THE FIORDLAND EARTHQUAKE OF 10 AUGUST, 1993: A Reconnaissance Report Covering Tectonic Setting, Peak Ground Acceleration, and Landslide Damage

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ABSTRACT

On 10 August, 1993, a M_L 6.7 (M_S 7.1, M_w 7.0) earthquake occurred c. 10 km offshore of western Fiordland, New Zealand (45.2° S, 166.7° E). Its hypocentre is approximately 20 km deep which places it on, or close to, the interface between the subducted Australian plate and the overriding Pacific plate. The focal mechanism for the mainshock indicates reverse faulting on either a steep west-dipping, or shallow southeast-dipping plane. Analysis of a subset of the over 7,000 recorded aftershocks defines a shallow (c. 15°) southeast-dipping plane, roughly 25 km long and 15 km wide, that probably represents the rupture surface of the mainshock.

The earthquake was strongly felt by fishermen offshore in the epicentral region. Onshore, there were no reports of damage to man-made structures. The maximum peak ground acceleration recorded was 0.08 g at Te Anau, about 73 km from the epicentre. The attenuation of peak horizontal ground acceleration for this event is similar to the attenuation of other shallow crustal earthquakes in New Zealand.

The number of landslides triggered by this event is at least an order-of-magnitude less than the number of pre-existing landslide scars. The highest concentration of new slides appears to be in the Vancouver Arm/Hall Arm region, c. 45 km south-southeast from the epicentre. Many of the new slides were narrow, shallow seated failures, or small reactivated portions of older slides. The two largest earthquake-triggered landslides observed are located near Hall Arm, and in the Freeman Burn north of Lake Manapouri. Except perhaps for these two slides, all other observed earthquake-triggered slides will be indistinguishable from storm-generated slides once re-vegetated.

INTRODUCTION

On 10 August, 1993, a large earthquake occurred off the coast of western Fiordland, approximately 10 km west from Secretary Island (Figure 1). Immediately following the earthquake, which is the largest shallow earthquake to have occurred in New Zealand since the 1968 Inangahua earthquake, the Institute of Geological & Nuclear Sciences deployed a reconnaissance team to conduct an aftershock study, and to investigate any damage and tectonic deformation that may have resulted from the earthquake. The earthquake triggered a number of the Institute's strong motion accelerographs, and a separate team collected the resulting records.

The purpose of this paper is four-fold: 1) to describe the tectonic setting of the earthquake and its aftershocks, 2) to document the peak ground accelerations recorded for this event, and their attenuation, 3) to document the landslide damage observed by the reconnaissance team, and 4) to present observations pertaining to the tectonic deformation that resulted from the earthquake, and to assess the extent of this deformation using elastic dislocation modelling. Each of these aspects is presented separately in the sections below.

TECTONIC SETTING AND AFTERSHOCKS

The Fiordland earthquake of 10 August, 1993, occurred in an area where convergence of the oceanic Australian plate and the continental Pacific plate is manifest in a steep, southeast-dipping seismic zone of intermediate depth earthquakes and abundant overlying shallow activity. The steeply dipping zone of earthquakes represents the seismicity associated with the subduction of the Australian plate beneath the Pacific plate. The N65E direction of convergence of the Australian plate relative to the Pacific plate is highly oblique to the N45E-N25E strike of the dipping seismic zone.

Because the locations of earthquakes determined from the few permanent seismograph stations in the Fiordland region are

