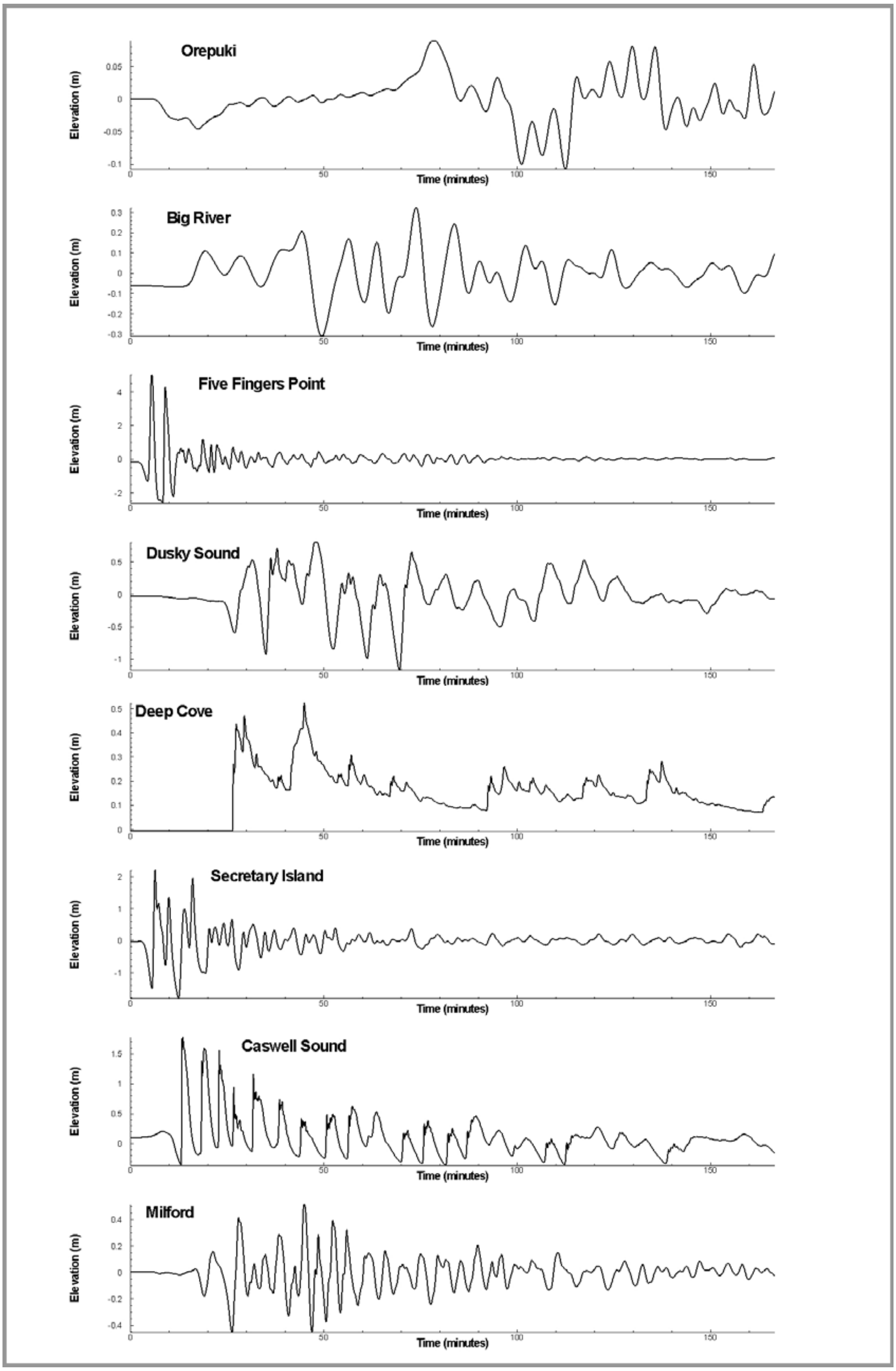


Figure 29 Time-series of water-surface elevation following the Alpine Fault earthquake scenario. The locations are shown in Figure 27.

5.4.2 Fiordland subduction zone events

In contrast to the Alpine Fault scenario, in which the tsunami was mainly generated at fault step-overs as point sources, the Fiordland subduction zone scenarios are characterised as line sources (compare Figure 22 with Figures 23 and 24). Hence the wave run-up effects are more uniform and are experienced over a larger area.

The initial wave decomposes into an offshore propagating wave that heads into the Tasman Sea and towards Australia, and a shoreward propagating wave towards Fiordland. Of the two scenarios, one with surface rupture and one without surface rupture, the former has the largest sea-floor deformation and hence the largest run-up (Figs. 30, 32). In addition, the surface rupture is located well seaward of the shelf break so that the resulting wave has strong amplification when it shoals on the continental shelf. On the other hand, the dislocation model without surface rupture is closer to shore with the result that the shoaling effect is not as pronounced. In both scenarios, maximum water levels occur along the Fiordland coastline with only small wave heights predicted for the Southland coastline. As with the Alpine Fault scenario, there is a mixed response in the individual fiords with Deep Cove and Bligh Sound generally having a large amplified response, and lesser amplification elsewhere. However we have not explored these variations in any detail. Time series of sea level after the two scenarios of fault rupture at various sites (Fig. 27) are shown in Figures 31 and 33.

For both subduction zone scenarios, arrival of the main wave at the nearest open coastline occurs within two minutes of the earthquake. Two locations that experience high water elevations are Deep Cove and Milford Sound, where the first main wave arrives 20 and 25 minutes respectively after the earthquake. Deep Cove experiences three large waves, the second and third large waves arriving at approximately 35 and 65 minutes after the earthquake.

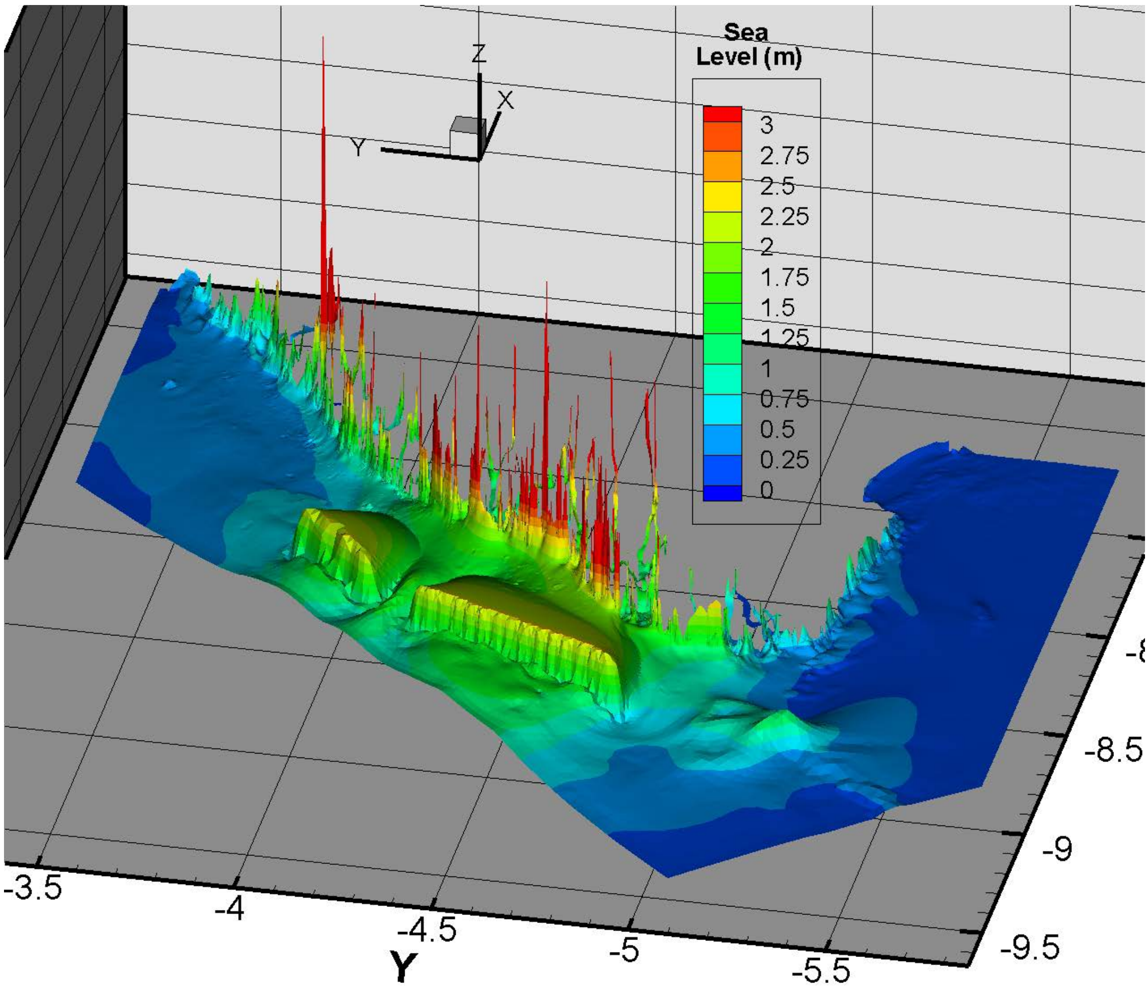


Figure 30 The envelope of maximum water levels during the Fiordland subduction zone earthquake with surface rupture scenario.

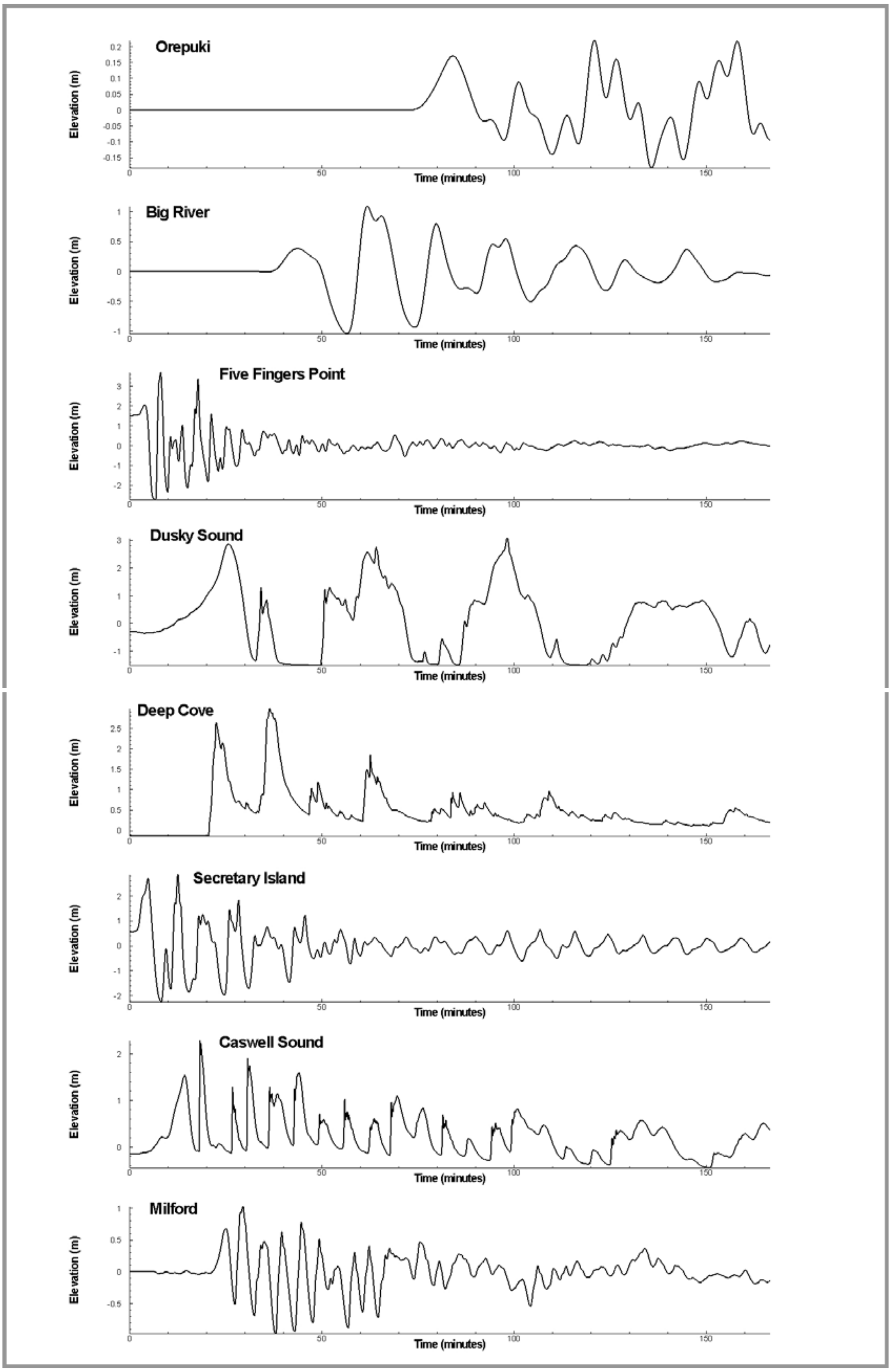


Figure 31 Time-series of water-surface elevation following the Fiordland subduction zone earthquake with surface rupture scenario. The locations are shown in Figure 27.

