THE M_W 7.2 FIORDLAND EARTHQUAKE OF AUGUST 21, 2003: BACKGROUND AND PRELIMINARY RESULTS

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ABSTRACT

The M_W 7.2 Fiordland earthquake of August 21 2003 was the largest shallow earthquake to occur in New Zealand for 35 years. Because of its location in an unpopulated area, it caused only minor damage to buildings, roads and infrastructure. It triggered numerous landslides on steep slopes in the epicentral region, where intensities reached MM9. Deployments of portable seismographs, strong motion recorders and GPS receivers in the epicentral region immediately after the event have established that the earthquake involved thrusting at the shallow part of the subduction interface between the Australian and Pacific plates. Recently installed strong motion recorders of the GeoNet network have ensured that the earthquake is New Zealand's best recorded subduction interface event. Microzonation effects are clear in some of the records. Current peak ground acceleration attenuation relationships for New Zealand subduction interface earthquakes underpredict the ground motions recorded during the earthquake, as was the case for previous large events in Fiordland in 1993 and 1989. The four portable strong motion recorders installed in the epicentral region have provided excellent near-field data on the larger aftershocks, with recorded peak ground accelerations ranging up to 0.28g from a nearby M_L 6.1 event.

INTRODUCTION

On 21 August 2003 at 12h12m UT (12 minutes after midnight on August 22 local time) a large M_W 7.2 earthquake occurred near Secretary Island in Fiordland. This was the largest event ever recorded instrumentally in the region, and the largest shallow earthquake to occur in New Zealand for 35 years. The earthquake triggered numerous landslides on steep slopes in the epicentral region, where intensities reached MM9. However damage to buildings and roads was only minor, because the epicentral region is unpopulated. The shock was felt as far away as Auckland in the North Island, some 1100 km distant. There were also reports of the shock having been felt in Sydney, Australia, 1850 km away. A small tsunami (300 mm peak to trough) from the event was recorded at Jackson Bay, 200 km northeast of the epicentre. The earthquake also produced a 150 mm tsunami at Port Kembla, Australia.

The earthquake was the latest in a series of moderate to large events in the Doubtful Sound region in the last 15 years (Fig. 1). The series began with the M_W 6.7 Te Anau earthquake of 1988, followed by the M_W 6.4 Doubtful Sound earthquake of 1989, the M_W 6.8 Secretary Island earthquake of 1993, and most recently the M_W 6.1 Thompson Sound earthquake of 2000. To understand the relationship of the 2003 earthquake to these previous events, and the nearby branches of the Alpine Fault, six portable seismographs were deployed in the epicentral region immediately after the mainshock. Four portable digital strong motion instruments were also deployed, with the aim of capturing near-field strong motion data from expected large aftershocks. Previously surveyed GPS sites were reoccupied so that the ground deformation due to the event could be quantified, and magnetotelluric measurements were made in order to track possible changes in the conductivity structure of the crust due to postseismic fluid redistribution. Also, a detailed inventory of landslides and other ground damage was completed.

Here we report on the preliminary results of the post-earthquake surveys, as well as data on the earthquake sequence recorded by the permanent GeoNet national seismograph and strong-motion network.

TECTONIC SETTING

In the Fiordland region, the motion of the Australian plate relative to the Pacific plate is c. 34 mm/yr at 062° (DeMets *et al.* 1994; Figure 1). Thus the plate boundary is characterised by highly oblique convergence. The convergent component of the relative plate motion is being accommodated largely by southeasterly subduction of the Australian plate, while the along-strike component of motion is mostly accommodated by the Alpine fault. The detailed morphology of the dipping seismic zone associated with the subducted Australian plate

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has been revealed by earthquakes relocated by Eberhart-Phillips & Reyners (2001). The dipping seismic zone steepens from south to north, becoming near-vertical below c. 75 km depth north of Doubtful Sound. This steepening is accompanied by a sharp change in strike of the subduction zone. South of Doubtful Sound, the average strike is 023°, whereas north of Doubtful Sound it is 040°.

Stresses and strains in the subducted plate are consistent with a model in which the sharp change in strike of the subduction zone is a consequence of the obliquely converging subducted plate having to bend around a zone with high seismic velocity ($Vp > 8.5$ km/s) in the uppermost mantle of the overlying Pacific plate (Reyners *et al.* 2002). Focal mechanisms and stress inversions for earthquakes in the overlying plate indicate normal faulting shallower than 16 km, and thrust faulting at greater depths (Reyners et al. 2002). This stress and strain regime is related to the 17° change in strike of the subduction zone near Doubtful Sound, which leads to an arching up of the subducted plate at the bend.

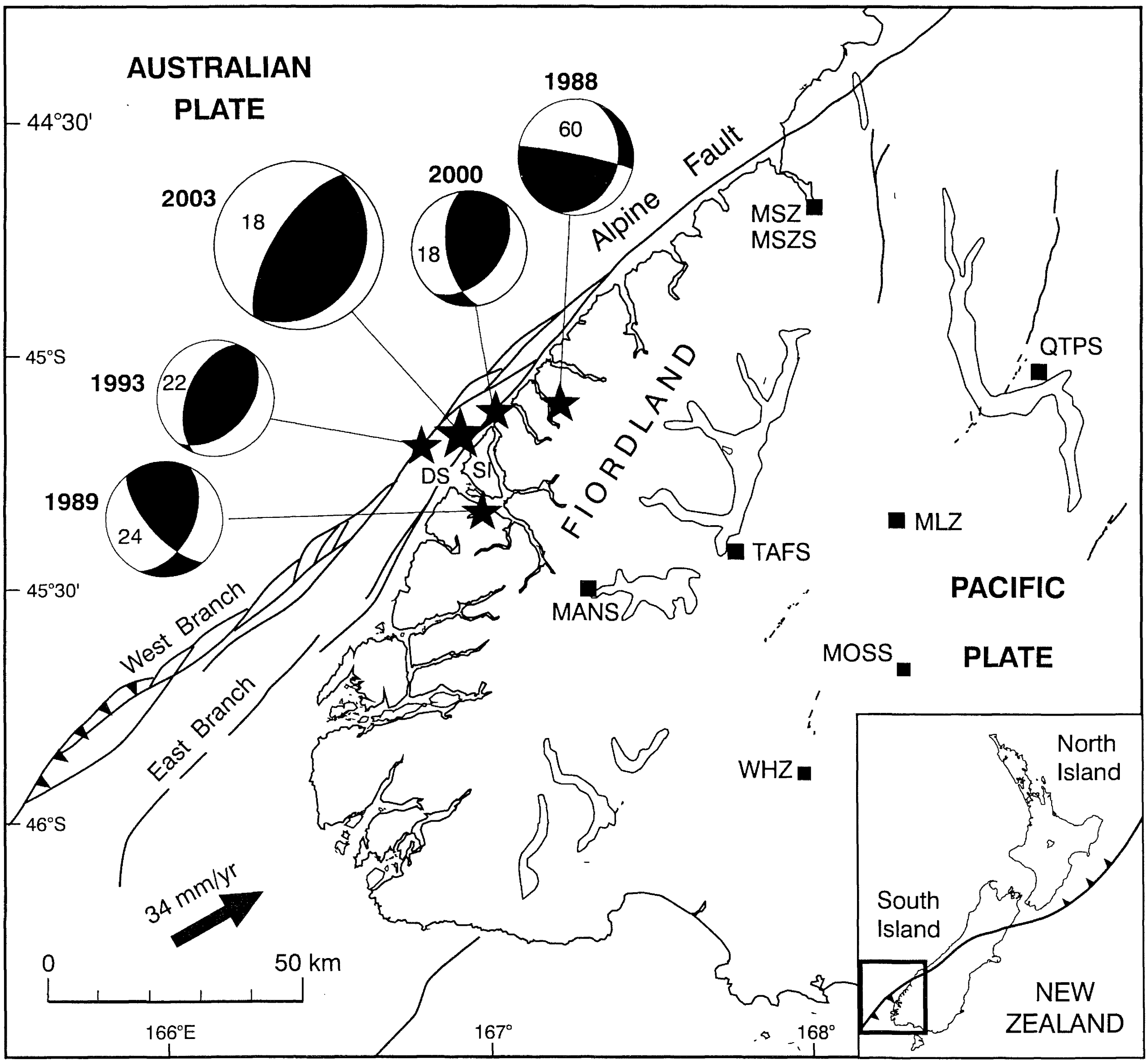


Figure 1. Tectonic setting of the 2003 Fiordland earthquake, and recent earthquakes of $M_W > 6$ in the same region. Focal mechanisms (lower hemisphere) and depths in kilometres are shown for each event. The mechanism for the 2003 earthquake is the Harvard centroid moment tensor solution, others are from Reyners & Webb (2002) and Robinson et al. (2003). Squares are seismographs and strong motion recorders of the GeoNet network used to locate the 2003 earthquake sequence. The arrow indicates the velocity of the Australian plate relative to the Pacific plate (DeMets et al., 1994), SI denotes Secretary Island, and DS denotes Doubtful Sound.