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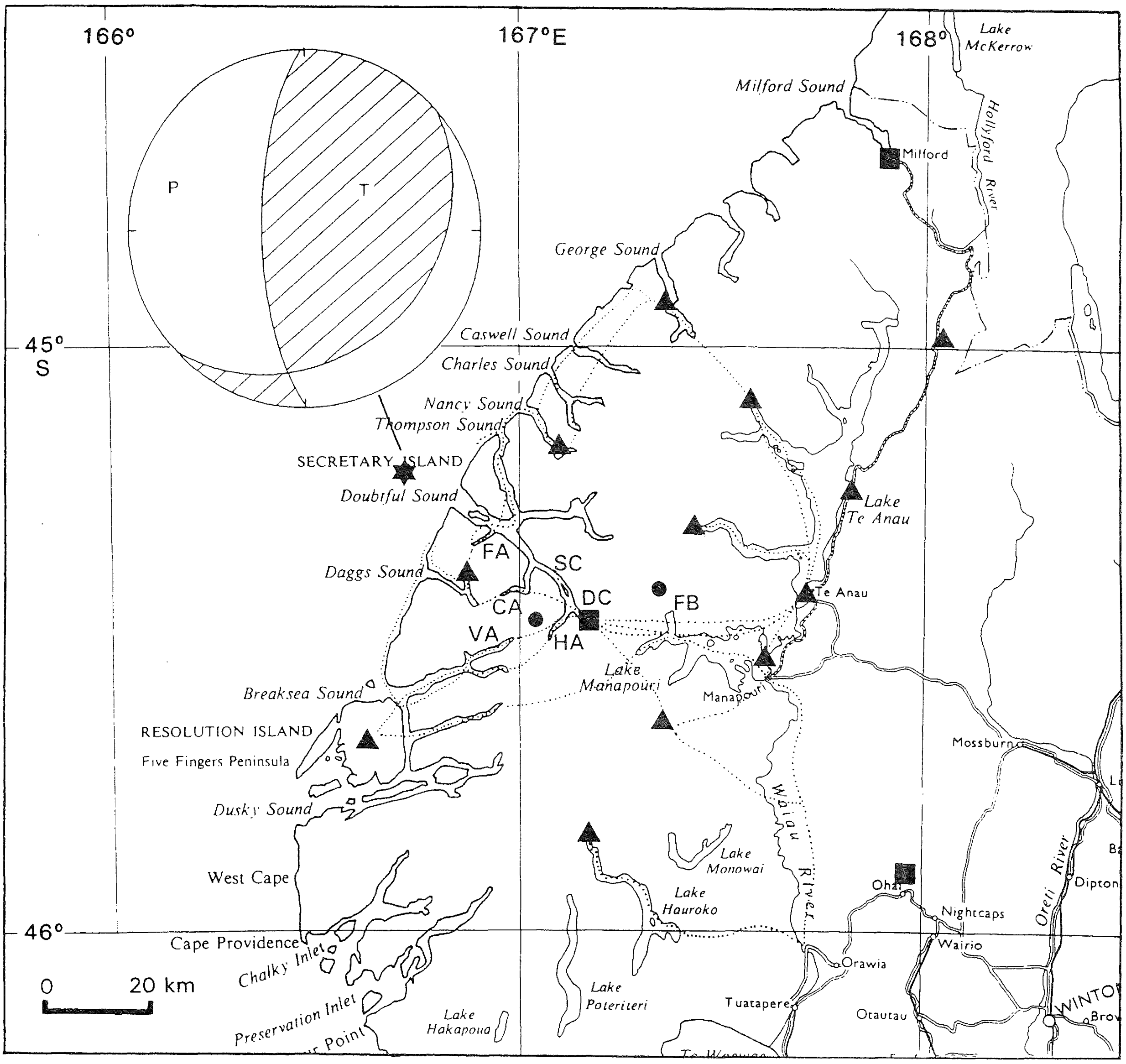

Figure 1. ***Location (star) and focal mechanism of the mainshock. The mechanism shown is the best double couple of the centroid-moment tensor solution determined by Harvard University. It is an equal area projection of the lower hemisphere, with compressional quadrants shaded, and P and T denoting the axes of compression and tension respectively. Squares are permanent seismograph stations in the Fiordland region, and triangles are temporary seismographs installed soon after the mainshock to determine accurately the aftershock distribution. Routes taken to deploy and retrieve the temporary seismographs are shown as dotted lines. Locations of the two largest earthquake-triggered landslides are shown as solid circles. Other locations noted in text are: CA, Crooked Arm; DC, Deep Cove; FA, First Arm; FB, Freeman Burn; HA, Hall Arm; SC, Shelter Cove; VA, Vancouver Arm.***

inadequate for detailed studies of the seismicity associated with the plate boundary zone, 25 portable digital seismographs were operated in the region from March to June, 1993, and 985 local events were well recorded. A joint inversion of hypocentres and 3-D velocity structure has produced a clear picture of the subducted Australian plate and new information on the highly heterogeneous seismic velocities in the Fiordland region [9].

The seismicity associated with the subducted Australian plate can be divided into three sections from north to south. The central section, which extends from just north of George Sound to Dagg Sound (Figure 1), is much more active than the others. The strike of the subducted plate changes from about N25E in the southern section to N45E further north, with the bending concentrated in the active central section. Here the dip of the subducted plate increases very rapidly to be nearly vertical below 65 km depth. Both the steepness and intense seismic activity of the subducted plate can be explained by its meeting strong resistance from a thickened continental lithosphere. This interpretation is consistent with the relatively low mantle seismic velocities east of the subducted plate.

The 1993 Fiordland earthquake occurred in this active central section, west-northwest from the entrance to Doubtful Sound.