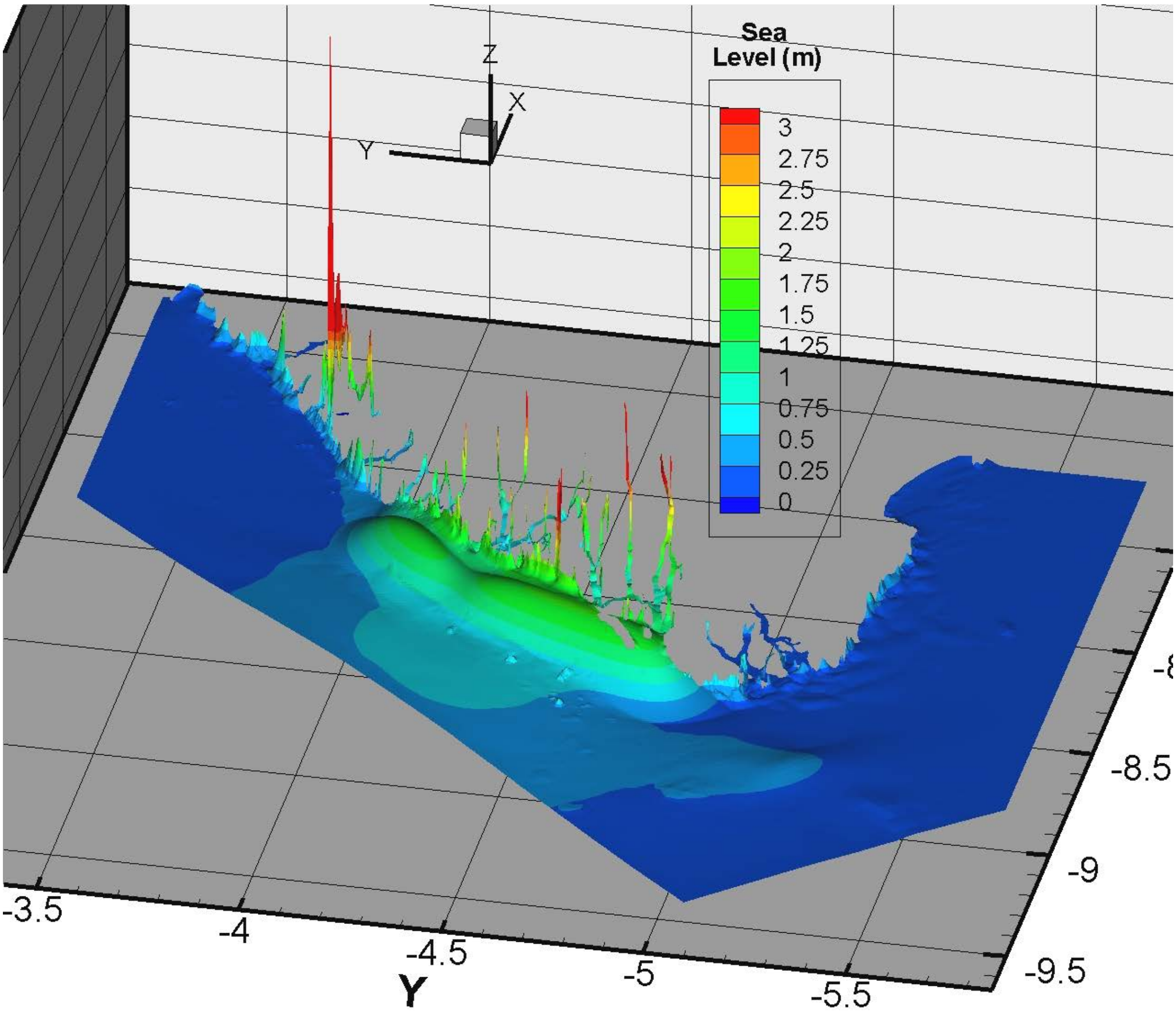


Figure 32 Envelope of maximum water levels during the Fiordland subduction zone earthquake with no surface rupture scenario.

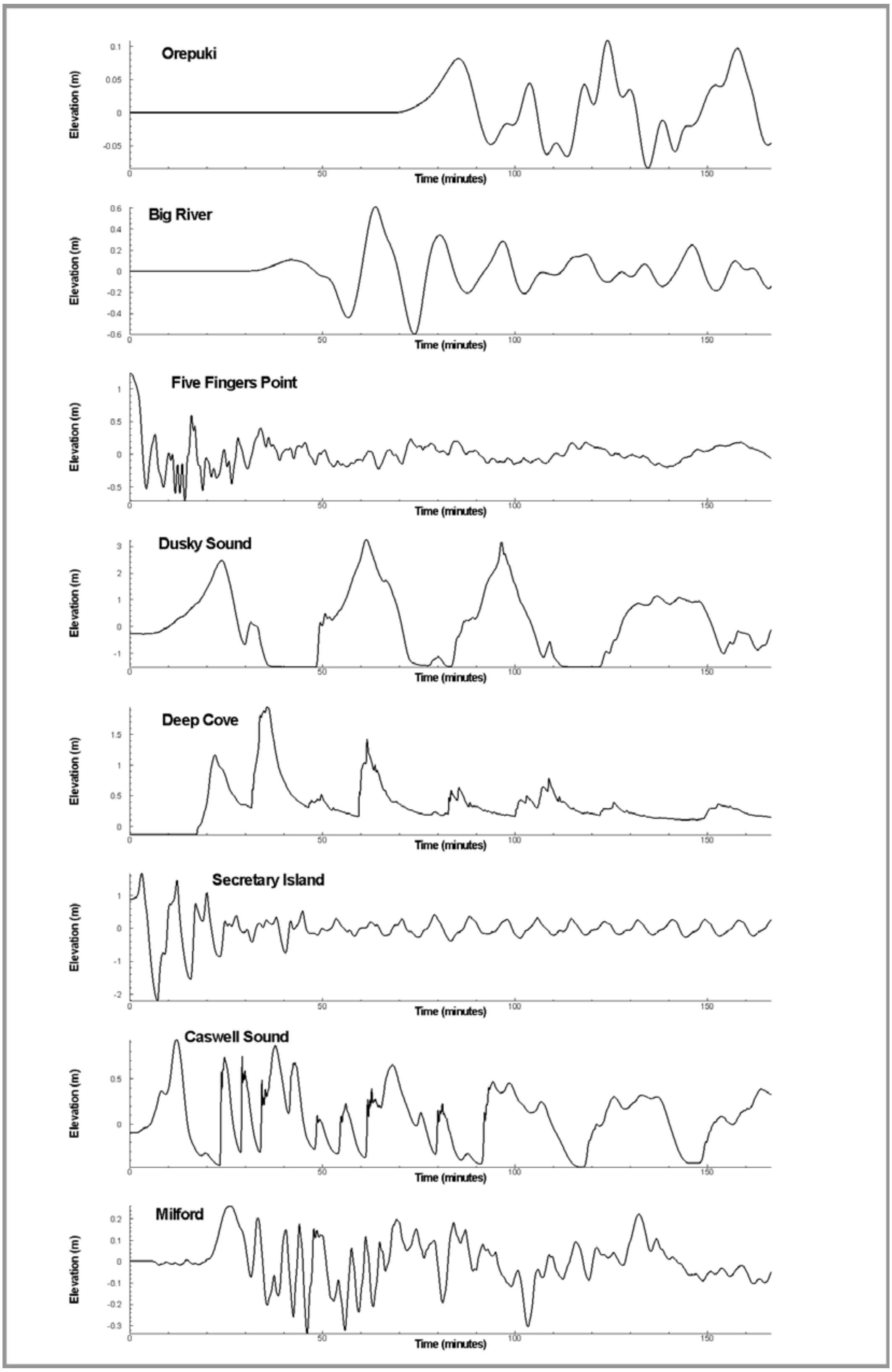


Figure 33 Time-series of water-surface elevation following the Fiordland subduction zone earthquake with no surface rupture scenario. The locations are shown in Figure 27.

5.4.3 Puysegur subduction zone events

There are two scenarios for the Puysegur subduction zone: one with surface rupture (Fig. 25) and one without surface rupture (Fig. 26). The former has the largest spatial extent, whereas the latter has a larger maximum sea-floor deformation.

In a similar manner to the Fiordland subduction zone events, the initial wave separates into two waves propagating in opposite directions. One wave propagates into the Tasman Sea, and the other towards the continental shelf south of Stewart Island.

The onshore/eastward propagating wave encounters the shelf area south of Stewart Island and the wave crest height above background sea-level increases dramatically. This appears to be a case of wave focusing where the incident wave refracts on a convex section of the shelf (see bathymetry on Fig. 27). The waves also refract on the eastern part of the shelf and propagate back toward Stewart Island and Bluff. Eventually, the wave passes westward along the southern edge of South Island. West of Puysegur Point, the wave undergoes refraction and diffraction and the amplitude is greatly reduced as it passes into the deeper Tasman Sea (Figs. 34 and 36).

In general, the wave crest height is 1 to 2 m above the background tide level as the wave passes Stewart Island and along the south coast of South Island. Locally, there are larger values in the range of 2 to 4 m. The greatest impact of both the Puysegur subduction zone scenarios occurs along the Southland coastline (including Stewart Island) with variable localised effects predicted for the Fiordland coastline. Time series of sea level after the two scenarios of fault rupture at various sites (Figure 27) is shown in Figs. 35 and 37.

In both scenarios (i.e. with and without surface rupture), waves inundate Halfmoon Bay, Stewart Island (see Figure 27) and Bluff (Figure 27) 135 and 165 minutes respectively after the earthquake

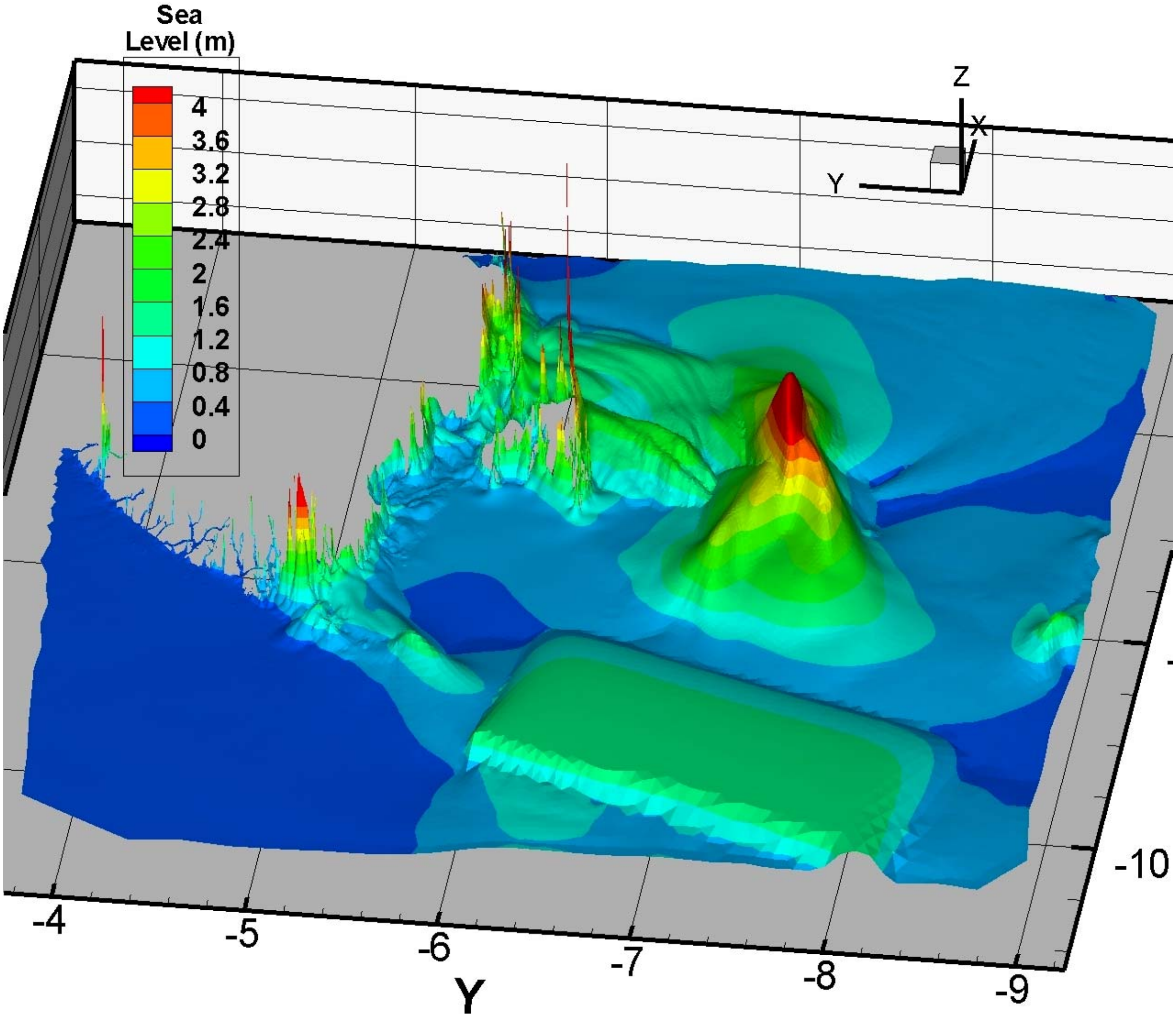


Figure 34 Envelope of maximum water levels during the Puysegur subduction zone earthquake with surface rupture scenario.

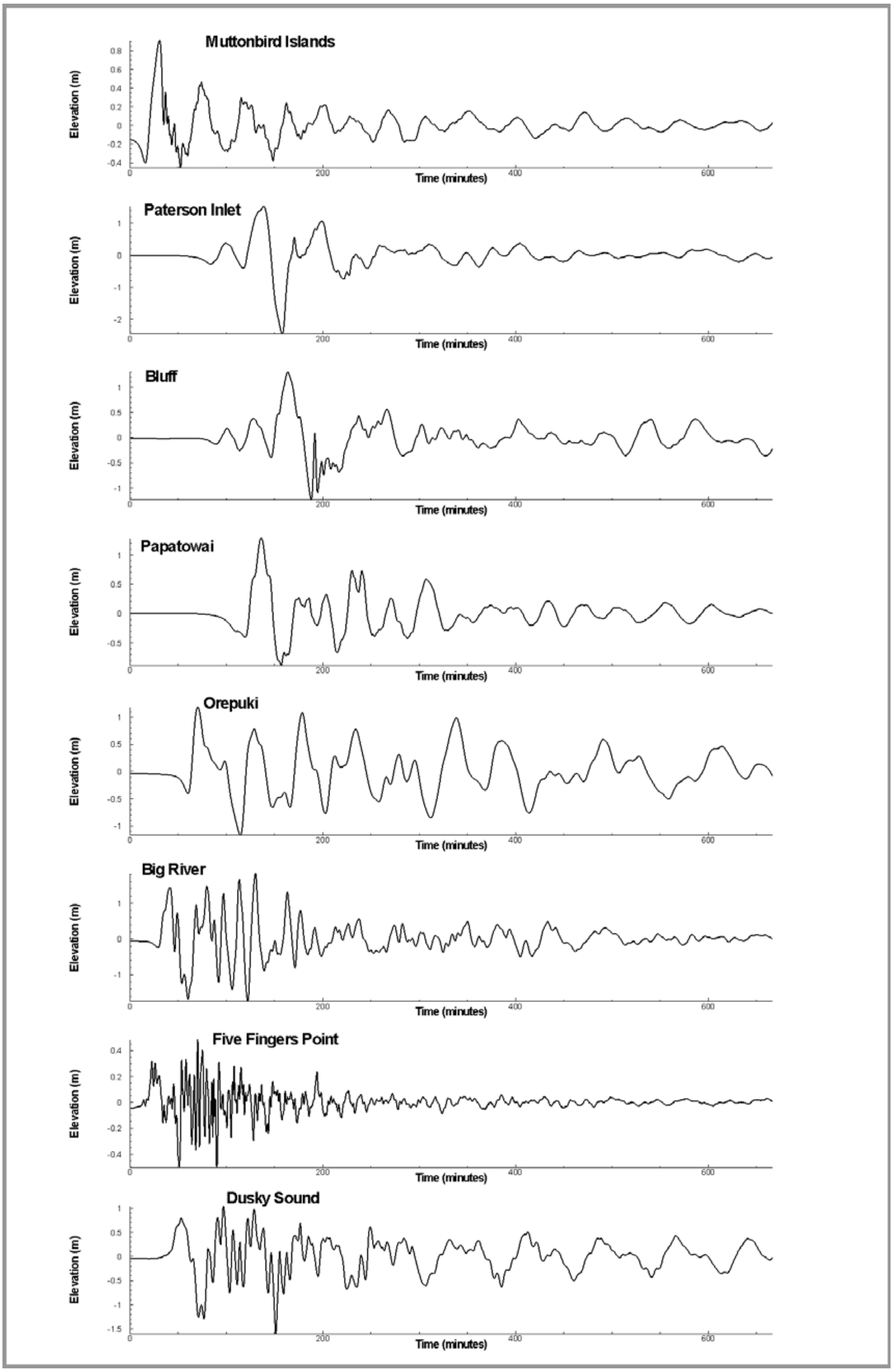


Figure 35 Time-series of water-surface elevation following the Puysegur subduction zone earthquake with surface rupture scenario. The locations are shown in Figure 27.