The focal mechanisms and depths of large historical earthquakes in Fiordland have recently been determined by Doser et al. (1999). In the period 1938-88, none of these large events appears to have occurred at the shallow part of the plate interface. Rather, they appear to be associated with complex deformation within both the subducted and overlying plates. The M_W 6.7 Te Anau earthquake of 1988 is an example of such an intraplate event. It involved down-dip tearing of the subducted plate in the 40-70 km depth range (Reyners et al. 1991). The M_W 6.1 Thompson Sound earthquake of 2000 was also an intraplate event, involving oblique thrusting within the overlying plate (Robinson *et al.*, 2003).

In contrast, the M_W 6.4 Doubtful Sound earthquake of 1989 and the M_W 6.8 Secretary Island earthquake of 1993 both involved slip at the plate interface (Reyners & Webb, 2002). Their rupture zones abut rather than overlap, and the region where they meet lies vertically beneath the surface trace of the East Branch of the Alpine Fault. Slip during the deeper 1989 event was approximately in the plate convergence direction, whereas slip during the shallower 1993 event was approximately down the dip of the subducted plate. This requires slip partitioning in the shallow part of the subduction zone, and suggests that the East Branch of the Alpine Fault is active in this part of Fiordland.

THE MAINSHOCK FAULT PLANE AS DETERMINED FROM GPS MEASUREMENTS

As the earthquake occurred close to the Fiordland coast, we have the opportunity to model the fault plane that ruptured by measuring coseismic displacements on land using GPS receivers. In February 2001, GPS measurements were made at sites throughout Fiordland by a joint Otago University – GNS team. A number of these sites in the epicentral region were reoccupied during the period 29 August – 3 September 2003 (Fig. 2). The measured displacements may thus include a small component due to some of the larger aftershocks, and any postseismic creep.

To determine the coseismic displacements, we first subtract expected interseismic displacements between February 2001 and the time of the mainshock. These have been estimated using a recent version of the New Zealand contemporary deformation model, which has been derived from campaign GPS observations over the past ten years (Beavan & Haines, 2001). We then model the earthquake fault using the disloc99 non-linear inversion code (Darby & Beavan, 2001), assuming uniform slip on a rectangular fault in an elastic half-space. For the modelling we fixed the strike of the fault to 030°, consistent with the Harvard centroid moment tensor (CMT) solution (Fig. 1). We omitted the observed displacement data from a GPS station on Secretary Island as it showed highly anomalous motion relative to the uniform slip model. This station may have been subject to local instability as a result of strong ground shaking, as it overlies the mainshock rupture zone. Alternatively, our assumption of uniform slip may be in error. We will investigate this in future using more sophisticated variable slip models.

The best-fitting fault model is shown in Figure 2. The modelled horizontal displacements fit those observed very well. The maximum horizontal motion recorded was 170 mm WNW at the Museum Range GPS site (DF4Q). Once the fault strike was fixed, the inversion code was able to solve for all other fault parameters, and these are shown on the lower part of Fig. 2. The derived moment corresponds to M_W 7.1, somewhat lower than the Harvard CMT value. Both the dip and rake (i.e. direction of slip) of the fault are similar to those of the low angle nodal plane of the Harvard CMT solution. The mainshock clearly involved low-angle thrusting in the direction of dip of the Fiordland subduction zone.

Vertical displacements predicted by the fault model are shown in Figure 3. The maximum vertical motion measured was 130 mm of subsidence at the Museum Range GPS site (DF4Q), though the model predicts only 95 mm of subsidence. At most other stations the modelled vertical displacement agrees with the observations to better than 10 – 15 mm. Apart from the Secretary Island site, there were no GPS sites in the high uplift region, so we are endeavouring to validate uplift along the coast through observations of bleaching of uplifted inter-tidal marine life. The observed tsunami is consistent with that expected from the predicted coseismic uplift.

THE EARTHQUAKE SEQUENCE

The earthquake occurred two years into the GeoNet project – a major initiative to modernize geological hazard monitoring in New Zealand funded by the Earthquake Commission (EQC) and operated by the Institute of Geological & Nuclear Sciences (GNS). As a result, this event was much better recorded than the previous large earthquakes in 1989 and 1993, and many of the data were available in near real time. In particular, data received by satellite telemetry from the seismograph at Wether Hills (WHZ), dial-up data from the seismograph at Milford Sound (MSZ), and dial-in data from strong motion recorders at Manapouri Power Station (MANS), Te Anau Fire Station (TAFS), Milford Sound (MSZS), Mossburn School (MOSS) and Queenstown Police Station (QTPS) proved very useful for rapidly locating the earthquake sequence (Fig. 1).

Here we use data from these stations in Fiordland and western Southland, together with available data from the seismograph at Mavora Lakes (MLZ), to relocate the mainshock and all aftershocks of $M_L \geq 5.0$ up until the end of September 2003. The earthquake had a rich aftershock sequence, with three events in the M_L 6.0-6.2 range, and 18 in the M_L 5.0-5.9 range up until the end of September 2003. For later events in the sequence, we supplement data from these stations with that from the network of portable recorders installed in the epicentral region.

The relocated events are shown in Figure 4, together with the fault model for the mainshock derived from GPS measurements. It should be noted that prior to the installation of the portable recorders, the nearest station to the sequence (MANS) was more than 40 km away. Thus the location accuracy of the first 15 $M_L \geq 5.0$ events is relatively poor, particularly in terms of depth. We will in future be able to improve on these locations by using aftershocks recorded by the temporary network to define station corrections relative to a three-dimensional seismic velocity model previously determined for this region by Eberhart-Phillips & Reyners (2001).

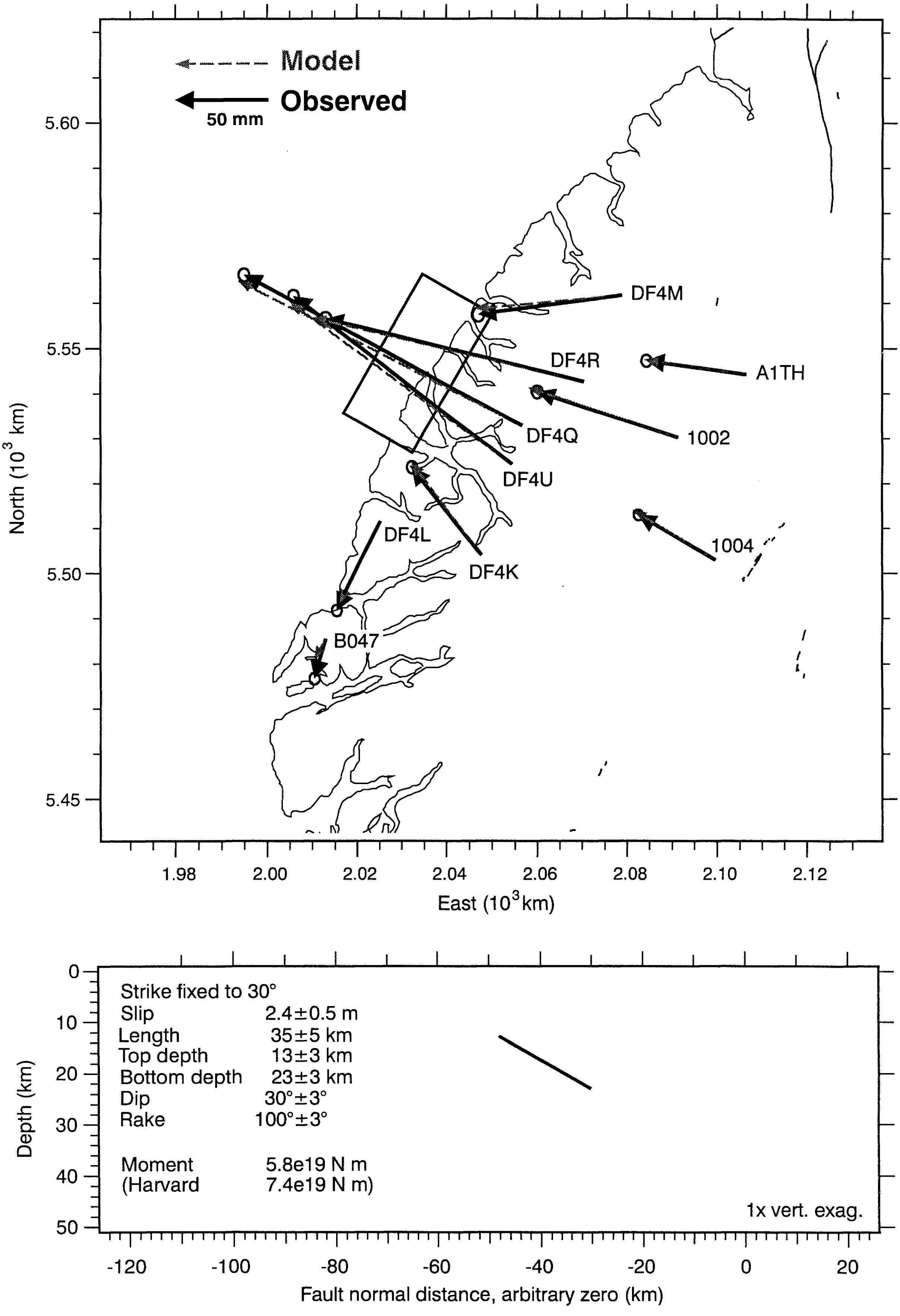


Figure 2. Preliminary fault plane for the mainshock from GPS observations. The ellipses at the ends of the displacement vectors are 95% confidence limits, and 1-σ errors in the fault parameters are given in the depth section.