TABLE 1. The 1993 Fiordland earthquake

Origin time:	00h 51m 51.61±0.65s (UT) 10 August, 1993
Epicentre:	45.21±0.02°S 166.71±0.07°E
Depth:	5 km (restricted during routine computation) ~ 20 km from aftershock distribution and mainshock relocation
Magnitude:	M_L 6.7, M_s 7.1, M_w 7.0
Moment:	3.4 x 10^{19} Nm

Parameters listed are those determined routinely by the New Zealand Seismological Observatory, except for M_s which has been determined from stations worldwide by the US Geological Survey, and M_w and moment, which have been determined by Harvard University.

Details of the mainshock are given in Table 1, and the mainshock epicentre and focal mechanism are shown in Figure 1. Since the earthquake occurred some six weeks after the removal of the portable seismograph network, only the stations of the permanent seismograph network recorded the event. Because the earthquake was located outside this network, and the nearest station was 44 km away, its location is poorly controlled, especially in depth. Consequently, the depth of the earthquake was restricted to 5 km during routine processing, in order to minimise travel time residuals at the nearest stations. The focal mechanism indicates that the earthquake was predominantly a thrust event, either on a fault plane dipping 70° to the west, or one dipping 29° to the southeast.

As expected for a shallow M 7 earthquake, the mainshock was followed by many aftershocks. Since these aftershocks can be expected to delineate the rupture zone of the mainshock, twelve digital seismograph stations were established soon after the mainshock in order to study the aftershock sequence in detail. These are shown in Figure 1; many were at sites occupied in the March-June seismograph deployment. Cars, boats, and helicopters were used to deploy the portable seismographs. The first began recording two days after the mainshock, and the last was removed on 1 September, 1993.

From the time of the mainshock until the end of 17 August, 6,496 aftershocks were recorded on the permanent and temporary seismographs. This number of events represents approximately half the number of earthquakes routinely located throughout New Zealand by the Seismological Observatory in a normal year. Consequently, these aftershocks have been set aside for future analysis as resources permit, and routine analysis has recommenced from 18 August. From that time till the end of August, a further 762 shocks were located in the region of the mainshock, and these are shown in Figure 2a. Many of these events - especially those that occurred prior to the installation of the temporary seismographs - were assigned restricted depths during routine analysis because the station coverage provided poor control of the event location. Nevertheless, 344 events were sufficiently well recorded for free depths to be determined. Depth sections of these are shown in Figures 2b and 2c.

The earthquakes in Figure 2 define a rectangular aftershock zone, some 25 km long and 15 km wide. It strikes N30E-N35E, approximately parallel to the strike of the subducted plate in this region, and dips to the southeast. The aftershock zone is centred at about 20 km depth, indicating that the mainshock rupture must also have occurred at this depth. This is confirmed by a preliminary relocation of the mainshock and a