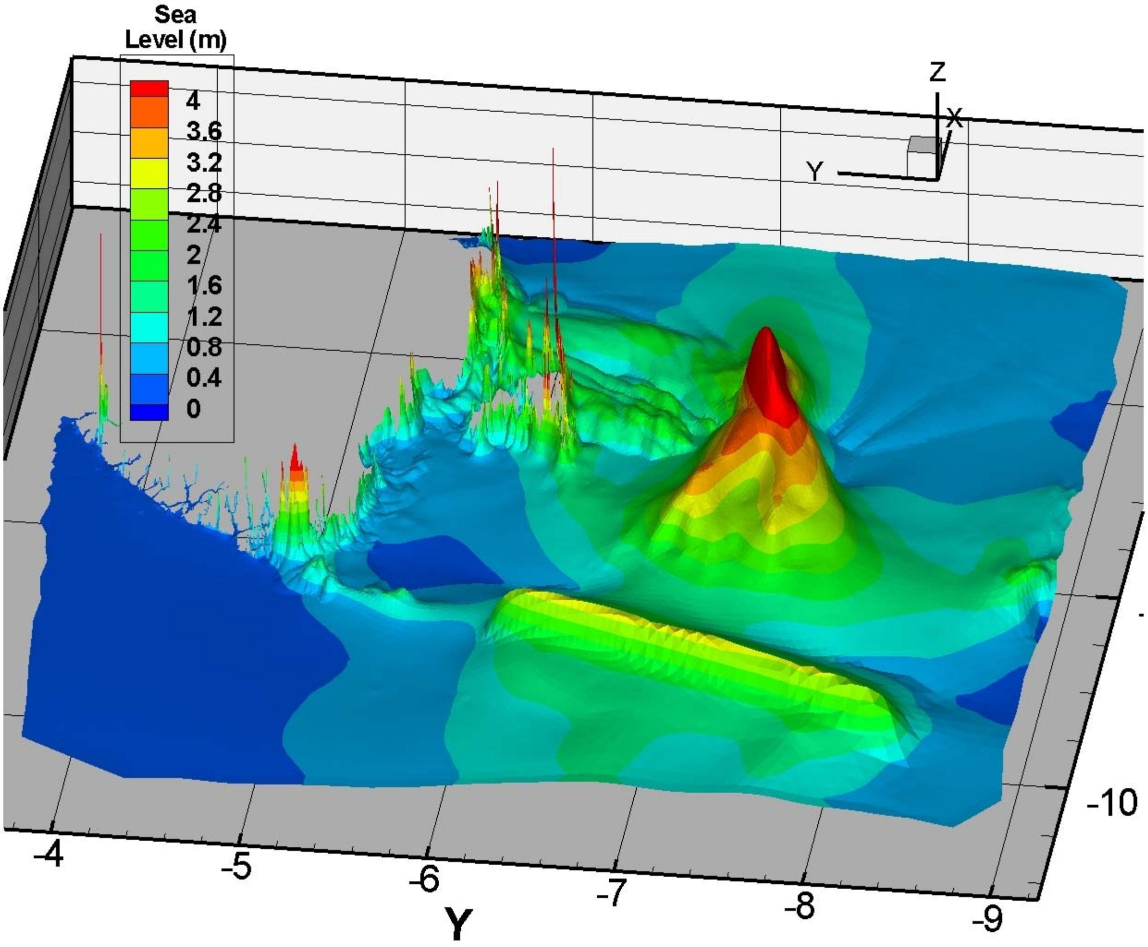


Figure 36 The envelope of maximum water levels during the Puysegur subduction zone earthquake with no surface rupture scenario.

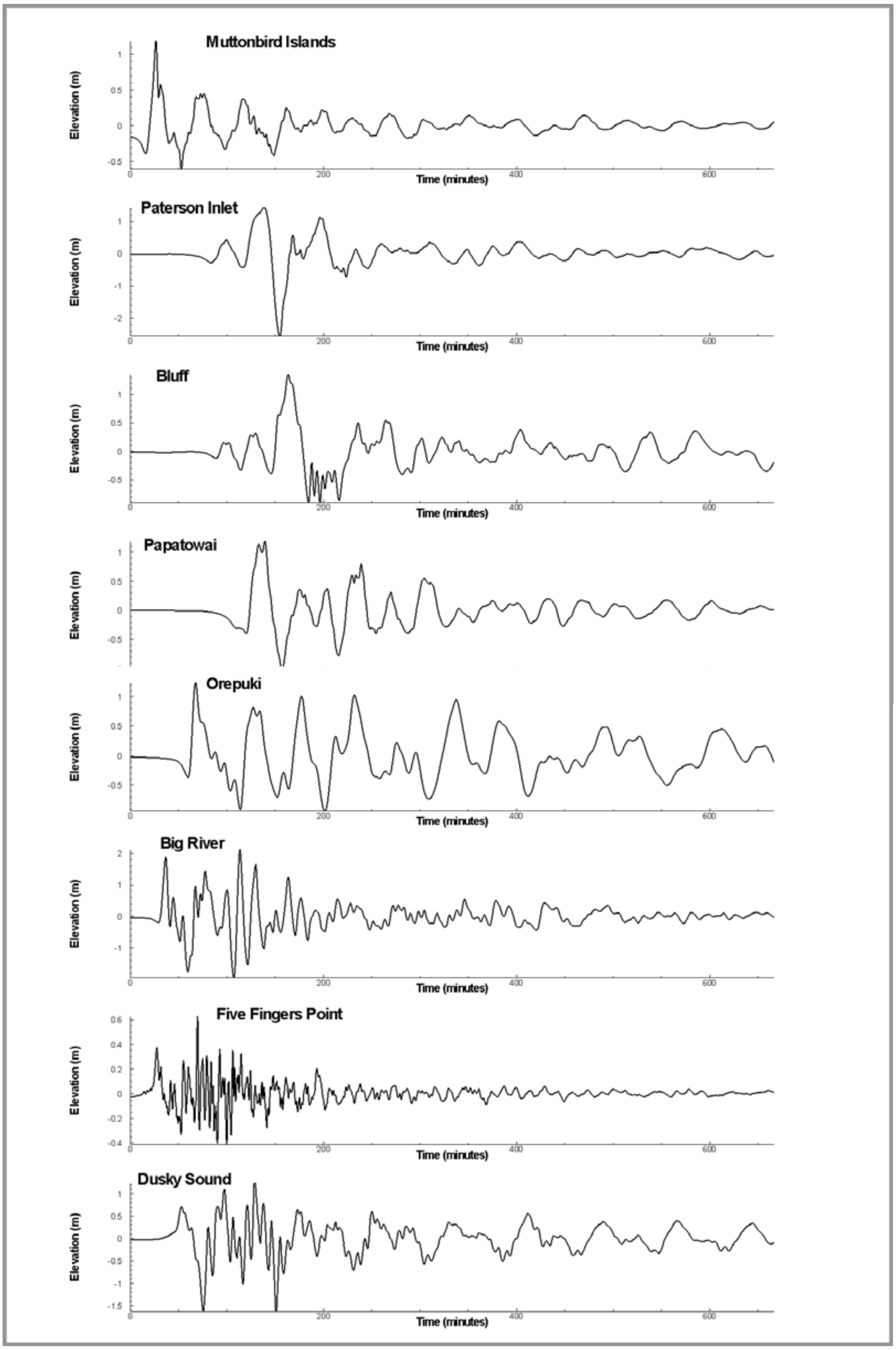


Figure 37 Time-series of water-surface elevation following the Puysegur subduction zone event with no surface rupture. The locations are shown in Figure 27.

5.5 Summary

All of the five tsunami generation scenarios are capable of creating waves with crest heights exceeding 4 m. In addition, the areas affected are local and waves do not propagate with significant amplitude along the coast away from the generation area. The highest water elevations for the Fiordland coastline are caused by the Fiordland subduction zone scenarios but the Alpine Fault scenario also produces locally high wave heights. The highest water elevations for the Southland coastline are produced by the Puysegur subduction zone scenarios.

Because of the initial displacement of ocean bottom and water surface, the initial wave separates into a wave propagating offshore and a wave propagating onshore. The Alpine Fault displacement acts like a point source for the resultant tsunami. The offshore wave dissipates rapidly in circular patterns and the onshore wave is experienced locally. On the other hand, the displacements for the subduction zone events act like line sources for the resultant tsunami. The offshore propagating wave dissipates from the ends of the wave, whereas the onshore propagating wave is more uniform in its effects.

The Puysegur subduction zone events are longer than the Fiordland subduction zone events thus the offshore propagating wave would have an effect at greater distance than that of the Fiordland events. The onshore-propagating wave increases in height as it shoals on the shelf and refracts around Stewart Island and eventually travels westward along the southern coast of the South Island.

A major limitation in the assessment of tsunami run-up height is that the available data for land and subsea topography lacks sufficient resolution to capture the spatial variability in run-up. The necessary resolution depends on the wave characteristics and land surface variability, and is generally of the order of tens of metres in the horizontal direction and a metre in the vertical. Nearshore bathymetric features that are commensurate with the tsunami wavelength can strongly affect the height of a tsunami as it approaches land, while inundation extent and potential damage are dependent on land topography and use (e.g. vegetation, assets). Lacking data for land topography, the shoreline was modelled as a vertical cliff. For much of Fiordland, this is a good approximation to the steep glaciated shore. At other locations such as at the valleys at the head of the fjords and at river deltas, more detailed data is needed to fully assess the run-up effects.

6.0 DISCUSSION

The period for which New Zealand has reliable written records of day-to-day events is very short – just 165 years. Even if the records of large earthquake and tsunami were complete for this period, it is not long enough to allow a reliable assessment of tsunami and earthquake hazards as the largest events may have return times of hundreds to thousands of years. Thus, the 1820s tsunami and 1826 earthquake, although they occurred before comprehensive records are available, with unreliable dates and without full details of their occurrence, nevertheless provide an insight into the larger events that might affect Southland and Fiordland.

6.1 1820s tsunami

Descriptive accounts of the 1820s event indicate wave heights that pose a significant hazard and risk to the Southland coast. Yet, the region has experienced no significant local- or regional-source tsunami since 1840, although it has had minor to moderate effects from tsunami from South America and minor effects from Southern Ocean sources (Balleny Islands, Macquarie Ridge) (Downes, unpub. GNS Tsunami Database). While it is uncertain when the "1820s" tsunami occurred, the descriptive accounts of the effects are plausible, and numerical modelling has shown that very large plate interface earthquakes in the Puysegur subduction zone could produce moderate wave heights at the shore of about 4 m along the parts of the coast compatible with the descriptive accounts.

The two source models developed for Puysegur subduction interface earthquakes equate to M_w 8.6 and M_w 8.5, and these are likely to be maximum possible events, that is, they extend along the entire fault zone. The potential of the zone to produce earthquakes of this size is unknown. It is possible that the zone ruptures in smaller earthquakes (and consequently smaller tsunami), and earthquakes of the size used in the models are either very rare or impossible. Alternatively, any redefinition of fault parameters, for example the dip of the interface, could result in realistic earthquakes with larger vertical displacement of the seafloor, and consequently larger tsunami. For example, a Puysegur subduction zone fault with a steeper dip will yield larger vertical displacements than an event with the same amount of slip on a shallowly dipping fault. In-depth seismological studies might be considered as a way of shedding more light on the earthquake potential of the zone. However, the Puysegur zone is so far offshore that only the larger earthquakes are identified and their locations, particularly their depths, are poorly constrained. Hence, the architecture of the zone is poorly defined (i.e. the dip of the fault at depth).

Sudden uplift or subsidence effects detectable in the geology of the coastal environment also provide information on past earthquakes and GPS can detect long-term deformation effects due to on-going locking of the plate interface. While the latter studies are possible along the Hikurangi margin, the Puysegur Trench is not amenable to such studies, as it is offshore and has no nearby coast that might be deformed. Further detailed mapping of the sea floor such as that of Barnes et al. (2005) is likely to provide better definitions of surface fault geometry and history.

Paleotsunami studies provide a relatively accessible option for better understanding the frequency and size of large earthquakes along the zone through their tsunami impact, and in this regard, the numerical modelling of tsunami can now be used to guide further selection of suitable sites for paleotsunami investigation. Figures 34 and 36 clearly show that the coast east of Bluff, Preservation Inlet, Dusky Sound and parts of Stewart Island have much greater wave heights than other parts of the coast. The modelling will also aid in the interpretation of the distribution of the tsunami deposits. However, the numerical modelling scenarios presented here are limited to the largest events and to the case where the slip (and hence, uplift of the seafloor) is the same along the length of the fault. Further modelling for other magnitude earthquakes and to evaluate the effect of variable uplift along the fault will be needed to provide the full range of possible tsunami impact scenarios. Not only will these scenarios inform the paleotsunami studies, but also they are needed by emergency management to both plan for, and respond appropriately to, future tsunami and the earthquakes that cause them.