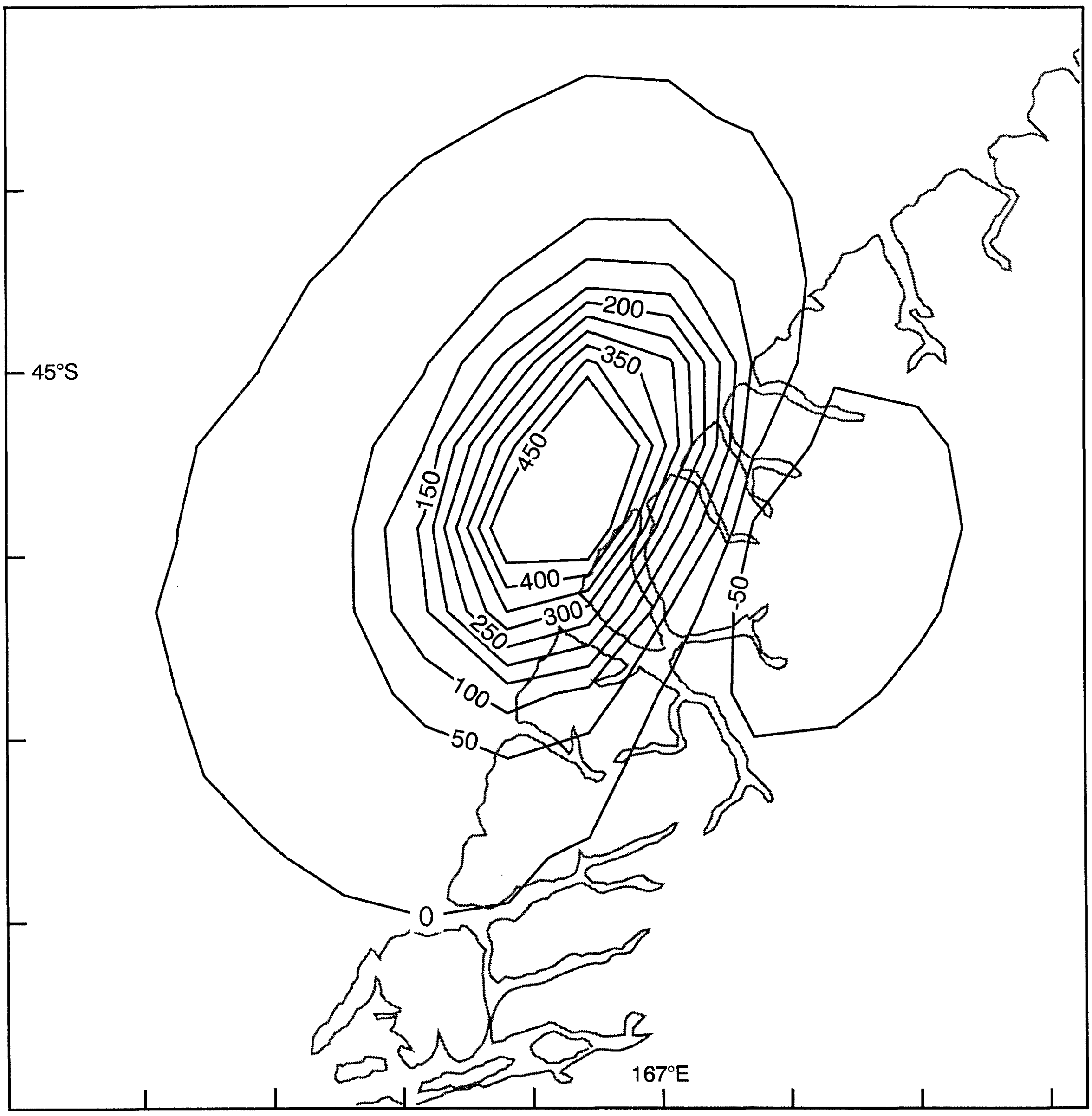


Figure 3. Predicted vertical displacement (in mm) from the GPS-derived fault model.

All but one of the large aftershocks occurred near the plate interface as determined by Reyners & Webb (2002). In particular, the seven aftershocks which were well recorded by the portable network lie close to the interface, and the fault modelled from the GPS observations (Fig. 4). These data, and the Harvard CMT solution (Fig. 1), confirm that the mainshock involved thrusting on the shallow part of the plate interface. The CMT solution for the M_L 6.1 aftershock of September 04, which lies close to the modelled fault plane, also shows thrusting (Fig. 4).

There were no obvious foreshocks in the epicentral region, apart from an M_L 4.6 earthquake 84 km deep within the subducted Australian plate directly downdip of the rupture zone some 33 hours before the mainshock (Fig. 4). The epicentres of the large aftershocks cluster around the preliminary fault plane as defined by the GPS data, except to the south, where they extend to Breaksea Sound, 30 km from the fault plane and well beyond the rupture zones of the 1993 and 1989 earthquakes. How this distribution of aftershocks relates to dynamic and static stress changes associated with the mainshock, and the effect of these earlier earthquakes, will be the subject of further study. An event that is likely to have been triggered by the stress changes induced by the mainshock is the shallow M_L 5.8 event which occurred on September 30 near Wet Jacket Arm (Fig. 4). This earthquake is located in the overlying Pacific plate, and the normal faulting indicated by its CMT solution is similar to that documented for smaller events in this plate shallower than 16 km (Reyners *et al.*, 2002).

STRONG MOTION DATA

Over the last two years, there has been a major upgrade of strong motion recording in New Zealand as part of the GeoNet Project. Nearly all of the obsolete stand-alone accelerographs in the strong-motion network have been replaced and new sites instrumented to give a total of 170 well-instrumented sites. The strong-motion stations are mostly equipped with Kinemetrics Etna recorders and Episensor accelerometers. Instrument specifications are as follows: triaxial force-balance accelerometer, range ±2g, 18-bit resolution, on-site storage capacity 48 minutes, and threshold triggering. GPS timing is provided at all stations, and data transmission is mostly by cell phone.