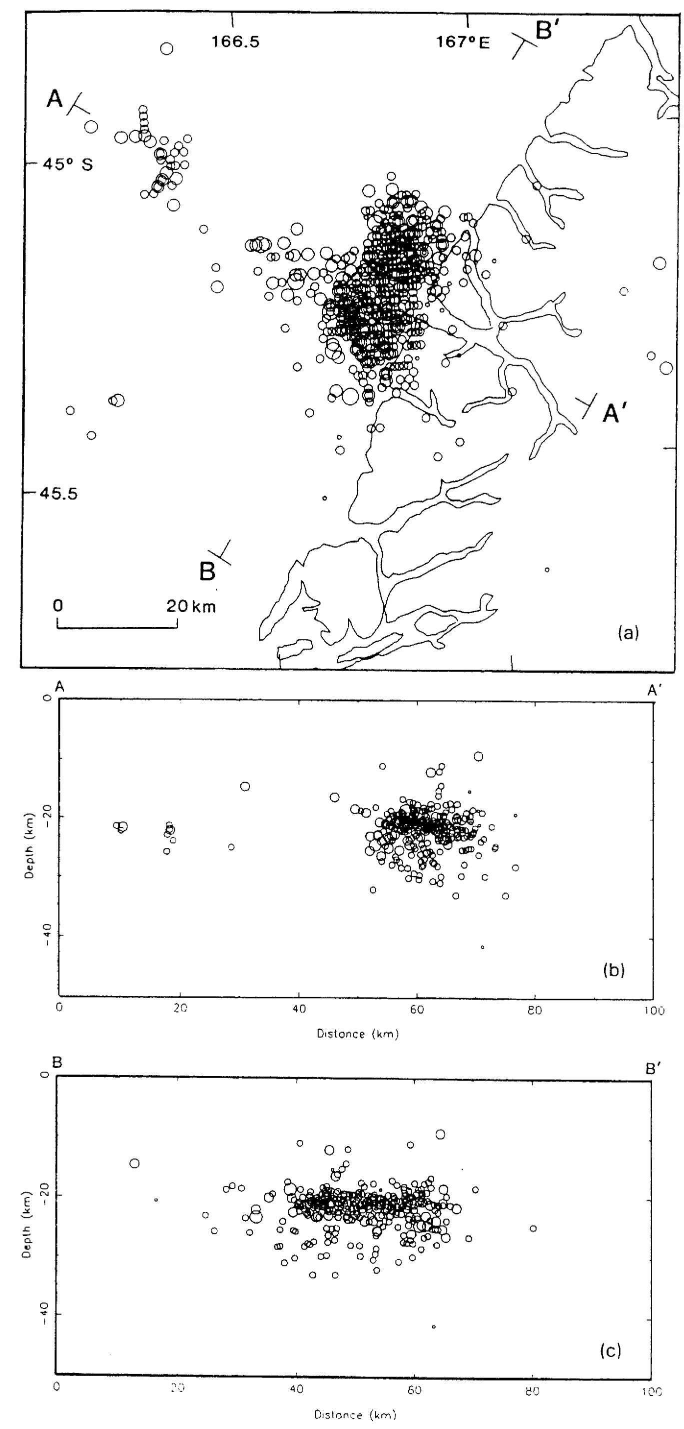

***Figure 2.** Aftershocks occurring during the period 18-31 August, 1993, inclusive. All located events are shown in 2a, but only events for which free depths could be determined are shown in the depth sections, 2b & 2c. Circle size is scaled to magnitude, with the smallest denoting events in the range M_L = 1.0-1.9, and the largest those with M_L = 4.0-4.9.*

few early aftershocks using the seismic velocity model determined from the March-June seismograph deployment.

When the aftershock zone is compared to the focal mechanism of the mainshock, it is clear that the shallow-dipping nodal plane represents the fault plane. The question now arises as to whether thrusting on this plane represents thrusting on the interface between the subducting Australian and overlying Pacific plates. This is likely to have been the case, since the aftershock zone shown in Figure 2 lies close to the plate interface defined by relocated seismicity recorded during the March-June seismograph deployment. A more detailed study of the aftershock sequence is planned to confirm this. The study will also investigate the relationship between the 1993 earthquake and previous large earthquakes near Doubtful Sound on 1989 May 31 (M_w 6.4) and near Charles Sound on 1988 June 3 (M_w 6.7) [8].

PEAK GROUND ACCELERATIONS

Currently, the New Zealand-wide strong-motion accelerograph network comprises 45 digital accelerographs, 180 film-recording accelerographs, and 70 scratch-plate acceleroscopes [2].

Twenty-two of the strong-motion accelerograph recording sites were within 300 km of the earthquake. Film records were obtained from 3 sites, Manapouri Power Station (at an epicentral distance of 57 km), Te Anau (73 km), and Queenstown (139 km), acceleroscope records from two further sites, Mossburn (123 km) and Gore (194 km), and digital accelerograms from Haast (218 km) and Dunedin (295 km). Peak ground accelerations from the records, excluding the one from the Manapouri Power Station, are plotted in Figure 3 (solid triangles). The maximum PGA recorded for this event was 0.08g at Te Anau. The apparent PGA from the Manapouri record was 0.03g, but the accelerograph had not triggered until during the "S-wave" part of the earthquake motions and so is likely to have missed the true peak.

The dashed line in Figure 3 is an expression derived from strong-motion data recorded throughout New Zealand [5]. Using ordinary least-squares regression techniques, various subsets of the data were fitted by expressions similar to

$$\log_{10}PGA = A + B \cdot M_w - C \cdot r - D \cdot \log_{10}r$$

where A, B, C, and D are fitted constants, PGA is the stronger of two horizontal components of ground acceleration, M_w is the moment magnitude, and r is the slant distance from the recording site to the nearest part of the fault plane at "centroid depth" h_c. McVerry et al. [5] found that the expression for deep earthquakes, $h_c > 50$ km, was significantly different from curves for shallower, crustal, earthquakes. Because the earthquake of 10 August, 1993, has a focal depth of about 20 km, the formula appropriate to "shallow upper crustal events" ($h_c \leq 20$ km) was used:

$$\log_{10}PGA = 0.309M_w - 0.00297r - 0.618\log_{10}r - 1.81. \qquad (1)$$

Also plotted in Figure 3 are PGA values from three previous shallow earthquakes centred in Fiordland. Note that the PGA values have been scaled to a magnitude of 7.1 using Eq. 1. The curve derived from New Zealand-wide data appears to be a satisfactory fit to the Fiordland data.