Of further importance to emergency management is the result that the off-shore Alpine Fault and Fiordland subduction zone modelled earthquakes do not cause major damaging tsunami in Southland, and therefore were probably not the cause of the 1820s tsunami. They would, however, cause wave heights of perhaps 1-2 m that require a warning and appropriate response at near-shore facilities. As with the Puysegur scenarios, the earthquake dislocation scenarios are limited to the case where the slip is the same along the length of the fault. Use of other earthquake dislocation scenarios would be needed to confirm that maximum wave heights in Southland do not significantly exceed those from the scenarios already developed. Also it must be remembered that the modelled earthquakes present a hazard in themselves (refer to earthquake shaking hazard below).

The sources of the 1820s tsunami that have been investigated in this study are local to the southern part of New Zealand. Distant sources have not been considered. Tsunami from distant sources that pose the greatest threat to Southland are those from South America. As noted already, there is no known South American earthquake/tsunami event in the period 1800-1840 that could account for the 1820s tsunami in Southland. However, recent numerical modelling research into the effects of South American tsunami, has recognised that larger earthquakes, and consequently larger tsunami, than have occurred in historical times are possible (William Power, pers. comm., 2005). Further work is proceeding on this.

6.2 1826 earthquake and tsunami

In contrast to the somewhat vague descriptive accounts of the 1820s tsunami, the accounts of the 1826 earthquake and tsunami experienced by the sealers at Dusky Sound are much more definitive. The event seems real, and if not correct to the year, then correct to within relatively few years, that is, the era of sealing from 1792 onwards. The searching out of all potentially relevant material was beyond the scope of this study. However, it is felt that there is some scope for more information being found, perhaps in early Tasmanian newspapers, or by searching out early sketches, maps and descriptive accounts and comparing them with what was seen and described in the landscape, particularly landslides and uplift, after 1840 and with what is seen today.

As noted in section 3.0, mapping the distribution of landslides can give a good indication of MM intensity, and earthquake location and magnitude. Descriptive accounts of the 1826 earthquake suggest a much greater area of Fiordland was subjected to landsliding in that earthquake than in the 2003 Fiordland earthquake. Therefore a magnitude greater then M_w 7.2 can be inferred for the 1826 earthquake. Widespread forest destruction also occurs as a result of earthquake-induced landsliding and floodplain aggradation. Stripping of coastal forest can also occur in a tsunami. Dendrochronological studies of trees established after an episode of destruction can be used to date the earthquake or tsunami responsible. While such research has been carried out in South Westland and this shows strong correlation with onshore Alpine Fault events and correlation with an event in 1826, presumed to be the 1826 Fiordland earthquake, there has been little mapping of the distribution of landslides in Fiordland and as far as is known, no dating using dendrochronology. Thus, there is scope for these studies for confirming the approximate date and the magnitude of the 1826 event, and extent of any associated tsunami.

A Fiordland subduction zone source is favoured over an Alpine Fault source for the 1826 earthquake because the former would be more likely to cause coseismic uplift at the coast. Coseismic uplift is the inferred explanation for historical descriptions such as that of the cove named The Jail, “After these shocks the sea completely retired from the cove so that it ceased to be any longer a harbour, as the boat could only come to its mouth” (section 3.0). Although such changes could be brought about by deposition of a boulder bar (as currently exists at the mouth of Cascada Bay), uplift is the preferred explanation because of the existence of the flight of raised bedrock platforms observed in Cascada Bay (section 4.0). Dating of the platforms would be required before attributing uplift in Cascada Bay to deformation in the 1826 earthquake. However, assuming these platforms have been raised in previous large earthquakes, and possibly that platform B (Figures 13 and 14) was raised one metre in 1826, we can conclude that much greater earthquakes than the 2003 M_w 7.2 earthquake have occurred along this coastline in the past. Even the Fiordland subduction zone earthquakes of M_w 7.7 and 7.9 modelled in section 5.0 do not achieve one metre of uplift at Cascada Bay. The models are based on uniform slip along a credible length of fault for the Fiordland subduction zone but higher uplift at the coast could be achieved through slip on a deeper part of the interface, variable slip along the fault, or a greater amount of slip in a larger earthquake. It is also likely that other tectonic elements, not modelled in this work, have played a significant part in deformation of the Fiordland coast. For example raised platforms would be expected to recover (subside) interseismically if they were the result of purely elastic deformation in subduction zone earthquakes. Therefore, there possibly is an upper plate structure or another mechanism ensuring that uplift is permanent. If an upper plate fault exists in the region, it could rupture occasionally causing episodic uplift and abandonment of the raised platforms, in addition to the subduction zone events.

6.3 Earthquake shaking hazard

In this study, the prime objective has been to investigate the tsunami hazard for Southland in the first instance, but also for Fiordland. Very large earthquakes have been used as the source models for these events. It is useful to compare the magnitude of these events with the maximum magnitudes used in the National Seismic Hazard Model (Stirling et al. 2002). It should be noted that the Model is being revised (Stirling, pers. comm.) at present to incorporate the most recent seismological and fault data, including latest offshore fault data interpreted from swath bathymetry and seismic reflection profiles from Barnes et al. (NIWA).

At present, large earthquakes on the Puysegur subduction zone are not included in the National Seismic Hazard Model (Stirling et al. 2002). Based on the attenuation models of Dowrick & Rhoades (1999), we have predicted isoseismals for an M_w 8.6 Puysegur subduction interface earthquake (Fig. 38). These show that Southland is predominantly in the MM6 zone. This zone is a little more extensive than the expected effects of Fiordland subduction zone earthquakes of maximum magnitudes M_w 7.7 to 7.8, which are already included in the National Seismic Hazard Model (Stirling et al. 2002). Thus, although these earthquakes may not have been recognised in planning and engineering design, they have little potential to cause significantly greater earthquake shaking damage beyond what has been planned for already.

Large earthquakes on the offshore Alpine fault have been included in the National Seismic Hazard Model with maximum magnitude of M_w 7.2, while the Fiordland subduction zone has a maximum magnitude of M_w 7.8, both less than the scenario earthquakes we have used here. Stirling et al. (2002) note the lack of large historical earthquakes and paucity of paleoseismic data with which to constrain both the Alpine Fault and Fiordland subduction zone fault segmentation, recurrence times and magnitudes. Barnes et al.'s current offshore work will undoubtedly update this, as will further geological and seismological research. The predicted intensities for Alpine Fault (M_w 7.8) and Fiordland subduction zone (M_w 7.9) earthquake scenarios developed here are shown in Figure 38.

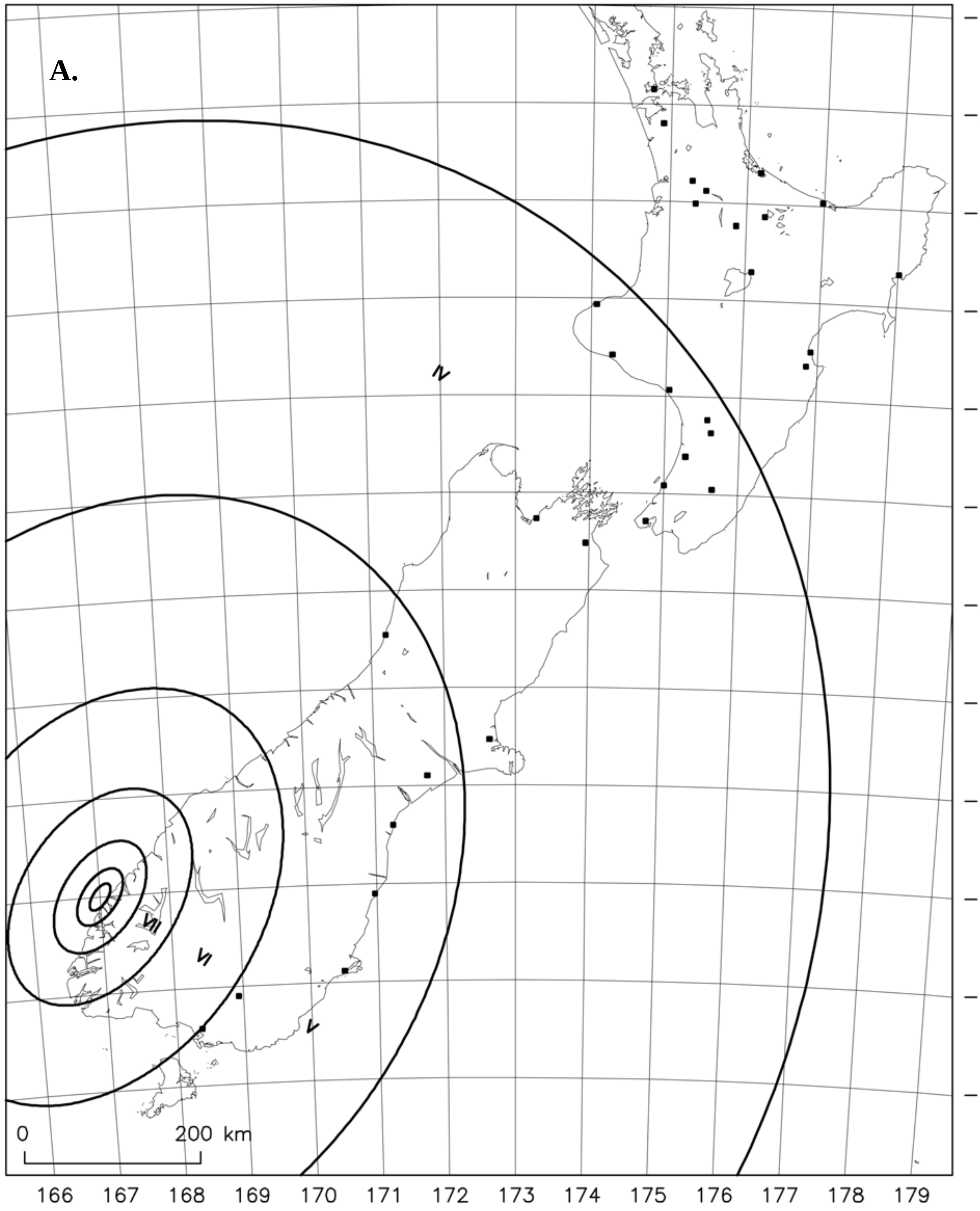
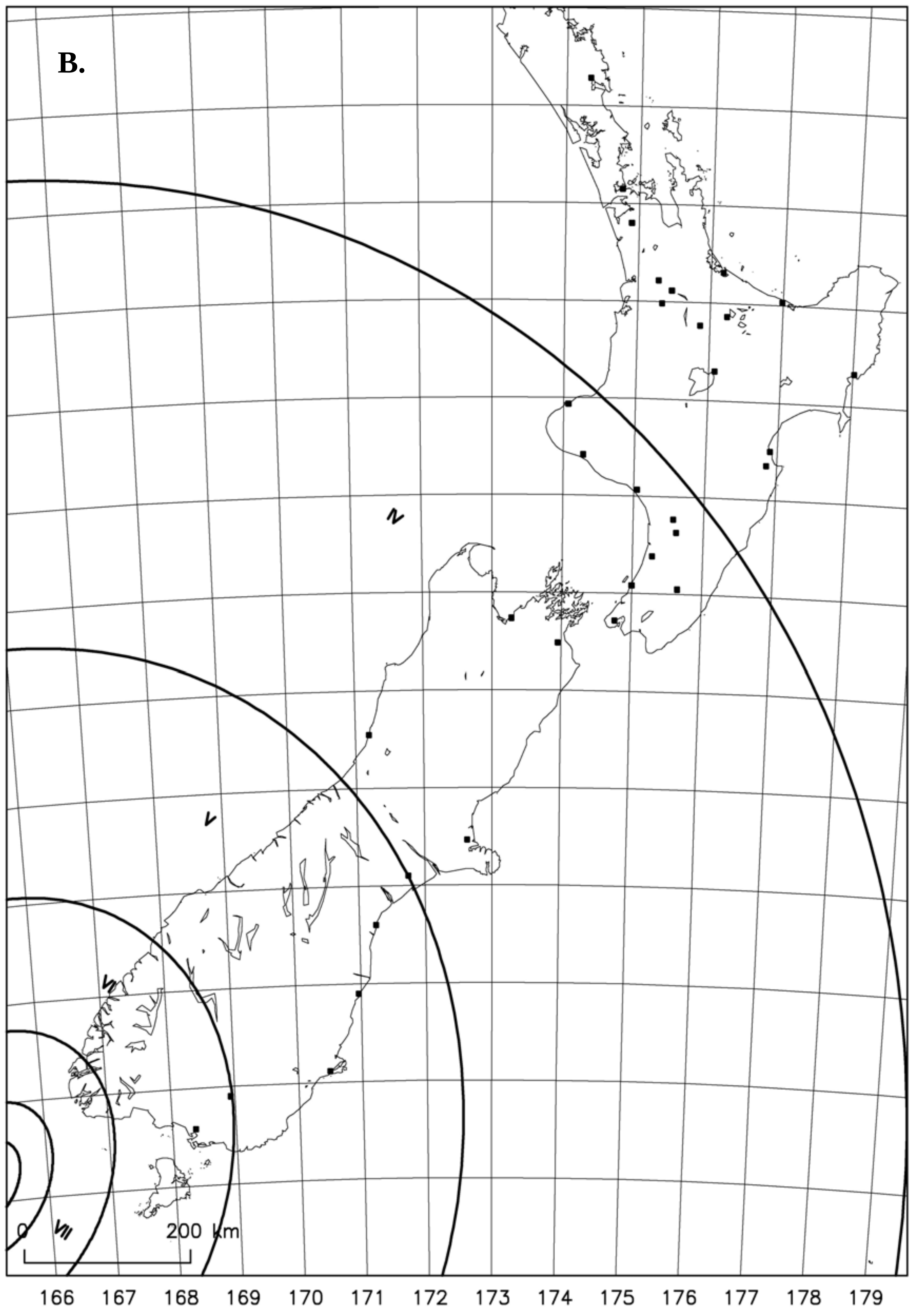


Figure 38 Expected intensities from A. an M_w 7.8 on the offshore Alpine Fault; B. an M_w 7.9 on the Fiordland subduction zone; C. an M_w 8.6 earthquake on the Puysegur subduction zone.