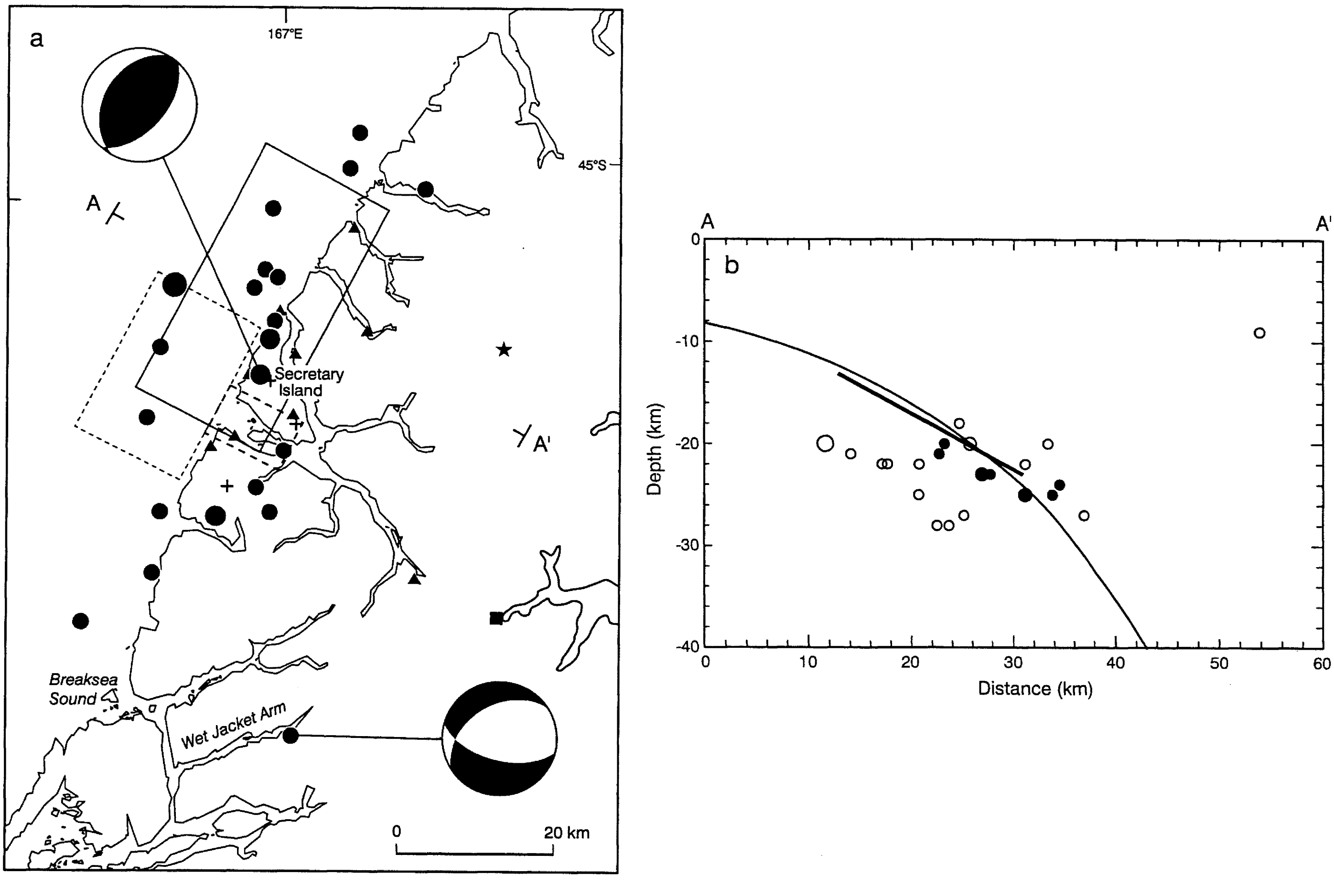


Figure 4.(a) Relocated $M_L \geq 5.0$ events in the earthquake sequence (circles) up until the end of September 2003. Circle size is scaled to magnitude with the largest circle denoting the epicentre of the mainshock. Harvard centroid moment tensor solutions (lower hemisphere) are shown for the M_L 6.1 aftershock beneath Secretary Island on September 04 and the M_L 5.8 triggered earthquake in the upper plate near Wet Jacket Arm on September 30. The star is the location of the M_L 4.6 earthquake 84 km deep within the subducted plate 33 hours before the mainshock. Triangles show the locations of portable seismographs and strong motion recorders, pluses are magnetotelluric stations, and the square is the permanent strong motion recorder at Manapouri Power Station. The solid box is the mainshock fault plane as determined from GPS observations, and the large and small dashed boxes are the fault planes for the 1993 and 1989 earthquakes respectively (Reyners & Webb, 2002). (b) Depth section of the earthquake sequence along the line A-A' shown in (a). Filled circles are the better located events, for which readings from the portable network were available. The straight line is the mainshock fault plane from GPS, and the curved line is the plate interface as determined by Reyners & Webb (2002).

Table 1: List of strong-motion recordings within ~ 400 km of the rupture plane of the Fiordland earthquake of 21 August 2003. Peak ground accelerations were less than 100 mm/s/s for all of the other stations. The ground subsoil categories are defined in Table 2.

Source Distance (km)	Peak Ground Acceleration (mm/s/s)			Name of Recording Site	Ground Subsoil Category
	Vertical	Horizontal 1	Horizontal 2		
44	1169	1277	1685	Manapouri Power Station (Meridian)	A
65	1216	1463	1427	Te Anau Fire Station	D
77	650	738	667	Milford Sound	B
112	583	707	668	Mossburn School	C
122	732	868	1021	Queenstown Police Station	C
167	120	245	194	Invercargill City Council	B
164	371	498	570	Wanaka National Park Headquarters	D
168	281	580	1126	Jackson Bay	C
190	358	399	332	Makarora Emergency Centre	D
191	299	693	715	Haast DOC Workshop	E
242	50	105	103	Balclutha District Council	B
248	45	51	61	Benmore Dam (Meridian)	B
258	27	39	53	Aviemore Dam (Meridian)	B
279	78	150	146	Dunedin Corstophine Substation	B
280	77	88	93	Dunedin Civil Defence	B
280	52	81	63	Dunedin GNS	C
280	123	304	331	Dunedin Kings High School	E
281	176	273	377	Dunedin St Kilda Fire Station	E
277	93	211	166	Mount Cook National Park Headquarters	D
288	37	50	56	Tekapo A Power Station (Meridian)	D
289	132	161	163	Fox Glacier Park Board Headquarters	D
334	40	66	61	Oamaru North Otago Museum	C
311	54	74	100	Fairlie District Council Garage	C
333	31	59	68	Timaru Roncalli College	C
345	103	148	142	Harihari Fire Station	D
361	27	28	26	Waitaha Valley (rock site)	A
388	14	26	45	Ashburton District Council	D
399	51	69	71	Kokatahi	C
401	26	69	62	Hokitika Medical Centre	D

Upgraded stations in the National Seismograph Network also have strong-motion capability. Further information on the network can be obtained from the project web site www.geonet.org.nz.

The new strong motion network recorded the mainshock very well, with 60 stations triggering. The closest was at Manapouri Power Station (surface facility), 44 km from the rupture zone, and the most distant was at Taumarunui High School, 958 km away. The peak ground accelerations of the three recorded components, distances from the source, and site classes for stations out to 401 km are listed in Table 1. None of the accelerations were particularly strong, which is not surprising given the epicentral distances involved. The largest peak ground acceleration, 0.17g, was recorded at the Manapouri Power Station site. Few records were obtained from Marlborough and none from the southernmost North Island, resulting in a data "shadow" as indicated by the dashed lines in Figure 5. As discussed later, a shadow zone was also apparent in the recorded motions, in that stations to the east of the extension of the western boundary of this shadow zone produced weaker motions than stations to the west at a similar distance from the source.

Microzonation effects are clear in some of the records. Acceleration response spectra from the sites closest to the source, Manapouri Power Station and Te Anau Fire Station, are compared in Figure 6. The Manapouri instrument is on strong rock and the Te Anau one on deep, compact gravels. The distances from the rupture plane are 44 and 65 km respectively. On going from rock to deep gravels there is a clear attenuation of short-period motions (periods < 0.25 s) and amplification of longer-period motions (periods > 0.7 s). Site effects are even more dramatic in Dunedin, at about 280 km distance from the source. In this case the subsurface materials include weathered rock, compact gravels above basalt rock, and soft alluvium above basalt rock. Short-period motions at the gravel site are attenuated relative to the weathered rock site, and the soft alluvium site shows considerable amplification relative to the other two sites in a period band of 0.3 to 1.8 s (Fig. 7).

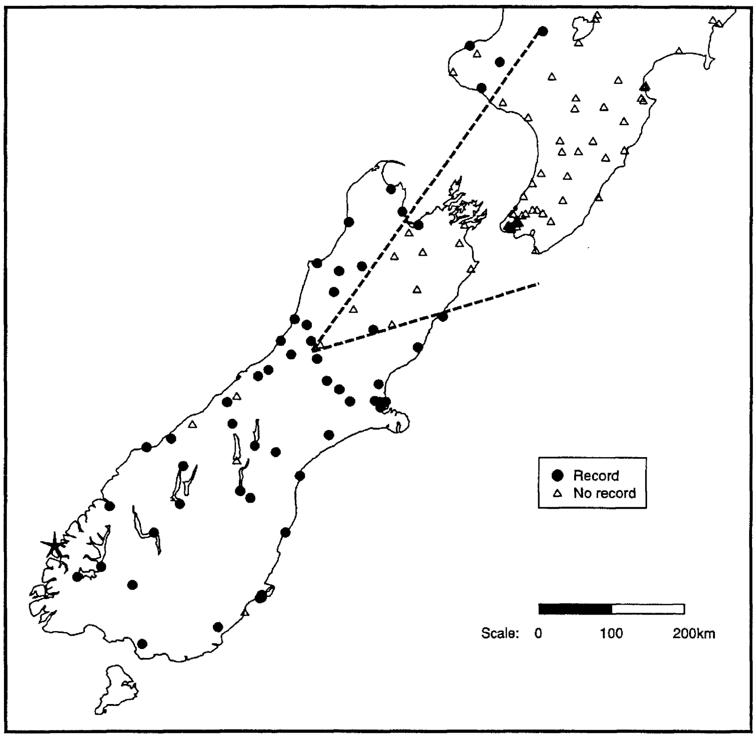


Figure 5. Sites from which strong-motion records were obtained. Note the data "shadow" encompassing the north-eastern part of the South Island and the southern part of the North Island (indicated by the dashed lines). Four recorders in the southern South Island were inoperative at the time of the earthquake.