The "Doubtful Sound" earthquake of 31 May, 1989 (open squares on Figure 3) is of special comparative interest because its hypocentre maybe within 10-20 km of that of the 10 August, 1993, earthquake. The 1989 earthquake appears to have resulted in relatively larger PGAs than the 1993 earthquake, at least out to 150 km (Figure 3), but the spectral characteristics of accelerograms recorded at Te Anau showed some similarities (Figure 4). As previously noted by McVerry & Sritharan [4] the site response at Te Anau is broad-band in nature with spectral peaks out to 3 seconds period (0.3 Hz). Fourier spectra from the 1989 event had prominent peaks close to 3 seconds period in both horizontal components, whereas peaks at 1.6 seconds (0.6 Hz) tended to be more prominent in spectra from

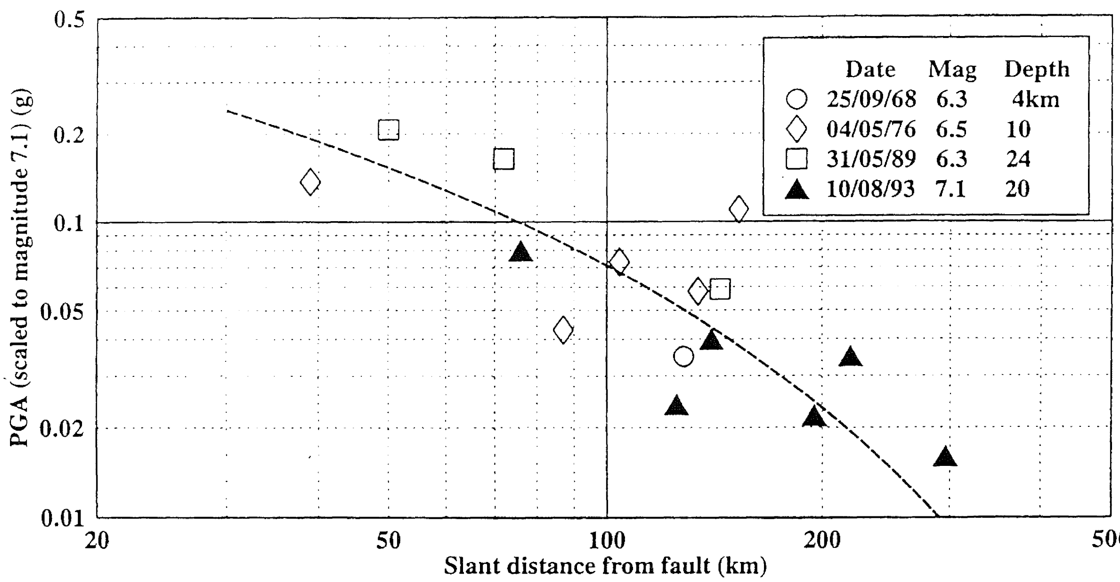

Figure 3. Peak ground accelerations recorded during four shallow earthquakes centred near the Fiordland coast. The accelerations have been scaled to a common magnitude of 7.1 using an empirical attenuation relationship derived from New Zealand data by McVerry et al [5]. The dashed line shows the PGA predicted by the relationship.

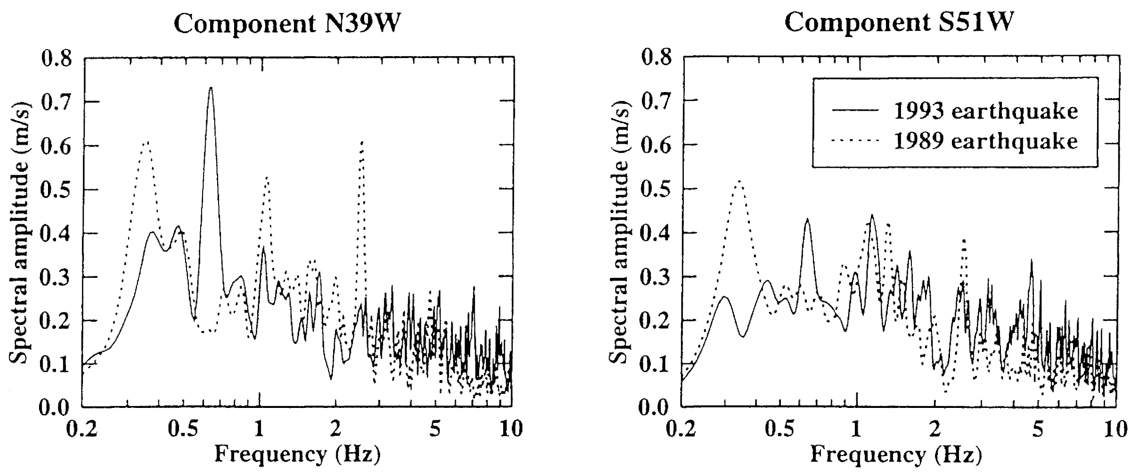

Figure 4. Comparisons of Fourier spectra of acceleration calculated from accelerograms recorded at Te Anau during the Doubtful Sound earthquake of 31 May, 1989 and during the earthquake of 10 August, 1993. All of the spectra show broad-band character.