B.
−35
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166 167 168 169 170 171 172 173 174 175 176 177 178 179
0 200 km
IV
V
VI
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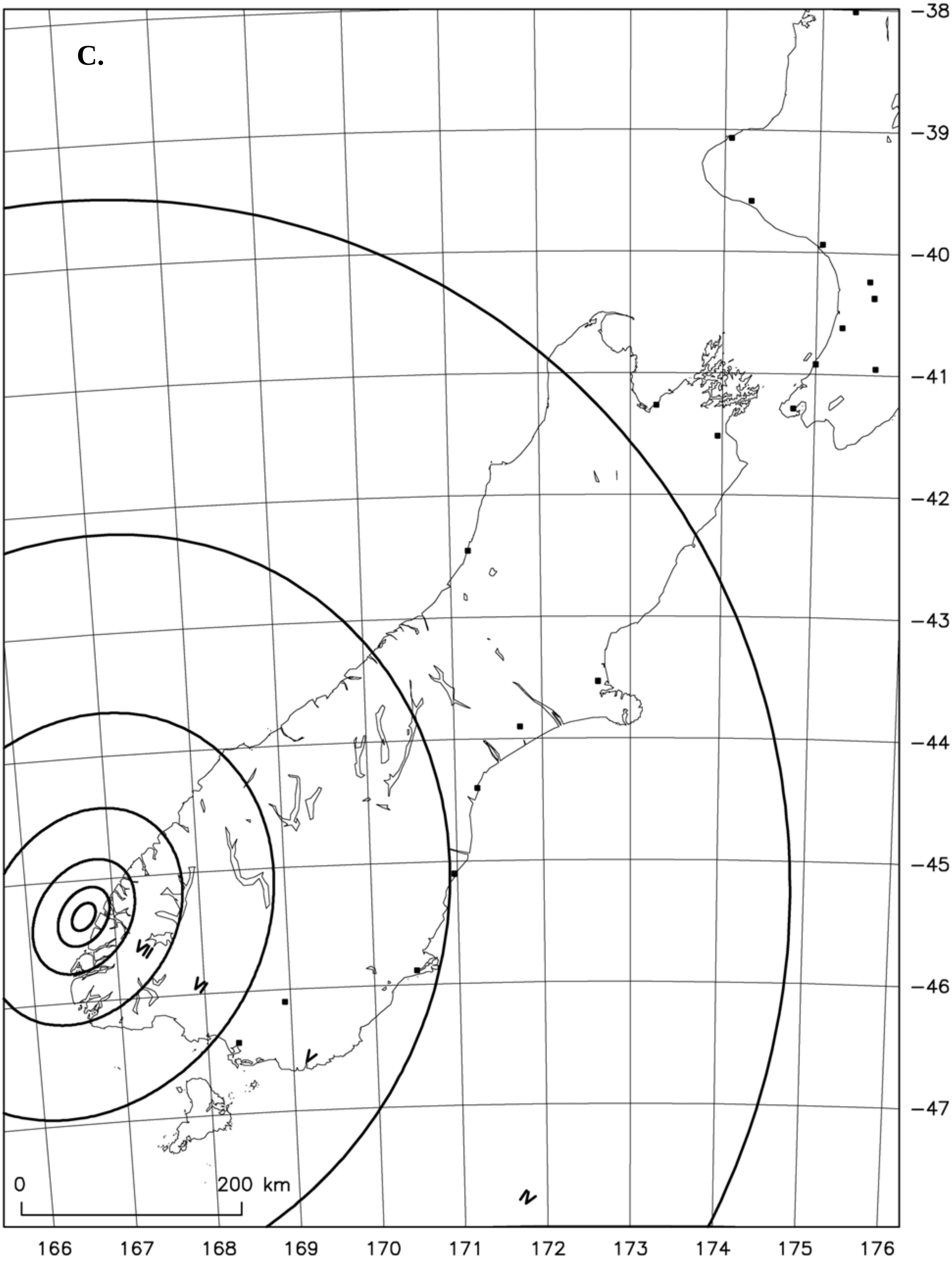

C.
−38
−39
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7.0 CONCLUSIONS

This project has been successful in achieving its goals, and there are some key results that will contribute to the understanding of tsunami, and to some extent, earthquake hazards in Southland and Fiordland.

In summary, the key results are:

7.1 Historical

The historical part of the project shows that there are more uncertainties in the historical data than previously recognised. The dates of both the 1820s tsunami and the 1826 earthquake are questionable. Because the 1826 earthquake was recorded by sealers and whalers, it probably occurred after their arrival in New Zealand waters, i.e. post-1792. The date of the 1820s event, however, is more uncertain. However, the descriptive accounts of both events are credible, and credible sources for each can be found.

The descriptive account of the 1826 earthquake, in particular the uplift, is most consistent with a Fiordland subduction interface earthquake, and credible large earthquakes on this fault, or indeed, on the offshore Alpine Fault, do not propagate large tsunami into Southland. Hence, it is unlikely the 1820s tsunami and the 1826 earthquake are related. However, out of the five potential sources of the 1820s tsunami investigated in this study, large earthquakes on the Puysegur subduction zone are capable of producing damaging tsunami in Southland consistent with descriptions of the 1820s tsunami.

7.2 Paleotsunami reconnaissance

The field reconnaissance part of the project indicates that despite the steep topography, high rainfall and high-energy wind and wave regime of Fiordland, there is potential for preserving physical evidence of past earthquakes and tsunami in the form of episodes of landsliding, forest destruction, and geomorphological changes to coastal waterbodies. Further work would be required to verify the findings listed below and to maximise the paleotsunami and paleoseismological records available at each site.

- Possible evidence for the occurrence of repeated large earthquakes in central Fiordland was found in the form of four raised platforms in Cascada Bay near the mouth of Doubtful Sound. Initial surveying implies single-event uplift amounts of at least one metre.
- Possible evidence for the occurrence of an historical tsunami was found in the form of very recent establishment of vegetation on a c. 8 m high surface in Cascada Bay, Doubtful Sound.
- An earthquake and / or tsunami at c.1450 AD may have been responsible for increasing the marine influence in the tidal inlet at the head of Goose Cove, Dusky Sound.

- Possible evidence of a much older tsunami (c. 6100 BC) in the form of a pebbly, sandy silt unit incorporating shells from diverse environments is preserved in the banks of the lower Hollyford River at Martins Bay.

7.3 Numerical Modelling

- The water elevations of a tsunami produced by an M_w 7.8 earthquake on the offshore part of the Alpine Fault is mostly very local to the uplifted / subsided areas produced by step-overs on the predominantly strike-slip fault, with minimal propagation of the tsunami to the south and north. The maximum shoreline water elevations at the outer Fiordland coast were over 4 m (at Five Fingers Peninsula), with mixed response within the fiords.
- The water elevations of a tsunami produced by a surface rupturing, M_w 7.9 earthquake on the Fiordland subduction interface centred beneath Doubtful Sound are more uniformly distributed than the Alpine Fault scenario, and are experienced over a larger area. Nevertheless, significant waves are mostly confined to the Fiordland coast with some propagation to the north and into parts of Southland. The maximum elevations at the outer Fiordland coast are close to 4 m (Secretary Island), with mixed response within the fiords. Fome fiords experience shoreline water elevations greater than 4 m. Along the Southland coast, maximum shoreline water elevations approach 1 m at Big River (in western Southland, Figure 27), but are less than 1 m at Orepuki.
- The water elevations of a tsunami produced by a non-surface rupturing M_w 7.7 earthquake on the Fiordland subduction interface centred beneath Doubtful Sound are similar in extent to the previous (surface rupturing) scenario with somewhat smaller, but still significant, maximum shoreline water elevations (less than 3 m along the open coast, about 3 m within some fiords).
- The surface rupturing M_w 8.6 and non-surface rupturing M_w 8.5 earthquakes on the subduction interface along the Puysegur subduction zone south of Fiordland produce more significant tsunami along the Southland coast, Stewart Island, and around Puysegur Point than the previous scenarios. In general, the shoreline water elevations are 1 to 2 m in these areas, with some locations experiencing 2 to 4 m. Notable effects occur to the east of Bluff, where diffraction of the tsunami around Stewart Island causes higher waves than at locations closer to the fault.

8.0 ACKNOWLEDGEMENTS

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We thank Mauri McSaveney (GNS) and Scott Stephens (NIWA) for the careful reviewing of this report. We also thank Penny Murray and Carolyn Hume for their word processing and graphics input.

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10.0 APPENDICES

Appendix 1 Source Material: 1820s Southland tsunami

Source: MacIntosh, J. 1985: From waste land to wealth, the history of Otara, Southland. Invercargill [NZ]: Otara Centennial Committee.

p1 [In Introductory Chapter: Origins of a name]

> In “Saga of the Pioneers” a story describing Otara as “quite historical for it was on the Otara beach that there was a large Māori pa which was attacked by the northern Māoris. While they were indulging in a great cannibal feast, a huge tidal wave came up from the sea and swept everything within radius away After the great storm which plainly showed the anger of the Great Spirit, the beach lay bare and all the looted treasures of greenstone, battle axes and ornaments were covered with sand washed up by the seas. Many of the Māories were engulphed and lost their lives. It was a strange fact that at this same period at the beginning of last century (1800s) this great tidal wave also swept over the Orepuki beach and a large party of Māories who were journeying to the Waiau mouth to obtain food for the winter were exterminated.”
>
> Note: This story, sometimes attributed to folk lore, was told long before so many Māori artefacts were found in the Otara area.

Source: Reece, A. G. (ed.) 1984. Pahia 1820-1985 Invercargill, NZ: Craig Printing co ltd.

p15-17

> Pahia is named after Tahu Pahi, or Pahi as he was usually called, chief of the pa which was known to the early sealers and whalers as Pahees, the most westerly of the settlements on the northern shore of Foveaux Strait. This settlement was often used as a source of provisions for those in need.
>
> The ruins were still visible to W. B. D. Mantell in 1852, when he made his first visit to the area, in a walk from Dunedin to the Waiau River, taking a census and setting out Māori reserves.
>
> …
>
> Pahia was originally a Maori pa. In the early days it was at Otatara, Colac and pahia or the Aropaki district that the natives of the western district congregated. In and around Pahia are many evidences of the old days when a powerful Maori race lived in this district. Farmers when ploughing, found battle axes and skulls and greenstone weapons, and it must have been a period just before the landing of the whalers and sealers that the great war between the northern and southern Maories , which almost exterminated the powerful tribe, took place. Following this terrible war was the catastrophe of the great tidal wave which overwhelmed the whole tribe while they were travelling along the beach from Orepuki to the Waiau Beach. It was at Kaitangata Point that this tragedy occurred. The place is most suitably named for Kaitangata in Maori means ‘eat a tribe’.

Source: Richards, Rhys, 1995. "Murihiku" re-viewed – a revised history of Southern New Zealand from 1804 to1844. Wellington [NZ]: Lithographic Services.

p38

> In June 1823 when Captain Kent visited Ruapuke in the Mermaid, he noted that most of the local people were away gathering muttonbirds from the southern islands. He visited Te Wera's village at the Bluff which was very small and almost deserted, and Port William which was then the base for some sealers (including Thoms) under Captain Day in the Sydney brig Wellington. Kent also mentioned the drowning of the chief 'Pihee' and about forty others in June 1823 when their double canoe ['unua'] broke up while returning from the muttonbird islands.

p39

> Further west, the largest and oldest village was Pahi's kaika, though it was much reduced in numbers as forty of its people had drowned in 1823 [storm, not tsunami, see above].

Source: Saunders, M., 1934: The conquerors, saga of the stations. *In:* Southland Times, 17 February 1934.

> Added to the terrible war was the great catastrophe of the great tidal wave which overwhelmed the whole tribe while they were travelling on the beach from Colac to Waiau Beach. It was at Kaitangata Point that this tragedy occurred.

Source: Saunders, M., ca. 1936. *In:* Southland Daily News (date uncertain).

> According to Maori tradition, 1820 or thereabouts, was a year of great disturbances, earthquakes in the north and tidal waves in the south … the Maories of Murihiku, both on the west and south-east coasts affirm that a great tidal wave swept over the area where Orepuki now is, and all along the western coast, and thereby many hundreds of the Ngati Mamoe tribe met their deaths by drowning. It appears that it was the usual custom for tribes to meet together at the Waiau mouth in the autumn to provide themselves with enough fish to carry them over the long winter…. They were traveling along the beach track when great tidal waves rose from the ocean, and carried them out to sea, and it is believed no survivors remained to tell the story; a similar adventure is also recorded from an area beyond Fortrose.

Source: Smith, M. A. 2003. From Goldfield to Fields of Green. Waimahaka, Southland, NZ: M.A. Smith

p. vi (Introduction)

> The land at one time formed part of a great Maori settlement with a large pa at Pahia, under the chief Pahi. Another village was situated further along the bay which later bore the name of its chief, Te Wae Wae. The people of the two villages were closely related and friendly and a fair degree of social interaction and trade took place between them. Walking from Pahi's village to Te Wae Wae's

took about 4 hours and as they travelled westward towards the valley of flax, rising hills and the tops of mountains, they drank etutu, a tea made from the tutu plant, to refresh them on the way.

Before crossing a rapid freshwater river (the Waimeamea) in the canoe that was kept there for that purpose, the party would travel along the sandy beaches, to the projecting headland named Kaitangata (meaning eat a man), where legend had it that retreating enemies were once trapped by the tide and eaten by their pursuers.

About 1811, a large party from Pahi's village and Colac Bay were making their way westward to gather food for the winter, when they were overwhelmed by a huge tidal wave which dashed them all against Kaitangata's cliff. Not one of them survived this disaster, and this incident took its toll on the Maori population of the area. Incidentally Kaitangata today, situated just west of Monkey island, is much diminished to what it was then, having been washed by many tides over the years since it is now hardly noticeable as a feature on the beach.

Notes:

The book is about the settlement of Orepuki, Southland. Source of the information above is not referenced, but according to the author (telephone conversation) the source was an article by L. Esler in the Southland Times in the last ten years. L. Esler has been contacted but has no further information on the source.

Source Material: Possible 12-14th century tsunami

Source: Beattie, H. 1915. Traditions and Legends collected from the natives of Murihiku. (Southland, New Zealand). Part II. *Journal of the Polynesian Society* 24, 98-112 .

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THE TAKITIMU CANOE

If *Arai-te-uru* is *the* canoe of the natives at Moeraki and Waikouaiti, then Takitimu is certainly *the* canoe of the Murihiku people. The canoe is said to have been wrecked in Foveaux Strait and the takitimu mountains were named to keep the canoe in memory, but this prosaic explanation has no vogue amongst the southern natives. They assert that this great range of Takitimu is the veritable canoe turned into stone, and I did not risk my life by contradicting them. There are two versions of how this happened, and both are very interesting specimens of folk-lore.