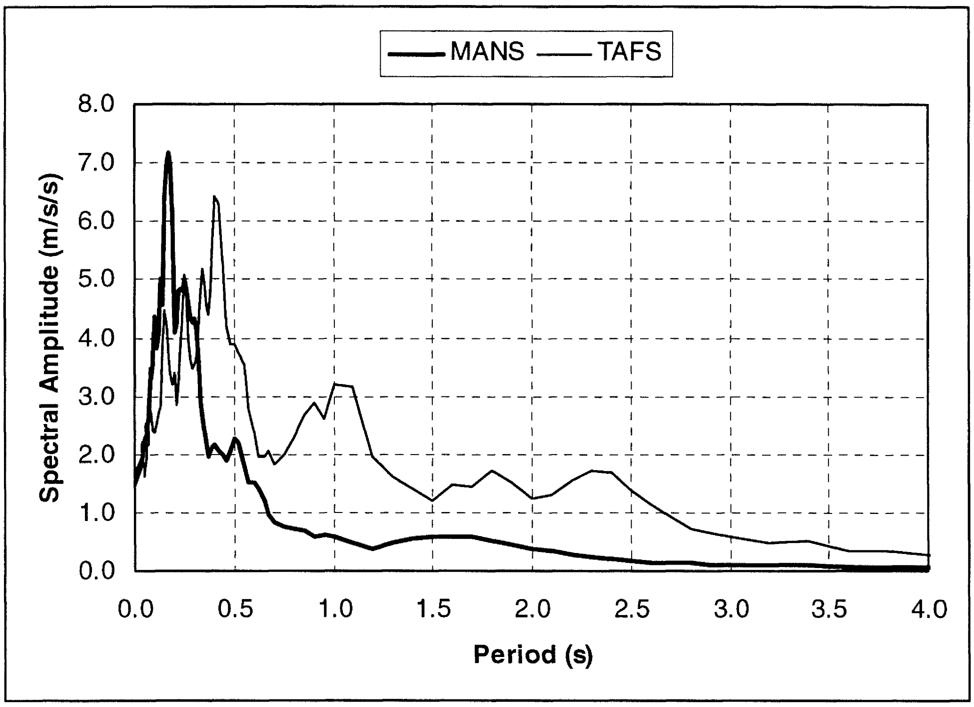
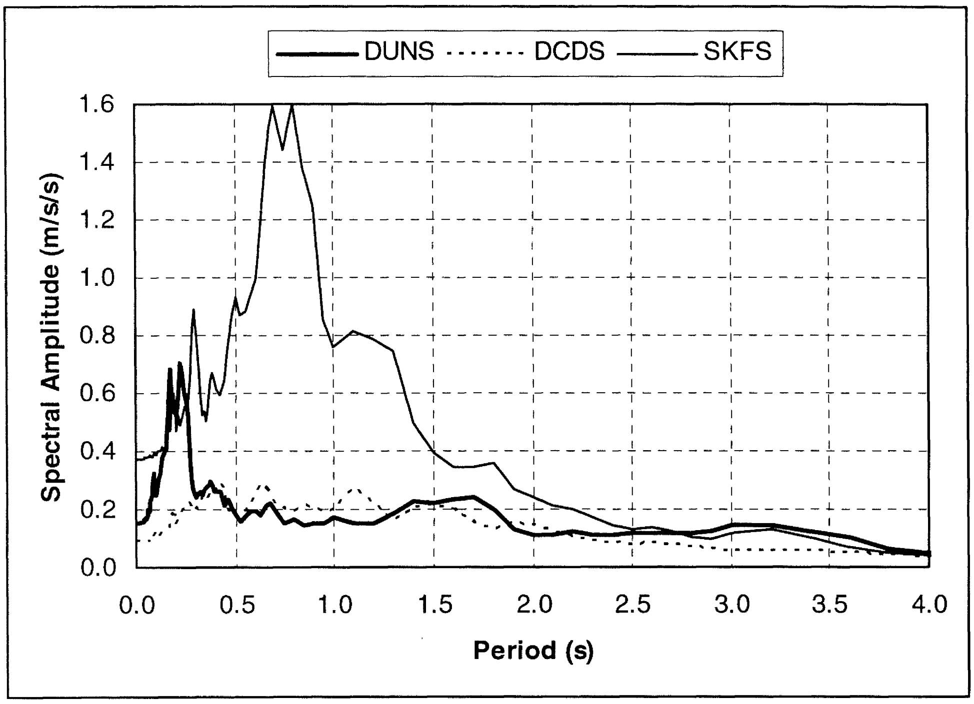


Figure 6. Acceleration response spectra from the two sites closest to the rupture zone. Site MANS (Manapouri Power Station – surface control room) is on strong rock, and site TAFS (Te Anau Fire Station) is on deep, mostly tightly packed, gravels.

Figure 8 shows a comparison of the attenuation of peak ground acceleration (pga) with attenuation models for subduction interface earthquakes developed from world-wide data (Youngs *et al.*, 1997), Japanese data (a preliminary version from Saiki *et al.*, 2003) and New Zealand data (McVerry *et al.*, 2000). The plots are for the geometric mean of the peak accelerations of the two horizontal components. The attenuation curves correspond to the 50-percentile values of this quantity. Their associated distributions are log-normal, with large variances. For example, the New Zealand model has a standard deviation of 0.45 in ln(pga) for magnitude 7.2 earthquakes, corresponding to a factor of nearly 1.6 for pga.



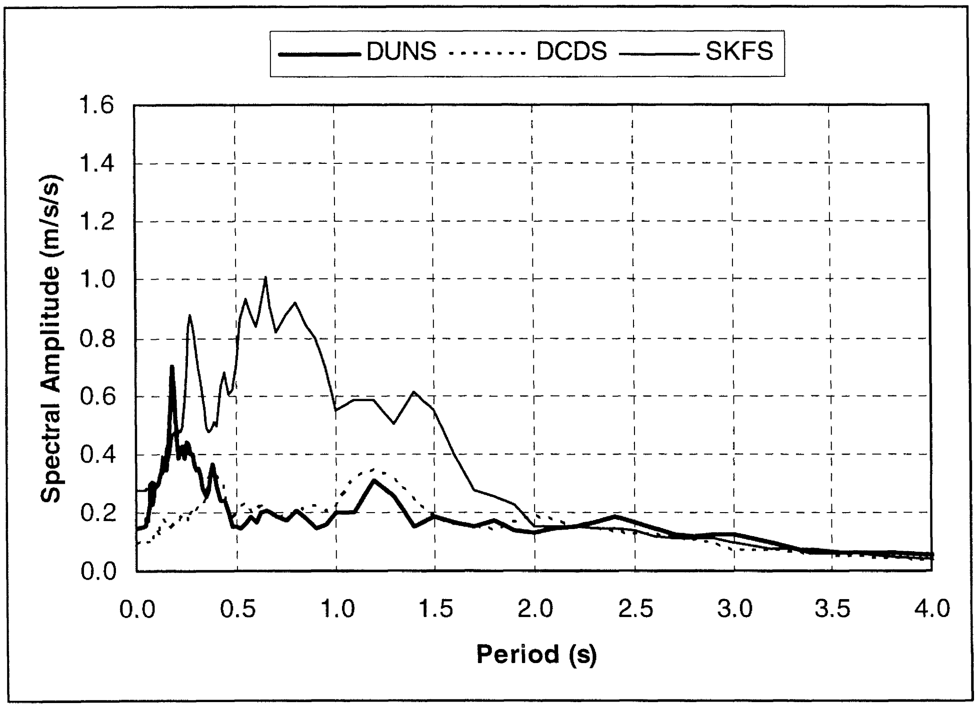


***Figure 7.** Acceleration response spectra from sites in Dunedin showing strong site effects. Site DUNS is on weathered rock in the suburb of Corstophine. There is apparent de-amplification of high-frequency motions from site DUNS to site DCDS (Dunedin Civil Defence – c. 5 m gravel above basalt), but strong amplification over the period range 0.25 s to about 1.8 s at site SKFS (St. Kilda Fire Station – 25 m soft soils above basalt). The upper set of curves is for the stronger horizontal component from each site and the lower set for the weaker one.*