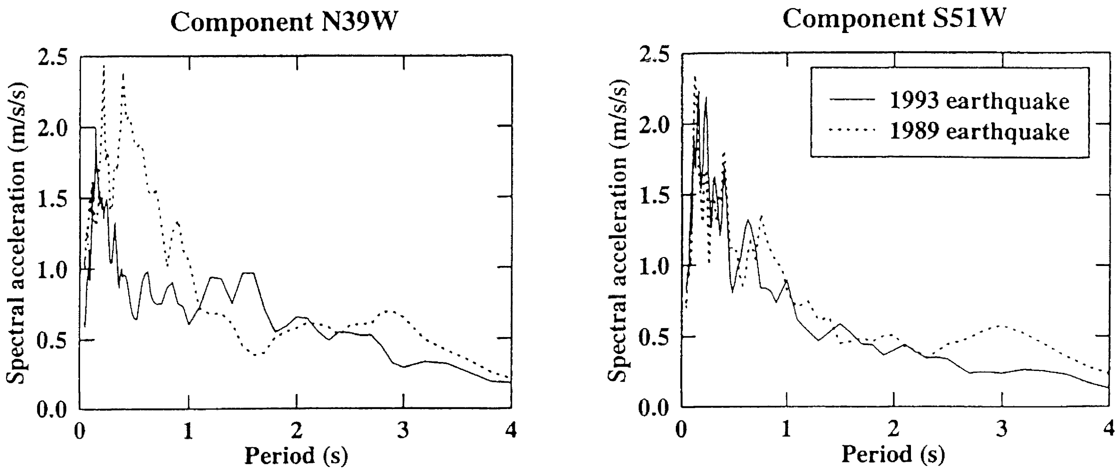

Figure 5. Comparisons of 5% damped acceleration response spectra calculated from the same accelerograms as in Figure 4. The enhanced low-frequency content of the 1989 record is clear.

the 1993 event. These differences are also reflected in the 5% damped acceleration response spectra (Figure 5).

LANDSLIDE DAMAGE

In the course of deployment, and retrieval, of the seismographs for the aftershock study, the locations of earthquake-triggered landslides were noted, and 150 photographs were taken of the epicentral region, mainly from the air. The routes taken to deploy and retrieve the seismographs are shown on Figure 1, and give an indication of the area covered by our landslide observations and photos.

The earthquake was strongly felt by fishermen offshore in the epicentral region, but caused little if any damage to man-made structures in the southern South Island. The primary record of the damage caused by the earthquake are the landslides it triggered. The vast majority of the earthquake-triggered landslides were narrow, shallow-seated failures involving mainly areas of dense vegetation cover. These slides looked in every respect similar to the narrow, shallow-seated, rain-triggered slides that are so common in the Fiordland region. In fact, the number of old landslide scars, presumably rain-generated, is at least an order of magnitude greater than the number of new earthquake-triggered slides.

The landslides triggered by this earthquake were all located west of SH 94, which, at its closest, is about 75 km from the epicentre. Many of the new slides were small reactivated portions of older slides. Two of the most striking earthquake-triggered landslides were located in the region near the mouth of Hall Arm, Doubtful Sound (Figure 6). However, it was the reconnaissance team's impression that the highest concentration of new slides was located in the Vancouver Arm/Hall Arm region (VA & HA on Figure 1), approximately 45 km south-southeast from the epicentre. The reason why this concentration of slides is removed from the epicentral region has not been determined. The two largest earthquake-triggered landslides that were observed by the team were located near Hall Arm about 10 km west of Deep Cove, and in the Freeman Burn north of North Arm Lake Manapouri (Figure 1). Except perhaps for these two slides, all other observed earthquake-triggered slides will be indistinguishable from storm-generated slides once re-vegetated.

Figure 6. Shallow-seated landslides near the mouth of First Arm, Doubtful Sound, triggered by the 10 August, 1993, Fiordland earthquake. These slides are typical of many of those triggered by the earthquake.

Following the earthquake, the water in Crooked Arm of Doubtful Sound was discoloured, presumably the result of the new slide at the head of Crooked Arm. Muddy water was also entering Doubtful Sound at an inlet 2 km southeast of Shelter Cove (grid ref. S139 232220). Within a week after the earthquake the water was again running clear.

Only one or two small rock falls were observed along the steep shores of Lakes Manapouri and Hauroko. The boat driver on Lake Hauroko noted that most of the prominent old landslide scars were the result of heavy rains in January, 1984.