The first runs that the Takitimu canoe had run down the east coast till just below the Otago peninsula, when she ran off a great wave which the legends avers is represented by the Mauka-atua (now called Maungatua) range. The canoe ran off this sea and broached-to and dropped her *tata* (bailer) which turned into rock, and now is the Hokanui hill near Gore. Then the other wave (represented by the Okaka ridge west of the Waiau river) struck her and she upset, and there she lies as the Takitimu Mountains. When the first wave struck her one of the crew named Aonui was washed overboard, and being turned into stone, still stands on the Tokomairaro beach as the tall basaltic pillar known to the white man as Cook's head. One of my informants said that Aonui was swept off the Arai-te-uru, but the others all said it was off the Takitimu.

The other version of the story relates that at that time Southland was under the sea, the Bluff hill being an island, and the sea laving the foot of the Hokanui Hills. The canoe had come around the coast and was somewhere near where Gore is when disaster overtook her. The first wave struck her unexpectedly, and she dropped her bailer right on the spot, and then two more seas completed the disaster. The old song says: [In Māori]

Ko te tipaka mai ano Takitimu
Ko te Poroporo huariki
Ka tae mai ki te kutuawa Waimeha
Ka makere te tata
Na ka karu (? karo)
Nau O-te-wao, nau Oroko, Nau Okaka,
Koe tukituki —e —e!

My informant knew that these three waves were now ranges, but could not define the localities with any certainty.

Notes: See Beattie (1917) below for a translation and further comment on the song.

Source: Beattie, H. 1915. Traditions and Legends collected from the natives of Murihiku. (Southland, New Zealand). Part II. *Journal of the Polynesian Society* 24: 130-139.

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TWO TRADITIONS

There is spot below the seacliffs at Orepuki (rightly Aropaki, on Foveaux Strait) called Kaitakata (N. I. = Kaitangata), and this is so called not because of man-eating man, but because the sea there rose in its might and devoured some people in its capacious maw. A large party of Te-Mano-o-te-Rapuwai were travelling along the beach from the Waiau river when a great wave (Tai-koko) swept them away, and the name Kaitakata commemorates the catastrophe.

Notes: Te Rapuwai was one of the earliest South Island tribes (McFadgen, pers. comm.). In this regard, Beattie writes in the same paper:

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The Rapuwai people increased rapidly for a time and became so numerous that they are often spoken of as "Te-Mano-o-te-Rapuwai" (the great number of Rapuwai). One of the districts they most densely populated is about Lake Kaitangata, in the Otago Province, and that district is sometimes called in consequence "Te-Mano-o-te-Rapuwai".

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THE COMING OF KAI-TAHU [NGAI-TAHU]

By the time the Rapuwai, Waitaha, and Kati-Mamoe had become almost one people by intermarriage, we can safely say that there had elapsed about 300 years from the Great Migration, as it must have been about the year 1650 that the Kai-Tahu invaded Te Wai-Pounamu (or South Island).

Source: Beattie, H. 1917. Traditions and Legends collected from the natives of Murihiku. (Southland, New Zealand). Part VI. *Journal of the Polynesian Society*, 26: 75-86.

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In the song as published (Journal Polynesian Society Vol. XXIV., page 109 [see above]) it is questioned if *karu* should not be *karo. Karu* is the southern rendering of *ngaru* (wave), and that line would be read in the northern form as "*Na nga ngaru*".

The song was roughly translated to me as follows:

> "With regard to the broaching of Takitimu,
> She came from the North Island.
> She arrived at the mouth of the Waimea Stream
> And dropped the bailer.
> By the waves known as
> O-te-wao, Oroko and Okaka
> She was utterly destroyed. Alas!

These three waves are now represented by ridges. Okaka is "The Hump" at Waiau Rive, Oroko (Orokoroko) is the southern portion of the Hokanui Hills, while O-te-wao is "a ridge up Oreti river way".

Source: Downes, T.W. 1914. History of Ngati-Kahu-Ngunu (Chapter I, part 2). *Journal of the Polynesian Society* 23: 11-125.

p111

[After travelling from Cook Strait] He [Tamatea] landed near the place where Lyttelton now stands, and after cooking his fish he climbed a hill, which he called Tama-totea, to see the lay of the country. Apparently he was disappointed with the view, for he said to his men, "This place won't do, we must go further on". So on they went, and when they lay opposite Waiau between the White Cliff and Wai-tangi, his tohunga Rua-wharo and Te Tu-rongo-pa-tahi (Uenuku's grandson) said, "This place will do, we will turn in here." So the Taki-tumu [the canoe] was turned, but as they paddled in to the shore, lo, she stuck fast on a sand bank, and the united efforts of all of the crew could not get her off. When Tamatea found himself in this plight he became very angry, and cried out to Rua-wharo, “What do you mean by this, have you brought me here to drown me?” Rua-wharo also grew angry, and standing up he called to the sea to rise to his help. Immediately a great tidal wave came to his assistance; the canoe was lifted right into the river, where they all held her fast, while the scour back of the great wave made that river the deepest in Aotea-roa. There the Taki-tumu went ashore. And there she still lies, the proud canoe that had braved storm and tide, that had carried the hardy navigators all the way from Hawaiki to Aotea-roa – truly a record to be proud of.

Notes: The location in this account seems to be Waiau, Southland, as assumed by Bruce McFadgen (pers. comm.). This is supported by an early map of Fiordland drawn by Edward Shortland (redrawn on p31 of Hall-Jones, 1984) showing an area to the west of Te Wae-Wae Bay call “White Cliffs”.

Source: Stack, J.W. 1877. Sketch of the traditional history of the South Island Maoris. *Transactions of the New Zealand Institute*, 10: 57-92.

p 61

[Heading] UNCERTAIN TRADITIONS [of the Te Rapuwai]

Te Rapuwai or Nga ai tanga a te Puhirere succeeded the Kahui Tipua and rapidly spread themselves over the greater part of the [South] island. They left traces of their occupation in the shell heaps found both along the coast and far inland. It was in their time that the country around Invercargill is said to have to have been submerged, the forests of Canterbury and Otago destroyed by fire, and the moa exterminated.

Notes: In his article, Stack estimates that the Waitaha occupied the Southern part of NZ from mid-15th (approx. 1477) to mid-16th century, and that the Rapuwai were a pre-Waitaha people.

Appendix 2 Source material: 1826 Fiordland earthquake and tsunami

Source: Taylor, R., 1855. *Te Ika a Maui or New Zealand and its inhabitants*. Wertheim and MacIntosh, London, England. 499pp.
Keywords:
p235

> It appears that from 1826 to 1827 there was an almost constant succession of earthquakes, some of which were sufficiently violent to throw men down. At times he [a person formerly engaged in sealing in Dusky Bay] and his party, who then resided on a small island, were so alarmed lest it should be submerged, that they put out to sea; there, however, they found no safety, for such was the flux and reflux of the ocean, that they were in the greatest danger of being swamped, and were thankful to get on shore again. The sealers were accustomed to visit a small cove called the jail, which was a most suitable place for anchorage, being well sheltered with lofty cliffs on every side; and having deep water in it close to the shore , so that they could step out on to the rocks from their boats. It is situated about 80 miles to the north of Dusky Bay. After the earthquakes the locality was so completely altered; the sea had so entirely retired from the cove, that it was dry land. Beyond Cascade Point the whole coast presented a most shattered appearance, so much so that its former state could scarcely be recognized. Large masses of the mountains had fallen, and in many places the trees might be seen, under water.

Source: Taylor, R., ca. 1854. Notebooks, including drafts of *Te Ika a Maui or New Zealand and its inhabitants*. Auckland Public Library, GNZMSS297/27 Taylor.

p13-14 [The section starting with page 12 is headed Meurant, this name obviously having been added at a later date in a different pen.]

> In 1826 the greatest shocks were felt, they were sufficiently severe as to throw a man down. They continued five or six months to the beginning of 1827, seldom 2 days elapsed without one occurring. ~~sometimes~~ The narrator who was engaged in sealing there in Dusky Bay stated that some times they were so alarmed lest the island sh^{d} [should] sink that they used to push off to sea & there the flux and reflux of the sea was so violent that they were afraid of being swamped at sea & had to return again. [The narrator also stated] That there was a small cove which was shut in on every side with high land and thence named the Jail in which they were accustomed to anchor their sealboat as there was deep water in it and they could jump out of their boat on [to] the rocks. After these shocks, the sea completely retired from the cove so that it ceased to be any longer a harbour, as the boat could only come to its mouth.
>
> This [information] was furnished by Mr Meurant the native interpreter who has been many y[ear]rs in the middle island Oct 5 1847.
>
> The jail cove is in Grono's Harbour about 80 miles N of Dusky Bay. Beyond Cascade Point the

whole coast presented a most shattered appearance. After the shocks large masses of the mountains fell, so that in many places trees might be seen under the water. [Here is inserted a sketch of the location of The Jail.]

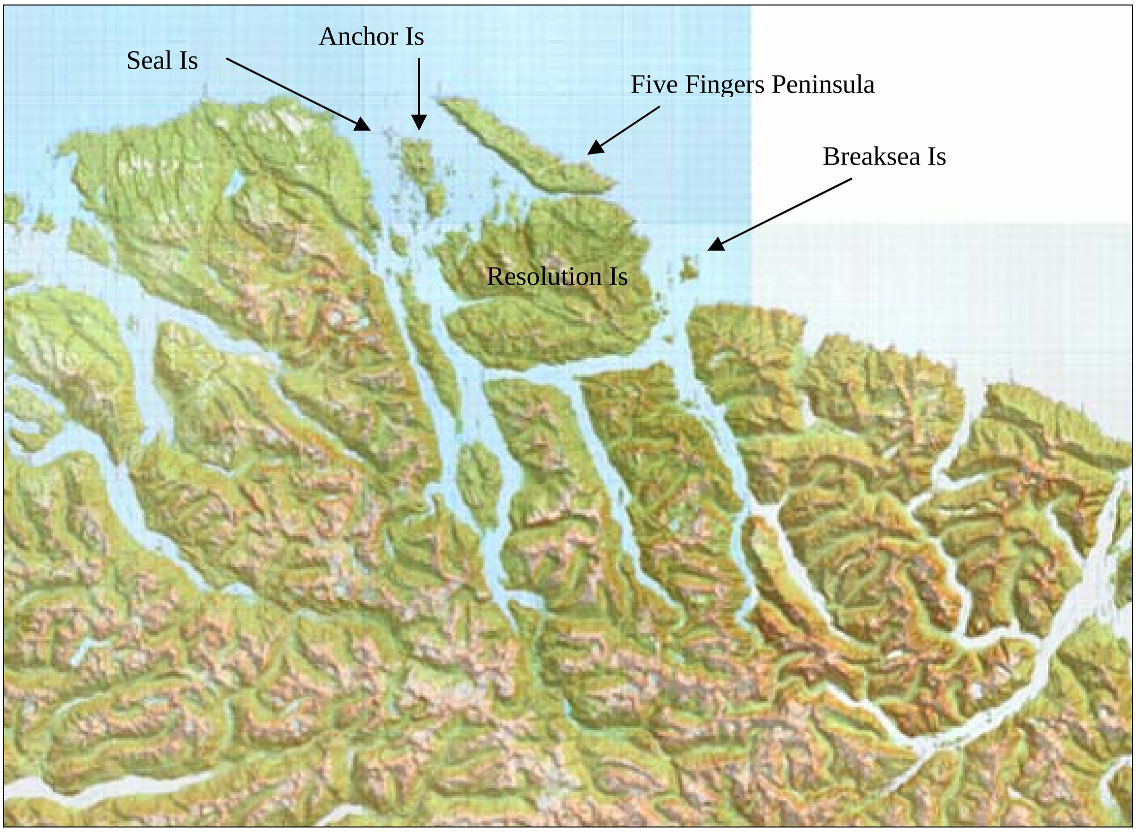

Figure A1 Map of features of Dusky Sound relevant to Taylor's (ca. 1854) diagram below.

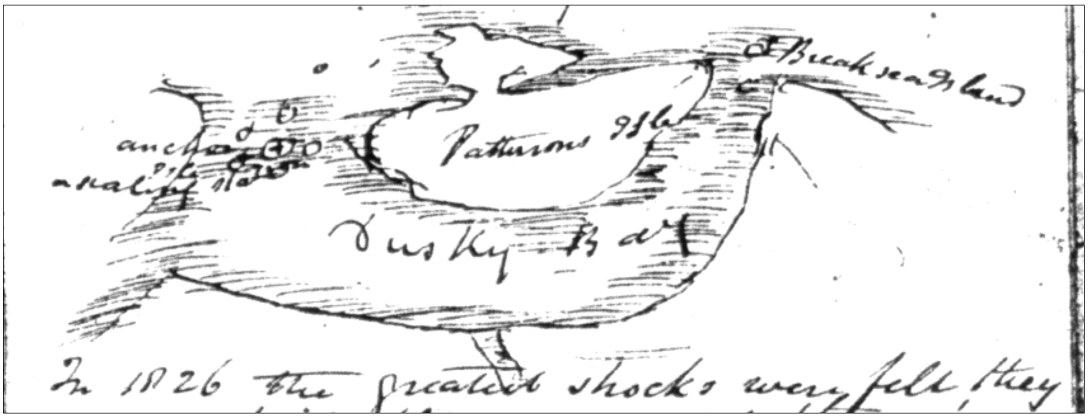

Figure A2 Sketch map of Dusky Sound from Taylor (ca. 1854) showing the general area where the sealers were stationed when the earthquakes occurred.

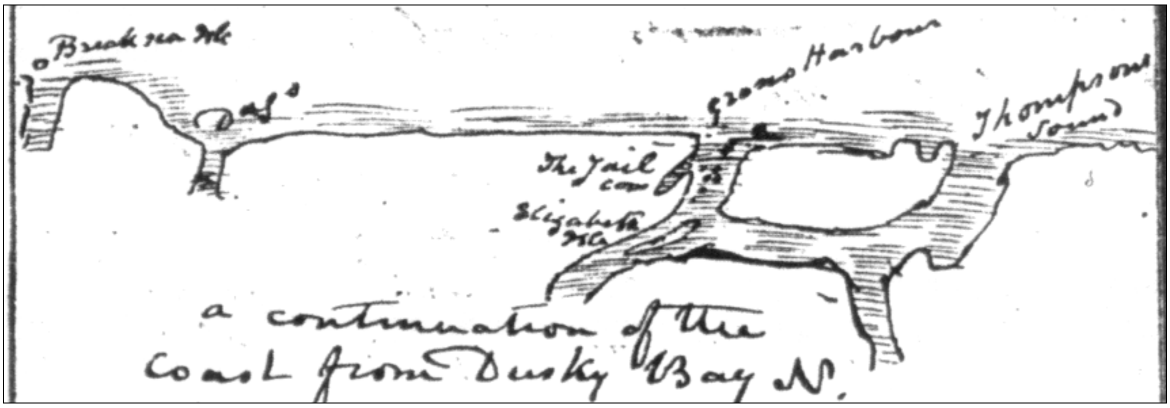

Figure A3 Sketch map from Taylor (ca. 1854) showing the location of "The Jail" within Grono's Harbour, which is now known as Doubtful Sound.