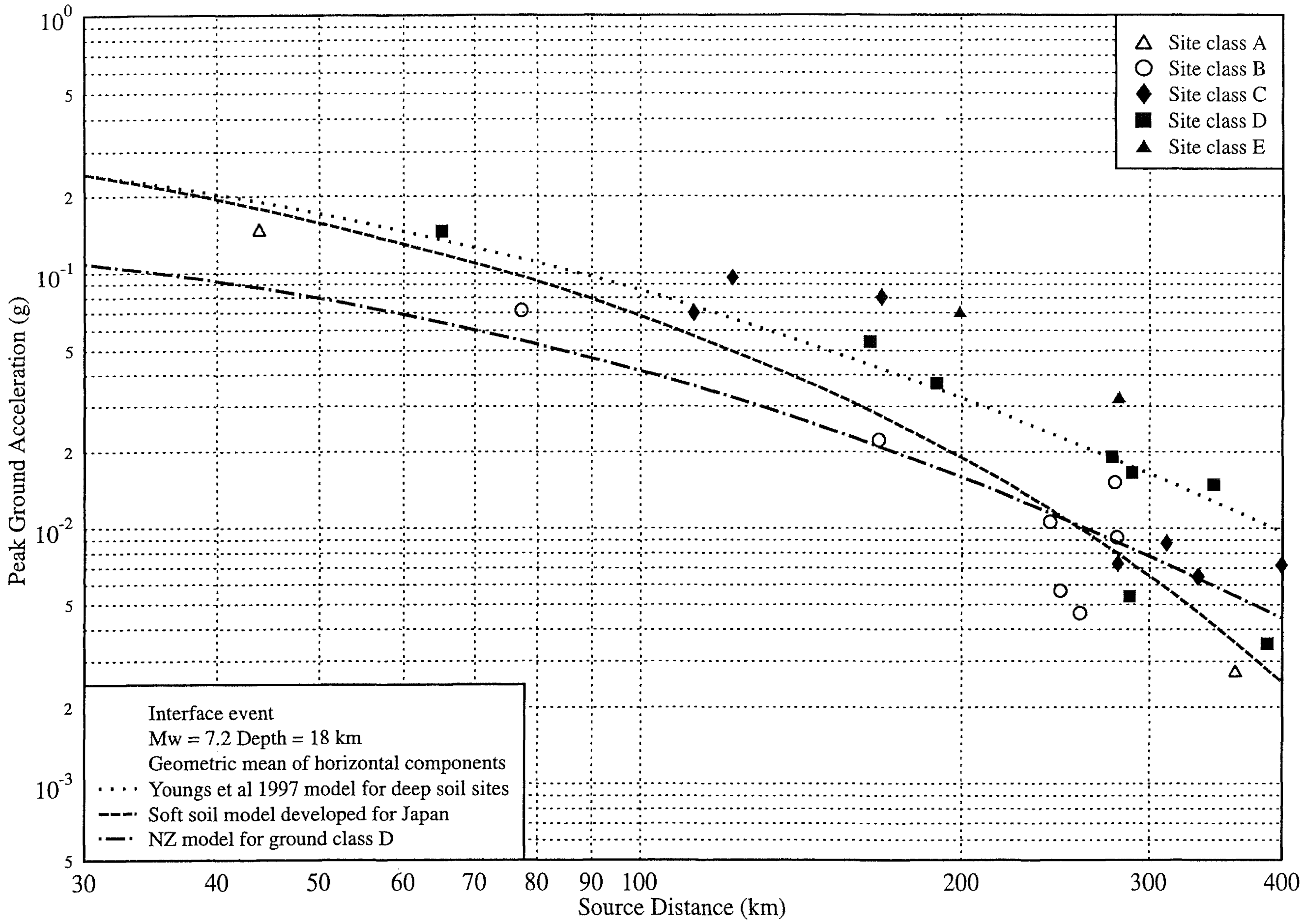


Figure 8. Comparison of the 2003 Fiordland mainshock peak ground acceleration data with attenuation models.

The site classes of the recorded pgas in Figure 8 are denoted by symbols for each of the New Zealand classes A to E, as summarised in Table 2. The peak ground acceleration attenuation curves shown in this figure are for the site classes believed to be most appropriate for New Zealand deep or soft soil sites (excluding very soft soil sites), namely for the soil class of Youngs *et al.* (i.e. "those where the depth to bedrock is expected to be greater than 20 m."), Class D (deep or soft soil) of McVerry *et al.*, and for the soft soil class of the Saiki *et al.* model. The three models use different site classification schemes, but the sites covered by their deep soil classes are reasonably similar. The New Zealand deep or soft soil class is for sites with periods in excess of 0.6 s, excluding very soft soil sites, as summarised in Table 2. The deep soil class gives the strongest peak ground accelerations for each of the models, and corresponds to the data plotted as solid squares.

It is apparent that the New Zealand model considerably underpredicts the recorded peak ground accelerations for this earthquake, by a factor of about 1.6 on average, for the data at distances up to 400 km when allowance is made for site conditions. The New Zealand model was modified from the Youngs *et al.* model to better match data from previous New Zealand subduction zone earthquakes, but for this earthquake the Youngs *et al.* model provides a much better match to the recorded peak ground accelerations. The model developed from the Japanese data is intermediate between the Youngs *et al.* and New Zealand models. The attenuation curves of the New Zealand model for 5% damped response spectral accelerations for periods of 0.5 s, 1 s and 2 s also underpredict the values recorded in this earthquake.

The New Zealand model has provided similar underpredictions of the recorded motions from the only two previous subduction interface earthquakes in Fiordland for which there are data. The M_W 6.4 Doubtful Sound earthquake of 1989 was underpredicted by an average factor of 1.7, with a factor of 1.4 for the M_W 6.8 Secretary Island earthquake of 1993. Peak ground accelerations from New Zealand subduction interface earthquakes overall are matched well by the model, although it appears underprediction of peak ground accelerations from Fiordland interface earthquakes may be balanced by a 10-15% overprediction of those from the Hikurangi subduction zone.

Fiordland subduction interface earthquakes provide only a small fraction of the overall New Zealand subduction zone earthquake data. In the New Zealand data set from which the McVerry *et al.* model was developed, 6 of the 24 subduction earthquakes were interface events, with the interface earthquakes producing 67 of the 305 subduction records used in the analysis. Only two of the interface earthquakes were from Fiordland, providing only 11 records. The 2003 Fiordland earthquake, which is the best recorded interface earthquake to date in New Zealand, has provided a further 25 records for site classes A to D at distances of less than 400 km. It is not clear whether the discrepancies arise because the nature of the earthquakes and their attenuation is different for Fiordland than for the Hikurangi subduction zone, or because of deficiencies in the scaling with magnitude. The two previous Fiordland subduction interface earthquakes, and now the 2003 event, are of larger magnitude than any of the other interface earthquakes in the dataset, for which the maximum magnitude was 5.8.

Table 2: Seismic hazard ground classification scheme as used in the draft Australia/New Zealand Loadings Standard March 2002. In New Zealand there is very little construction on "Strong Rock" and so for most purposes classes A and B will be combined.

Category	Name	Description
Class A	Strong rock	Strong to extremely strong rock with unconfined compressive strength greater than 50 MPa and an average shear-wave velocity over the top 30m greater than 1500 m/s.
Class B	Rock	Rock (i.e. material with a compressive strength between 1 and 50 MPa) with an average shear-wave velocity over the top 30m between 360 and 1500 m/s. A surface layer of no more than 3m depth of soil or highly-weathered or completely weathered rock may be included.
Class C	Shallow soil	Sites where the low-amplitude natural period is less than or equal to 0.6s [1]
Class D	Deep or soft soil	Sites where the low-amplitude natural period is greater than 0.6s [1]
Class E	Very soft soil	Sites with more than 10m of very soft soils with undrained shear-strength less than 12.5 KPA, or of soils with SPT N-values less than 6, or with more than 10m of materials with shear-wave velocities less than 150m/s.

Note:[1] The draft code and current standard NZS4203:1992 provide a table of depths of various types of soils that may be taken to correspond to the 0.6 s boundary

Another issue is that there is a very marked geographical variation in the strength of shaking in this earthquake for a given distance from the source. When the McVerry *et al.* peak ground acceleration attenuation curve is modified by the factor of 1.6 by which it underpredicts the data on average, it is found that the sites that are underpredicted and those that are overpredicted by the adjusted 50-percentile relations, with site conditions taken into account, can be separated almost perfectly by extending the western bounding line of Figure 5 to the south. Sites to the west produced stronger peak accelerations than predicted by the attenuation curve adjusted by the mean factor of about 1.6, while those to the east produced weaker motions than the adjusted prediction. Clearly, the amount of attenuation in this earthquake was governed by more than distance from the source, even when site conditions are taken into account. We plan to model the physical attenuation structure under the southern South Island to improve the modelling of motions from Fiordland earthquakes, using a similar methodology to that recently applied to mantle attenuation above part of the Hikurangi subduction zone (Eberhart-Phillips and McVerry, in press).