Considering the steep topography of Fiordland and the magnitude of the earthquake, there were surprisingly few landslides triggered by this event. However, it is important to note that the rocks in Fiordland are some of the strongest in New Zealand, and the average annual rainfall in Fiordland is one of the highest in New Zealand, ranging from 2.5 m/yr to greater than 6.5 m/yr. It is possible that the high rainfall, much of which falls in high intensity events, keeps the slopes washed free of loose or weathered material. As a consequence, there are few slopes that have an appreciable accumulation of weak material susceptible to failure during strong earthquake shaking, and deep-seated recent failures are relatively uncommon because of the general high strength of the rock mass. There are however, many huge deep-seated landslides in Fiordland, but these are ancient, and their inferred failure mechanism involves slope destabilization due to the retreat of the large valley glaciers at the end of the last glaciation [7].

TECTONIC DEFORMATION

Tectonic subsidence and uplift resulting from recent earthquakes in Japan and California have been documented using the horizontal banding and vertical zonation of intertidal plants and sessile animals as datums to measure the relative rise or fall of sea level resulting from the earthquake [1,3,6]. In Fiordland, there is a well defined zonation of intertidal organisms that can be used to measure vertical tectonic deformation. Also, there are several marine scientists who are experts in Fiordland's intertidal flora and fauna who have ongoing research in the Doubtful Sound area. To date, these scientists, for example Cameron Hay of the Department of Conservation, and Paul Meredith of the University of Otago, have noted no unequivocal evidence of tectonic uplift of intertidal organisms in the epicentral region, despite several visits to the area since the earthquake. The reconnaissance team also noted no compelling evidence of uplift.

In an attempt to assess the scale of tectonic deformation that may have resulted from this earthquake, forward elastic dislocation modelling of the mainshock was carried out. Elastic dislocation modelling assumes that the earth behaves as a homogeneous elastic half space with a cut whose opposite sides are displace with respect to each other. Input parameters required are the position and geometry of the fault in the half space, and the amount and direction of slip on it. For the Fiordland earthquake these values are constrained by the earthquake's moment and its aftershock distribution. Table 2 lists the model parameters used, and Figure 7 shows the elevation changes resulting from this model. The model