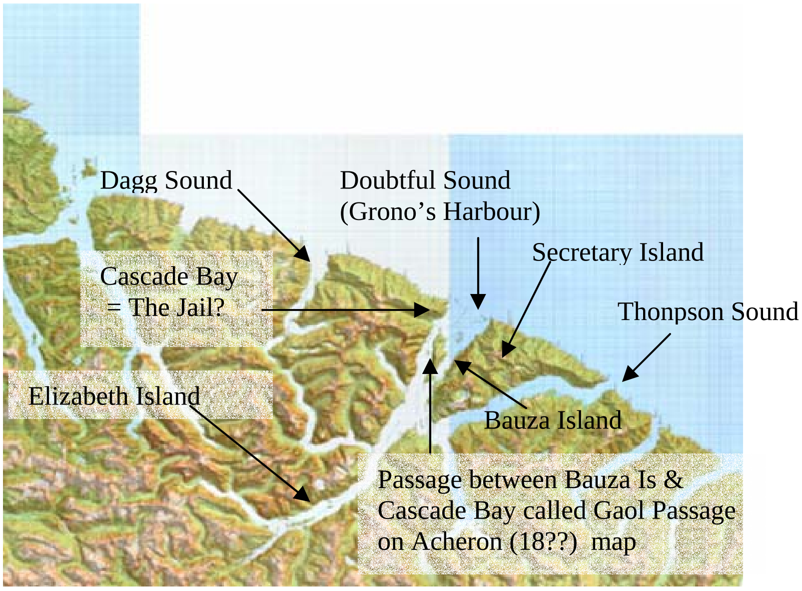

Figure A4 Map of features around Doubtful Sound relevant to Taylor (ca. 1854) diagram above.

Notes: The above document was found and transcribed by Mervyn Rodgers in 2002. Rodgers also supplied photocopies of the originals, his transcription and notes. One of the authors of this report was able to interpret more text and this is included in the transcription. Rodgers notes that the spelling and lettering is left as they appear in the diary and that his [insertions] are so marked. Where a word had been crossed out it is so indicated.

Source: Carrick, R. (Compiler & editor) 2005. Historical records of New Zealand prior to 1840. Christchurch, NZ: Cadstonbury Publications. (A reprint of Carrick (1903), with some introductory notes).

CHALKY INLET

Captain Edwardson, of the cutter Snapper, largely engaged attending seal gangs in this inlet—the upper reach of which is still known as Edwardsons Sound—communicated the following extraordinary coincidence to the Collector of Customs (Sydney) :—During 1813-1814 he had his first gang working the head of the bay. There was then a river of large volume, with a fall (40 or 50 feet) to the sound, opposite the beach on which his party camped. The length of the river was from two to two and a-half miles, fed by a lake; his men reported having visited several times, in quest of eel, ducks, etc. After a few months' residence, the party withdrew, and, again, after four years' absence, when the rookeries were expected to be replenished, the place was revisited. No one is known to have been there in the interim; the few camp appliances left on the ground remaining intact. The river, in the meantime, had entirely disappeared, and no new channel or outlet was found. The sound at the upper end terminated in a lagoon, which had all the appearance of silting up. Between the sound and the lagoon there was a narrow, intricate passage through which the tide rushed at high-water, and the surplus water discharged itself at low tide. It was not sufficient for the passage of a small boat ; being little more than a mere crack. Having spent the best part of two days on the lagoon shooting, Captain Edwardson could speak positively as to the change which had come over it on his return in 1820. it was so great that he was at first disposed to doubt the evidence of his own eyes. It was then opened out into a large lagoon-harbour, with a deep-water entrance channel, large enough for the accommodation of a whale ship. Instead of one head-stream as formerly, two good-sized rivers now fed the lagoon, so that the aspect of affairs had completely changed. Captain Edwardson's idea was the lake had broken out in an entirely new channel, running in amongst the hills, and that, instead of discharging its waters as formerly into the sound, it now found an outlet to the lagoon, and that was what had produced the second river or head-stream. Even then, the opening out of the entrance channel to the lagoon, which, from being tortuous, had become almost straight, remains unaccounted for. Evidently the convulsion which wrought so much havoc amongst the seal islands must have been severely felt at this place.—Patriot (Hobart), December, 1820.

Appendix 3 Site selection notes: Fiordland paleotsunami reconnaissance

The following site descriptions are based on information provided by Ian Turnbull, Rupert Sutherland and Mauri McSaveney, a report on the vegetation of Fiordland beaches by Johnson (1979), and examination of aerial photos. Sites are rated according to their apparent potential for preserving tsunami deposits – low, moderate, high or X (not appropriate for this study). Other factors such as tectonic setting and relevance of site to tsunami hazard considerations will also need to be addressed when choosing sites.

Site Descriptions	Site Ratings
Ruapuke Island (D48, E48)	X
Coastal lagoons exist on western and south-western shores. Turnbull – Permission for access could be time consuming, any fieldwork would have to be planned for the summer of 04/05 rather than 03/04. Mo knows of one thesis on the basement rocks (Webster 1981) but otherwise little geological work has been done.	Access too difficult
Steward Island (C49, D49)	Low
Turnbull – The western coast is either steep or sand-dominated beaches. The southern end of Mason Bay may be a possibility.	
South Coast between Preservation Inlet & Te Waewae Bay:	
Sutherland – Any events found along the southern Fiordland coast would be directly relevant to Southland – where high human interest lies in comparison with central Fiordland. It is also likely to be directly hit by a tsunami travelling from the Puysegur Subduction Zone. This coast is rocky and terraced for much of its length but Rupert suggests some of the small stream gullies could have good spots for tsunami deposit preservation.	See sections of coast listed below
Puysegur Point to Big River Mouth (B46)	X
Aerial photos (3996/4-6) – most of coast consists of high cliffs and wide intertidal rock platforms, there are a few sandy beaches backed by cliffs and occasional low-lying areas at the mouths of streams but not much potential for preservation of deposits.	
Johnson (pp.93-97) – Johnson has information on Sealers Creeks, Kiwi Burn and Green Islets and none of these sites look suitable for our purposes.	
Lake Hakapoua / Big River Mouth (B46)	Moderate
Turnbull – Originally a tidal fiord, the entrance was dammed by a landslide in ca. 1915. (See also Bishop (1986) Puysegur geological map – specifically Fig. 17).	
Johnson (pp. 98-100) – Big River drops only a few m over the kilometre between the lake outlet and sea (so it may be partially estuarine). Sandy beaches with low dunes occupy both riverbanks near the mouth.	
Oblique aerial photo in the National Park book p. 80-81 shows the small pond and wetland on the west bank of the river mouth. If this had a peaty sedimentary sequence it could be useful but it may be a very recent feature – even ponded in debris from the 1915 landslide.	
Aerial photos (3996/7-8) – Low-lying land with wetland on west bank doesn't appear to be part of same landslide blocking entrance but probably on recent colluvium. May contain a recent record. NB: smaller scale photos (1251/1-2) no good because of cloud cover.	
Knife & Steel Harbour (B46)	Low
Aerial photos (1251/6-7) – Two small streams are ponded behind a sandy beach here. There is little low-lying land but some potential for deposits in the back-beach stream channel area.	
Waitutu River Mouth (C46)	Moderate

Johnson (pp. 101-102) – on the west side of the river there is a sand and gravel beach, a low dune (4 m above high tide) and a low coastal terrace.	
Aerial photos (1252/7-9) – There is a sandy lowland on the west side of the river. It has several terrace risers cut into it parallel to the river and a small coastal cliff on the seaward side.	
Long Point to Sand Hill Point (C46)	Moderate-low
Aerial photos (1252/11-27) – Most of this section of coast has high cliffs. However there are a couple of low-lying sites that may be useful for recent deposits. There is a break in the cliff and a small triangular delta area at the mouth of Crombie Stream. There is a low terrace between Francis Burn and Sand Hill Point. The eastern half of this is dominated by sand dunes but the western end and mouth of Edwin Burn may be possibilities.	
Preservation Inlet:	
Turnbull – Much of the coast is very steep in this area. QMAP will be there with a boat and helicopter support around March 2004.	See sites listed below
Spit Beach, Preservation Inlet (B45)	Low
Turnbull – A wide sand flat backed by dunes, wide open to the SW. There are probably peaty areas between the dunes, although most of these are forested.	
Johnson (pp. 88-89) – Te Whara Beach (immediately west of Spit Islands) has a large dune and dune ridges to 15 m high, it appears to be too sand-dominated for our purposes. However the beach immediately east of Spit Islands may be less sand-dominated.	
Aerial photos (3995/4-5) – The eastern beach appears to have low-lying forested land behind the dunes, possibly colluvium but could be worth a visit.	
Seek Cove, Preservation Inlet (B45)	Low
Johnson (p.87) – Across the isthmus from South Port, this has a gravely beach with sand near water level and in small pockets near the forest edge. There is a gravel and driftwood levee near the back of the beach.	
Aerial photos (3995/3-4) – Appears to have a fairly abrupt cliff at the back of the beach.	
South Port, Chalky Inlet (B45)	Moderate
Turnbull – South Port is in fairly sheltered water but has some swampy ground between beach ridges that may preserve evidence of marine inundation.	
Johnson (p. 87) – This is a gravely beach very exposed to northwest winds. Land rises gradually behind the beach.	
Aerial photos (3995/3-4) – There don't appear to be any distinct beach ridges but forested back-beach area looks like a possibility.	
North Port, Chalky Inlet (B45)	Moderate
Johnson (p.85) – Between Breaker Point and Fisherman Bay, at the mouth of Shallow Creek **(A45)**, there is a sheltered cove with a small area of salt marsh.	
Aerial photos (3995/4-5) – Scale too large to gain much insight.	
Lake Cove, Edwardson Sound (B45)	X
Turnbull – Lake Cove consists of a sandy to muddy intertidal area at the head of the cove and a very swampy, boggy area at the mouth of the stream.	Too far from open sea for tsunami?
Cape Providence (A45)	Moderate
Turnbull – The coast north of this point has a raised platform that would be worth looking at. It is beyond the reach of high tide and storm surges so may preserve deposits at its landward edge. Peat sections may exist at the mouths of several small streams that cross the platform along this piece of coast.	
Aerial photos (3995/1-2) – Mouths of creeks west of Lake Hector could be possibilities.	
Goose Cove, Resolution Island (A44)	Moderate-High
Turnbull – Goose Cove faces south. The inner half is an estuary with a gravelly to sandy barrier at its southern end and a fan-delta barrier between it and Woodhen Cove at its northern end. The northern barrier could be overtopped in a very large event whereas the southern barrier is partially covered at high tide, and	

overtopped in strong SW winds and swells. Extensive mud flats are exposed at low tide. Storms probably cause re-suspension and deposition of sediment in the estuary so differentiation between storms and tsunami would be required for interpretation of any record of influx events. It is likely that the southern barrier is a source of sediment that a tsunami could mobilise and shift far into the mud flats, while a storm would probably only create gravelly to sandy washover deposits a short distance into the mudflats. This would be a good place to use QMAP's rubber dingy to get around – at high tide.	
Aerial photos (2025/3-4) – Steep-sided, extensive estuary, hard to tell how much sand in the system but definitely has potential.	
Woodhen Cove, Resolution Island (A44)	Moderate-Low
Turnbull – Woodhen Cove faces north and has a large build-up of driftwood behind its gravel beach. There is a swampy back-beach area but there are large pieces of driftwood in the mud and it's likely that the driftwood dam and beach would trap any incoming sediment before it reached the back-beach area.	
Aerial photos (2025/3-4) – Very steep-sided, narrow channel. Some potential at head.	
Duck Cove, Resolution Island (B44)	Low
Duck Cove gradually shallows to a sandy intertidal area at its head. It may be too sheltered from the open sea but is a possibility	
Supper Cove, Dusky Sound (B44)	Low
Turnbull – Large intertidal flats strewn with logs, and cut by numerous channels exist at the head of Supper Cove and there is an area of peat deposition near the mouth of Henry Burn. However the area is likely to be dominated by fluvial deposits and parts of the record eroded by flood events because of the size of the Seaforth River. It is also a long distance from the sea.	
Disappointment Cove, Resolution Island (B44)	Moderate-Low
This cove faces northwest and is sheltered by a number of islands in the entrance to Breaksea Sound.	
Johnson (p. 83) – This has a gravel beach with little sand and no dunes. The beach is backed by a storm ridge with lots of driftwood and a terrace. In places sand runs under the open mountain beech forest.	
Coal River Mouth (B43)	Moderate
Turnbull – There is a barrier at the mouth of Coal River which is breached at the northern end and the river is tidal for a fair way inland. Dune-fields exist on the southern side of the river with pockets of peat swamp nestled among them. These sequences of peat may record events that overtopped the barrier or flooded the valley. The river is large but much of its bedload would be filtered out by several lakes that exist upstream. This is Mo's number one site on the list.	
Johnson (pp.77-82) – Much of the mouth of the valley is mapped as dunes or alluvial fans. There are moist hollows within the dunes but these have "been eroded down to a gritty and stony base near the water table" so don't appear to contain sequences of peat. Profiles indicate that the dunes in the centre of the bay are at least 10 m high. Low ground occupied by swampy forest exists between older and younger dune systems and here "a stream has deposited alluvium and re-worked the dunes sands". This area occurs immediately south of the river so would be exposed to inundation of the river by the sea. However it would also be prone to flooding from the stream and river.	
Air photos (3990/1-2) – These confirm that the low ground described by Johnson would be the only possibility for investigation here.	
See also oblique aerial photo in National Park Book (p.41).	
Precipice Cove (C42)	Low
Turnbull – There is a large swampy area at the head of Precipice Cove. However it is a long way inland and may be too sheltered from the open sea.	
McSaveney – wave travel is likely to be impeded in the fiords because most of them have shallow lips near their entrances and long distances to their heads (the heads being where useful sedimentary records are most often preserved).	