The four portable strong motion recorders installed in the epicentral region immediately after the mainshock have produced numerous excellent recordings of strong ground motion from the larger aftershocks. In particular, the 22 km deep M_L 6.1 aftershock of September 04 occurred almost directly beneath the recorder on Secretary Island, and produced a peak ground acceleration of 0.28g. This strong motion dataset will prove valuable in defining the near-field end of attenuation models. It will also provide information on the slip distribution in the larger aftershocks.

LANDSLIDES

Landsliding was very extensive and widespread throughout the mountainous and unpopulated epicentral regional, 50-70 km west of Te Anau. More than 400 landslides were triggered by the earthquake. These range from small superficial failures involving a few tens of cubic metres of soil and a few trees, to large rock falls and debris slides and flows extending up to ~1000 m down slope, and involving shallow bedrock and regolith (surficial soils and completely weathered rock mass). Landslides were ranked on a scale from 1 to 3, as follows:

Ranking	*Number mapped*
1 – ***Small*** – minor rock or debris falls or slides. Estimated volumes ~1- 1,000 m^3	354
2 – ***Medium*** – these included larger features with greater area, length and / or volume (~1000 to ~5000 m^3, with some possibly up to ~10,000 m^3;	53
3 – ***Large*** – landslides, usually involving some bedrock and regolith with lengths of up to 500 m or greater and widths >200 m. These larger slides had estimated volumes of ~10,000–100,000 m^3 or greater, with some 15 larger landslides or landslide areas with volumes of ~200,000–700,000 m^3	15

The mapped distribution of landslides and liquefaction effects and other damage is shown in Figure 9. The main area of landsliding extending over an area ~65 km long and 40 km wide in the mountains west of Lake Te Anau. Most failures were initiated on slopes of 35–60° or greater, with average runout slope angles of 35–50°. Regolith failures are by far the most common landslides triggered by the earthquake, most of which were first-time failures, although many were on the margins of older slide areas. The larger regolith slides (some ranked as 2 and 3, but most as 1) were apparently initiated by point failures (at ridge-top level, or at the tops of very steep slopes) involving collapse of a single boulder or 'crag', which then spread and gained mass down slope. Large regolith failures are common along the shores of Doubtful Sound, where debris has completely disappeared under water. Other large regolith failures, and some involving bedrock, have quite fluid debris run-out zones with flow patterns still preserved. Some debris runout zones bifurcate and climb over ridges bounding gullies. Very similar regolith landslides were described by Van Dissen *et al.* (1994) following the 1993 Fiordland earthquake.

Where slides involve bedrock, the most common failure mechanisms noted were translational or wedge block failures in jointed granitic rocks. Metasediments and gneisses are less affected, except where slopes are extremely steep. Some bedrock slides were still active three days after the earthquake. Bedrock plus regolith slides are common on the outer coast of Secretary Island, and around Nancy and Charles Sounds. Bedrock in these areas is shattered (probably by faulting) and more weathered, and there is more loose debris on the slopes which are being actively undercut by wave action. These slopes are not of glacial origin, and are generally less steep than slopes in the fiords (30°-40° compared to 45°-65°). The fifteen largest landslides have estimated volumes of ~75,000 to ~625,000 m^3, and details of these are given in Hancox et al. (2003). The largest landslide, at Deas Cove in Thompson Sound, extended to within ~50 m of a Department of Conservation (DOC) hut (Fig. 10). At another location a large rock wedge failure caused a seiche (wave) which damaged a wharf and shorelines in Gold Arm of Charles Sound.

Overall, the landsliding is far more significant than occurred during the M_W 6.8 1993 earthquake in the same area. Based on these effects (Hancox *et al.*, 2002; INQUA 2003) the Modified Mercalli (MM) intensities for the 2003 earthquake are estimated to have been about MM9 in the epicentral area, MM8 at Deep Cove, and MM6 to MM7 at Te Anau and Manapouri. The main area of landsliding (~3000 km^2) fits well on the magnitude/area curve for worldwide data, but is slightly above the mean regression line for New Zealand historical earthquakes (Hancox *et al.*, 2002) – presumably reflecting the very steep terrain. Although the landslide damage was widespread, the slope failures were mainly superficial. There were no deep-seated very large landslides on the scale of about 40 very large (~10^7-10^9 m^3) prehistoric (post-glacial) landslides identified in Fiordland (Hancox and Perrin, 1994). A considerably larger earthquake than that of 21 August 2003 is thought to be required to trigger such very large bedrock collapses.

LIQUEFACTION

The distribution of liquefaction was largely controlled by the distribution of susceptible materials, which mostly occur in the far field. Minor liquefaction effects (sand boils and minor lateral spreading) and slumping of unconsolidated lake sediments and alluvium were observed in several places around the shores of Lake Te Anau (Fig. 9). A small rotational failure of the lakeshore with lateral spreading and sand boils occurred a few hundred meters west of the Te Anau lake control structure, and minor spreading and some spectacular underwater sand boils occurred on the right bank of the Waiau River just downstream from the control structure. These effects caused no damage to the structure. However, lateral spreading did cause minor damage (collapse of road edges) to Hillside Road east of Manapouri where it crosses a peat swamp.

DAMAGE TO STRUCTURES AND INFRASTRUCTURE

Damage caused by the earthquake was generally minor because of its isolated location. There were, however, numerous reports of items being thrown from shelves in buildings, especially in Te Anau and Queenstown. There was also minor damage to some chimneys in Te Anau. The most spectacular - but relatively minor - damage was to the Te Anau lake control structure. The embankment fill is retained by a series of concrete slabs, which moved outward by up to 50 mm and moved vertically relative to each other by up to 75 mm. Some inter-slab clashing knocked concrete chips off. Minor cracking of concrete was also observed at the Te Anau Marina.

Landslides and rock falls generally caused only minor damage to roads in the area. The Wilmot Pass road to Deep Cove was blocked by a road cutting collapse, and there were small rock and debris falls in several places on the road to Milford Sound, but this damage was quickly cleared. Other important infrastructure components (e.g., Manapouri powerhouse and electricity transmission lines, communications) were also largely unaffected by the earthquake.