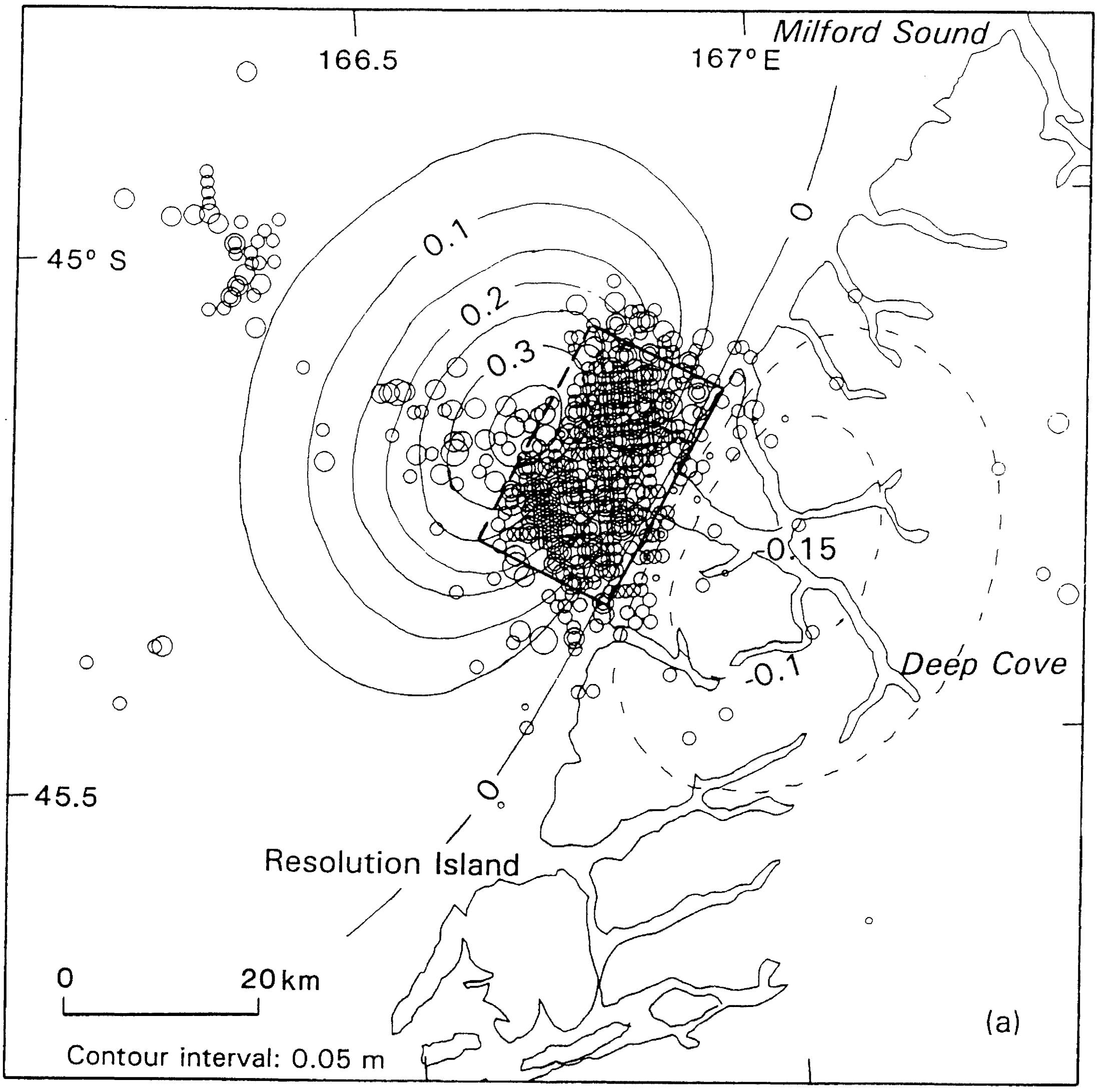

Figure 7. Elevation changes resulting from an elastic dislocation model of the 10 August, 1993, Fiordland earthquake. The rectangular box is the vertical projection of the fault plane with the shallow edge shown as dashed. Negative values are contoured with a dashed line. Model parameters are those given in Table 2.

TABLE 2. Elastic dislocation parameters

Top edge depth:	19 km
Bottom edge depth:	23 km
Length	25 km
Strike:	N32E
Dip:	15°SE
Slip:	3 m pure reverse dip-slip
Rigidity:	3×10^{10} N/m^2
Moment:	3.4×10^{19} Nm

predicts no uplift onshore, and is consistent with the field observations of no uplifted intertidal organisms in the fiords proximal to the epicentre. Figure 7 shows maximum uplift of 0.4 m occurring approximately 15 km offshore, and maximum subsidence of 0.15 m occurring 10-15 km inland from the coast, immediately southeast from Secretary Island. The modelling also predicts that the region near the southeastern end of Secretary Island moved horizontally to the northwest by 0.1-0.15 m.

If the Fiordland earthquake produced seabed deformations similar to those shown in Figure 7, then it would have displaced an appreciable volume of water and could have generated a tsunami. To date, however, the authors are not aware of any reports of a tsunami being generated by this earthquake. This maybe a consequence of the sparse human population in western Fiordland, and the general lack of people to note the occurrence of a minor tsunami.

CONCLUSIONS

The magnitude 7 Fiordland earthquake of 10 August, 1993, was located approximately 10 km west of Secretary Island, west-northwest from the mouth of Doubtful Sound. The aftershock distribution defines a shallow southeast-dipping plane centred at about 20 km depth that is roughly 25 km long and 15 km wide. This plane is consistent with the shallow dipping nodal plane of the mainshock's focal mechanism, and defines the rupture plane of the mainshock. The location and rupture zone geometry of the Fiordland earthquake suggest that it occurred on the subduction interface between the Australian plate and the overriding Pacific plate.

Peak ground accelerations recorded during the 10 August Fiordland earthquake, as well as those recorded during three other shallow earthquakes in Fiordland, appear to be well fitted by an attenuation relationship derived from shallow crustal earthquakes throughout New Zealand.

Several conclusions can be drawn from the reconnaissance investigation of the landslides triggered by this earthquake: 1) a heavy storm event does more damage to the Fiordland landscape than this magnitude 7 earthquake did; 2) the number of old landslide scars is at least an order of magnitude greater than the number of new earthquake-triggered slides; 3) the vast majority of landslides triggered by this event were narrow, shallow seated failures and/or small reactivated portions of older slides; and 4) once the new slides are re-vegetated, the effects of this earthquake will not be visible in the landscape.

Forward elastic dislocation modelling of the mainshock, constrained by the moment of the mainshock, and the location and geometry of the aftershock distribution, predicts that maximum uplift of c. 0.4 m occurred offshore of Fiordland, and maximum subsidence of c. 0.15 m occurred onland in the region immediately southeast of Secretary Island. The modelled deformation is consistent with field observations that onland there was no detectable uplift of marine organisms in the region about Doubtful Sound and Secretary Island.

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