Emelius Arm (C42)	Low
Turnbull – The head of the Emelius Arm also has swampy areas and is closer to the open sea. However the Irene River that flows into this arm is large and probably deposits and erodes sediment from the swampy areas in its lower reaches.	
Two Thumb Bay (B41, C41)	Moderate-Low
Turnbull – Rocks exist at the mouth of Two Thumb Bay which protect it slightly. A small lagoon at the river mouth is likely to be an ephemeral feature without a long sedimentary record. The valley is aggrading and has a high input of sediment from the steep catchment.	
Johnson (pp. 74-76) –The head of the bay has a small headland that divides the area into two beaches. No sand exists on the beach and there are no dunes. Both the beach and river bed are dominated by gravel. There is a small stony platform behind the beach on the south side of the river.	
Aerial photos (3973/ 1-2) – There is only a narrow entrance here because of a high-point / knob in the middle of the bay. Flat land behind the beach and south of the stream would be a possibility.	
Looking Glass Bay (B41, C41)	Moderate-Low
Turnbull – This bay is similar to Two Thumb Bay. The catchment and valley of Looking Glass Bay are also very steep. The stream regularly switches channels across the valley floor.	
Johnson (pp. 72-73) – Looking Glass Bay has a gravel beach and river bed and no dunes. There is a stony alluvial flat at the southern end that has 40 cm of silty soil on top of it. Damage to bay-head rata trees is thought to be a result of deer removing scrub fringe (also at Two Thumbs Bay).	
Aerial photos (3973/1-2) – Land behind the north end of the beach rises very steeply. To the south there is a small area of flat land between two river channels. However this is probably the alluvial flat described by Johnson – 40cm of silt on top of gravel and fairly regularly flooded.	
Catseye Bay (B41, C41)	Moderate
Turnbull – This has a sandy barrier at the mouth of the river and is otherwise fairly similar to the previous two bays.	
Johnson (pp. 65-71) – The river here is tidal for at least the lower 2 km. There are several dune ridges near the sea and these are lower than elsewhere (~10 m). Older dune systems may have been buried or eroded. There are damp hollows between the modern dunes but no mention is made of peat accumulation. Soil horizons become more distinct on older dunes so there may be some soil / sand stratigraphy preserved. Salt marsh also occurs about 1 km up the river but closer to the sea than many other sites.	
Aerial photos (3985/1-2) – Much of the valley floor appears to consist of slips off the valley sides so the only possibilities are those described by Johnson – pockets of salt marsh adjacent to the river and hollows within the modern dune sequence.	
Sutherland Sound (B41, C41, Pt. C40)	Moderate-High
Turnbull – Sutherland Sound could be a good site to obtain a long sedimentary sequence because it is more sheltered than most of the sounds by barriers near its mouth. It is still fairly deep so would require a deep water drilling system to sample from the floor of the Sound.	
Johnson (pp. 59-64) – An estuary separates the seaward part of this fiord from the inner brackish water 'lake'. At low tide about 92 ha of tidal flat are exposed. Salt marsh lines the edges of these tidal flats in the upper reaches of the estuary (2-3 km inland from the open sea). On the south side of the fiord there is an older, flat sand surface (just over 1 km inland from the open sea) that may be an uplifted estuarine flat or an old dune surface in which the hollows have been filled with peat. There is up to 1.7 m of peat overlying sand in one of the hollows here.	
Aerial photos (3985/3-6) – The sand flat area described above extends out into the valley so would be exposed to surges coming in from the sea but inundation would depend to a great extent on the state of the tide.	

Poison Bay (D40)	Moderate
Turnbull – This bay also has a sandy beach (grading to gravel at the south end) backed by dunes overgrown by scrubby forest,and may be another option with similar properties as Catseye, Two Thumb and Looking Glass bays.	
Johnson (pp. 57-58) – The beach at the head of the bay is 5 km from the open sea. It is sandy at the southern end with the rest composed of cobbles and boulders. There is a low (6-7 m above low tide), forest and scrub covered foredune.	
Aerial photos (3984/1-2) – A dune system may extend further back into the valley but it is all relatively flat and forested.	
Transit River Mouth (D40 & Pt C40)	Moderate
Turnbull – Garnet sands form the beach and dunes at Transit River Mouth and behind these are large areas of swamp. It would be worth checking the height of the beach/dunes at this site and if they're not too high, this could be a useful (and very scenic) site. The sandy beach grades into gravel at the south end, and the dunes are lower there. The 6000 yr paleoshoreline was about 2.5 km inland from the present coast, so any deposits within the Transit swamp will be younger than this.	No aerial photos?
Johnson (pp. 53-56) – Immediately south of the river mouth the beach (100 m wide) is backed by young hummocky dunes (40 m wide). Further south there is a steep foredune backed by forested hummocky dunes. Dunes are well over 10 m high on the profile 200 m south of the river.	
Milford Sound (D40 & Pt C40)	Moderate-Low
Sutherland – Milford Sound would be one of Rupert's highest priorities because of the risk associated with tsunami hazard and the tourism ventures at Milford. This site would have the greatest human interest for a tsunami study in all of Fiordland. He admits a wave may be greatly reduced in impact by the time it reaches the head of the sound but locally generated landslide tsunami would be of interest. Rupert suggests there would be several areas suitable for coring and we should try somewhere around Deepwater Basin and a short distance up the Arthur River.	
McSaveney – Landslide material near the mouth of the Arthur River is purported to be from an earthquake in the 1820s so any ponds in the area would be too recent for our use. Deepwater Basin doesn't contain a recent sediment record around its margins. The Cleddau River delta would be quite active so the best possibility would be the Harrison River delta.	
Aerial photos (3983/6-7) – Confirm that the river deltas would be the only possibilities.	
Yates Point to Madagascar Beach	Low
Johnson (pp. 49-50) – This coast is mostly rocky with bouldery platforms and bouldery or gravely beaches. There is a coarse sandy beach and low foredune at the northern end of Madagascar Beach and a raised bouldery platform. This has a number of dead trees at the forest margin.	
Aerial photos (SN 581/14-24 = Yates Point to Ruby Beach) – The coast is predominantly steep but there are a few flat back-beach, scrub-covered platforms at the mouths of streams between Yates Point and Madagascar Beach.	
Martins Bay (D39)	High
Turnbull – On the south side of the Hollyford River in Martins Bay there is a series of back-beach swamps and lagoons lying parallel to the beach between old beach ridges. There is potential to sample along a shore-normal transect and obtain records of older events further inland. Mo would put this as No 2 on the list, or even equal to Coal River. Lake McKerrow was probably an open fiord at the 6000 yr high sea level (with marine fossils at the head of the lake) so these dune-lagoon areas will post-date that. It would be wise to get flown in to the south side of the river because the Hollyford is a bit big to cross here. There is a good photo of the Martins Bay dune complexes in "The fold of the land" by Homer & Molloy – p.8. See also the National Park book p. 97.	

Johnson (pp. 41-46) – The beach is the biggest in Fiordland, very close to the open sea and very exposed to wind. The sandspit at the northern end has 6-8 m high dunes, the centre of the bay has a 16 m high dune and the southern end is much flatter with smaller dunes into which driftwood and seaweed are carried a long way inland. McKenzie Lagoon is brackish with wetland areas around the margins. There are also large areas of bog in the hollows between older dune ridges.	
Aerial photos (1512/3-8) – There is a high, forested dune in front of McKenzie Lagoon but at the southern end the ground is much lower and inundation directly over the dunes may be possible in a large event. The lagoon also has connection with the river about 3 km inland (or less if the spit is breached) so would be expected to flood in a large inundation from the sea. There are at least two extensive wetland areas further inland that could be investigated for older events.	
Sutherland – Martins Bay would provide a good outer coast site with which to compare any record investigated within Milford Sound.	
Wellman & Wilson, 1964. Notes on the Geology and Archaeology of the Martins Bay District. See reprint.	
Big Bay (D39)	Moderate-High
Turnbull – A similar possibility to that at Martins Bay may also exist at Big Bay. The back-beach dunes are lowest at McKenzie Creek at the south end of Big Bay where the beach is gravel. However, McKenzie Creek often floods and ponds behind the back-beach dune ridge, so any tsunami sediment signature will probably be overwhelmed by fluvial deposits.	
Johnson (pp. 35-39) – The beach at Big Bay is probably more sheltered than those at Barn Bay or Martins Bay because it is embayed a fair distance from the open coast. The beach is backed by numerous lines of low dunes and beach ridges. Dunes are generally 3-6 m high with the main foredune 9 m in places. Near the centre of the bay there are about 17 ridges on a shore-normal transect with undulations of between 2 and 4 m. The southern end of the bay is gravely.	
Aerial photos (1508/1-4 & 1507/1-4) – The beach ridge sequence does appear to be wettest and least vegetated at the southern end of the bay but as Mo noted, is prone to flooding, and a squarish patch suggests some of it has been cleared and possibly disrupted. The central part of the bay has the highest dunes and the northern end would have seen changes in the Awarua River course and flooding. The valley floor consists of numerous beach ridges / low dunes instead of several large dune complexes as at Martins Bay. Wetland areas are less extensive and probably hold much shorter sedimentary sequences but a shore-normal transect of ~8 major dune hollows may cover an equivalent period of time to that in the wetlands at Martins Bay. Waiuna Lagoon could potentially hold a long, high-resolution sedimentary record. However there is currently 1.5 km of heavily forested beach ridges between it and the sea or a much longer connection via the windy Awarua River so it would not be any use for recent inundation events.	
Barn Bay	Moderate-High
McSaveney – The Cascade River Mouth area is one of the best places Mauri has seen for paleotsunami work in this part of the country because extensive wetlands occur adjacent to the coast. Barn Bay is at the Hope River mouth south of the Cascade River and may be a slightly better site for not having the large river influence.	
Johnson (pp. 32-34) – The southern half of Barn Bay is a sandy beach backed by low dunes with swampy ground behind. Large dunes occur only at the northern end of the bay and in some places sand is burying the forest – iron pans in the sand are evidence of old forest soils buried by sand and re-exposed.	
Aerial photos (1495/1-3) – In the centre of the bay there is a small lagoon behind low dunes and a fairly extensive triangular shaped area of swamp. Although there is likely to be a lot of sand in the system, the southern end of this swamp may be far enough south not to be overwhelmed by the wind-blown sand at the northern end of the bay.	

Appendix 4 Model description

The numerical model used in this study is a general-purpose hydrodynamics and transport model known as RiCOM (River and Coastal Ocean Model). The model has been under development for several years and has been evaluated and verified continually during this process (Walters and Casulli 1998; Walters 2002, 2004, 2005a, 2005b). The hydrodynamics part of this model was used to derive the results described in this report.

The model is based on the Reynolds-averaged Navier-Stokes equations (RANS) that are time-averaged over turbulent time scales. For the simulation of weakly dispersive surface waves, these equations are averaged over water depth to derive a set of equations similar to the standard shallow water equations but contain additional terms that describe non-hydrostatic forces (Stelling and Zijlema 2003; Walters 2005b).

The free surface equation is derived from vertically integrating the continuity equation

$$\frac{\partial \eta}{\partial t} + \nabla \cdot (H\mathbf{u}) = 0 \tag{1}$$

where $\eta(x,y,t)$ is the water-surface elevation measured from the vertical datum, ∇ is the horizontal gradient operator, $h(x,y)$ is the land elevation measured from the vertical datum, and $H = \eta(x,y,t) - h(x,y)$ is the water depth. The vertical datum is arbitrary, but is usually set equal to the average water surface elevation (sea level). This choice minimizes truncation errors in the calculation of the water surface gradients.

After depth-averaging, the horizontal momentum equation becomes

$$\frac{D\mathbf{u}}{Dt} + \mathbf{f} \times \mathbf{u} = -g\nabla\eta - \frac{1}{2}\nabla(q) - \frac{q}{2H}(\nabla\eta + \nabla h) + \frac{1}{H}\nabla \cdot (HA_h \nabla \mathbf{u}) - \frac{\tau_b}{\rho H} \tag{2}$$

where D/Dt is a material derivative, **u** is the depth-averaged velocity, **f** is a vector representation of the Coriolis parametersetersetersete, q is dynamic pressure which varies linearly in the vertical with $q = 0$ at the free surface, A_h is horizontal eddy viscosity, and τ_b is bottom friction. Surface stress and atmospheric pressure have been neglected. Bottom friction is written as

$$\frac{\tau_b}{\rho H} = \frac{C_D |\mathbf{u}| \mathbf{u}}{H} = \gamma \mathbf{u} \tag{3}$$

where C_D is a drag coefficient and γ is defined by (3). Equations (1) and (2) with $q = 0$ form the classical shallow water equations.

Next, the vertical momentum equation and the continuity equation must be depth averaged to derive governing equations for vertical velocity w and dynamic pressure q. The former is

$$\frac{Dw}{Dt} = -\frac{(q_\eta - q_h)}{H} = \frac{q_h}{H} \tag{4}$$

where w is the depth-averaged vertical velocity, q_η is dynamic pressure at the free surface, q_h is dynamic pressure at the bottom, and the vertical viscous terms have been neglected. The vertically integrated continuity equation is expressed as

$$\int_h^\eta \nabla \cdot \mathbf{u} dz + w_\eta - w_h = 0 \tag{5}$$

This equation is written in finite volume form when it is discretized. In order to allow flexibility in the discretization of the model grid across the continental shelf, finite elements with unstructured triangular and quadrilateral elements of varying-size and shape are used for the spatial approximation. The time-marching algorithm is a semi-implicit scheme that removes stability constraints on gravity-wave propagation (Casulli and Cattani, 1994). The advection scheme is semi- Lagrangian which is robust, stable, and efficient (Staniforth and Côté, 1991). Wetting and drying of intertidal or flooded areas occurs naturally with this formulation and is a consequence of the finite volume form of the continuity equation and method of calculating fluxes through the element faces. At open (sea) boundaries, a radiation condition is enforced so that the outgoing wave will not reflect back into the modelled area.

The equations are solved with a split-step method. In the first step, the equations are solved with *q=0* to derive an approximate solution for the dependent variables (Walters and Casulli, 1998). In the second step, an equivalent of a pressure Poisson equation is solved for q and the velocity solution from step one is corrected (Walters, 2005b).

This method can be expanded in a straightforward manner to 3 dimensions when greater accuracy is required (Stelling and Zijlema, 2003) and provides an alternative to the Boussinesq equations which use higher-order velocity and geometry correction terms in the momentum equation (Lynett and Liu, 2002).