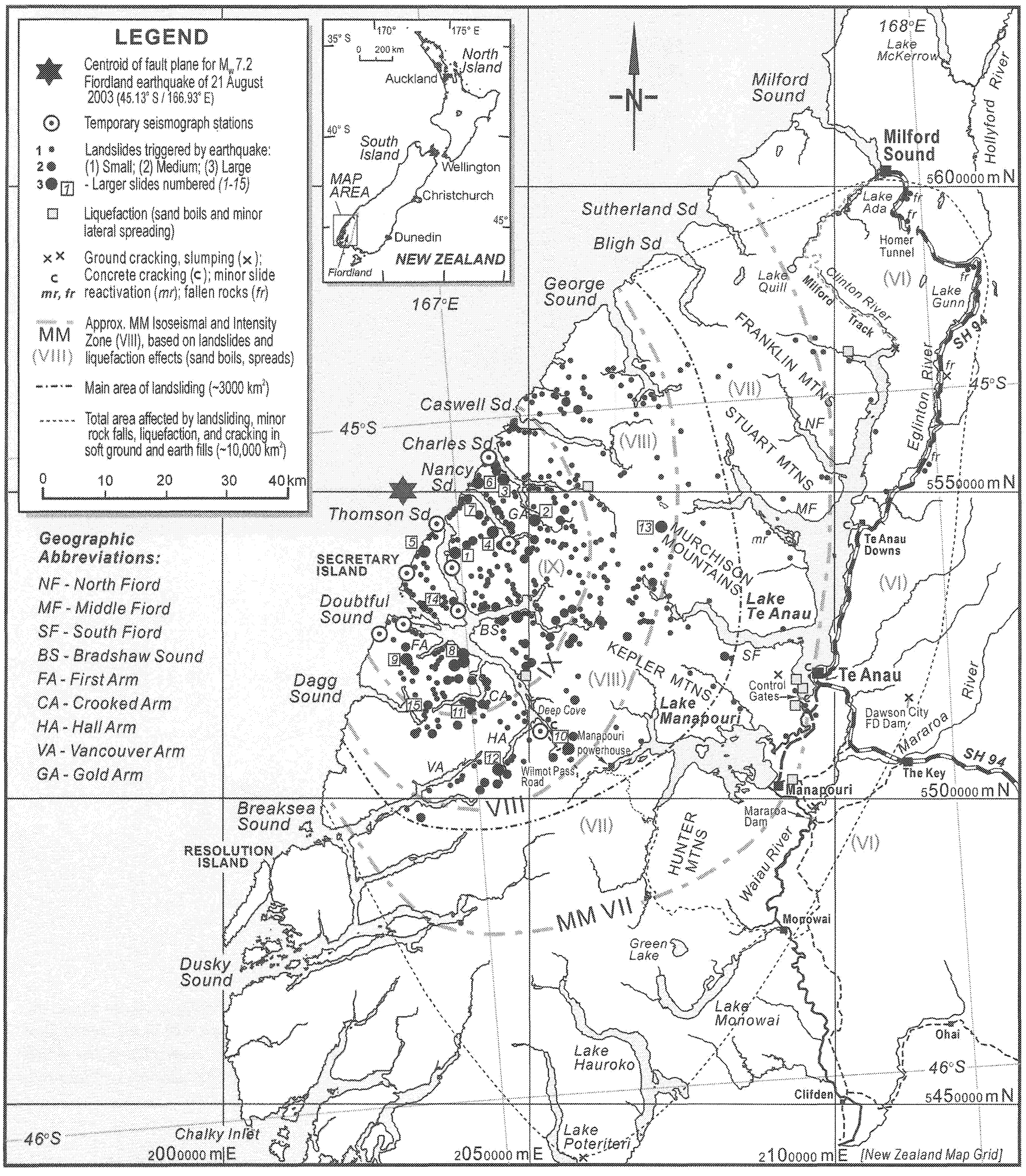


Figure 9. The distribution of landslides, liquefaction effects and other ground damage caused by the 2003 Fiordland earthquake. Numbers identify the fifteen larger (~100,000 – 700,000 m^3) landslides. Approximate Modified Mercalli (MM) isoseismals have been assigned using landsliding and liquefaction damage. The northern and southern limits of ground damage are approximate as the areas south of Breaksea Sound and Lake Manapouri, and north of Sutherland Sound were not surveyed in detail.

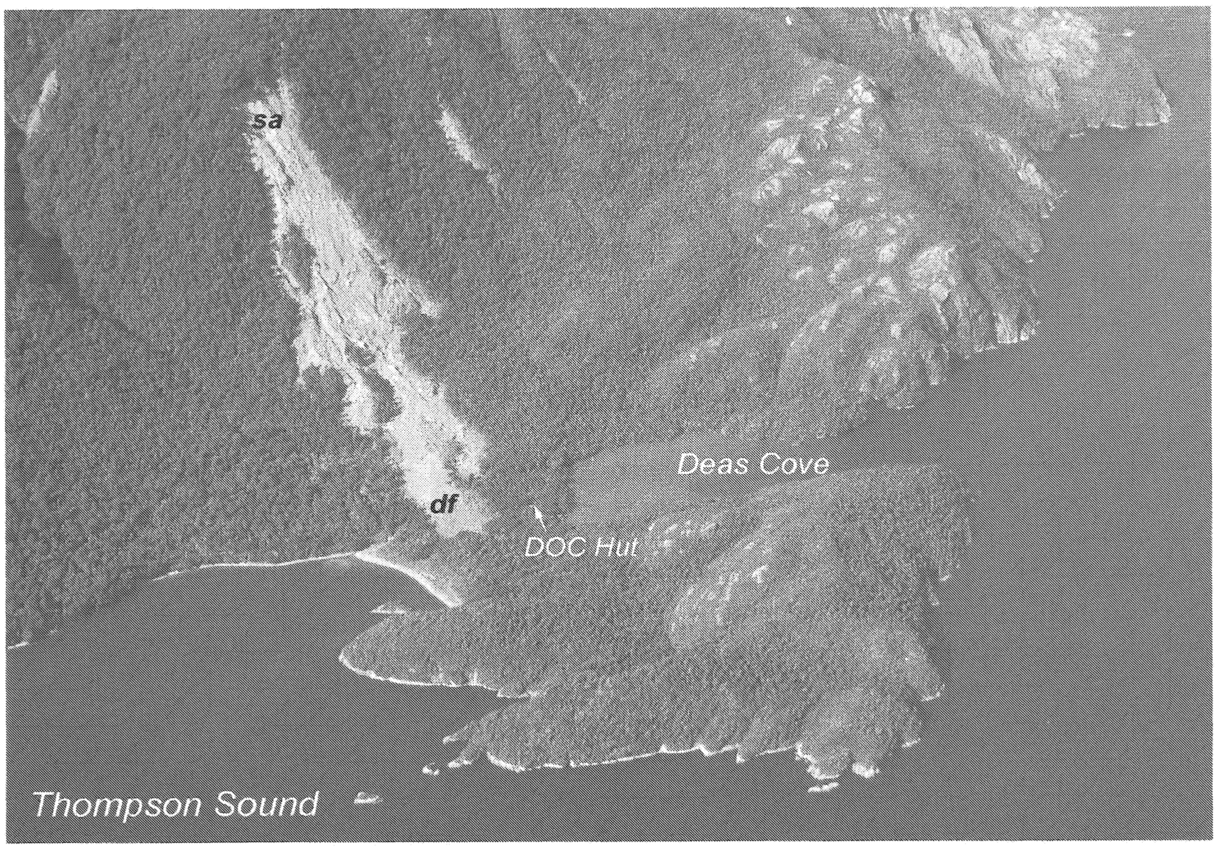


Figure 10. Aerial photos of the large (~625,000 m^3) debris slide at Deas Cove in Thompson Sound. The upper photo shows the location and overall extent of the slide. The vertical fall is about 500 m from the source area (sa) to the debris flow (df) at the slide toe, which extends to about 50 m from the DOC hut. The lower photo shows more clearly the debris flow at the toe and its close proximity to the DOC hut (in small clearing to the right of the slide toe). [Photos by G.T. Hancox, 29-8-2003]

MAGNETOTELLURIC MEASUREMENTS

Soon after the earthquake, three magnetotelluric (MT) stations were deployed in the epicentral region (Fig. 4), and these recorded for a period of two days. Each station uses three orthogonal induction coils and two perpendicular grounded dipoles to measure the surface components of the electromagnetic field.

The MT method utilises very low frequency electric currents induced in the earth by fluctuations in the strength of the earth's magnetic field to probe the electrical conductivity at seismogenic depths. It has been successfully used in the central part of the Southern Alps, where it has detected conductive zones which are thought to represent interconnected fluids. There is good geological evidence that high pressure fluids play an important role in the weakening of large faults. When the fault plane ruptures fluid trapped in the shear zone would be expected to be released, causing the electrical conductivity of the fault zone to change. We hope to re-measure the MT response at these sites periodically over the next few years to see whether measurable conductivity changes occur as the fault heals.

One noteworthy aspect of the MT data from Fiordland is that both the electric and magnetic field sensors also recorded the ground motion of the larger aftershocks that occurred during the period the MT instruments were deployed. Although there have been many claims in the literature of electromagnetic effects prior to major earthquakes elsewhere in the world, preliminary inspection of the MT time-series from Fiordland suggests that no unusual electromagnetic effects (in the 1-12 Hz frequency range) occurred immediately prior to the arrival of the elastic waves.

DISCUSSION

Despite its large magnitude, the 2003 Fiordland earthquake caused little damage to infrastructure and no loss of life, as it occurred in a remote, unpopulated and undeveloped region of New Zealand. Nevertheless, public interest in the event was high, with nearly one million hits on the GeoNet website on the day of the earthquake, and 3.8 million hits for the month of August. The rich aftershock sequence caused some concern amongst residents of Te Anau and Manapouri, as aftershocks of $M_L \geq 4.5$ were felt at these localities. To allay this concern, the Southland Regional Council organised a well-attended public meeting in Te Anau a week following the earthquake. A seismologist spoke at this meeting, explaining the tectonic context of the earthquake and the nature of its aftershock sequence.

The earthquake was the first big test of the new GeoNet network, which performed very well. In particular, the dial-in accelerographs proved very valuable in providing rapid, robust locations of the mainshock and larger aftershocks. This in turn led to accurate placement of portable seismographs, accelerographs, GPS recorders and MT instruments shortly after the mainshock. The excellent initial results from these portable deployments again attest to the value of such post-earthquake surveys.

An abiding question is what effect the 2003 earthquake has had on the southern part of the Alpine fault. Has it brought it closer to failure, or moved it further away? Further study of the data we have collected for the earthquake sequence to date will allow us to address this question.

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