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Landslides and other ground damage caused by the M_w 7.2 Fiordland earthquake of 22 August 2003

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Summary: The M_w 7.2 Fiordland earthquake of 22 August 2003 was located ~75 km northwest of Te Anau. It was the largest earthquake in New Zealand for 35 years, and involved thrusting on the subduction interface zone, not the Alpine Fault. Because of its location in an unpopulated area the earthquake caused only minor damage to buildings and infrastructure, but produced widespread superficial landsliding on steep slopes over more than 3000 km² west of Te Anau. More than 400 landslides with volumes up to ~700,000 m³ occurred during the earthquake, principally on very steep slopes within 25 km of the underlying earthquake fault plane where shaking intensity is estimated to have reached MM IX. Minor liquefaction effects (sand boils and lateral spreading) were observed in saturated sandy alluvium in many places. There were no very large bedrock failures similar to the very large (~10⁷–10⁹ m³) prehistoric landslides previously identified in Fiordland.

INTRODUCTION

On 22 August 2003 at 12 minutes past midnight NZ Standard Time, a large earthquake with a Moment Magnitude of M_w 7.2 occurred in Fiordland. The earthquake was centred near Secretary Island (45.13°S, 166.93°E), about 75 km northwest of Te Anau at a depth of ~20 km (Figure 1). It was the largest event ever recorded instrumentally in the region, and the largest shallow earthquake to occur in New Zealand for 35 years (Reyners *et al.*, 2003). Although the earthquake was strongly felt across Otago and Southland it caused relatively minor damage. Two months after the event, the Earthquake Commission had received some 2800 claims for damage with a total value of about \$11 million. It also caused numerous landslides on steep slopes in the mountains to the west of Lake Te Anau. The earthquake was the latest in a series of moderate to large events in the Doubtful Sound region in the last 15 years, including the M_w 6.7 Te Anau earthquake of 1988, and the M_w 6.8 Secretary Island earthquake of 1993 (Van Dissen *et al.*, 1994; Reyners and Webb, 2002).

To understand the relationship of the 2003 earthquake to these previous events, and nearby branches of the Alpine Fault, a 'Rapid Response' was initiated on the morning of 22 August 2003 by the *GeoNet Project* of the Institute of Geological and Nuclear Sciences (GNS). The main objectives of the response were: (i) to deploy a network of portable seismographs and strong motion instruments to record aftershocks; (ii) to resurvey GPS sites to quantify ground deformation; and (iii) to prepare a detailed inventory of landslides and other ground damage caused by the earthquake. The authors participated in the instrument deployment phase of the response and also carried out a detailed aerial and ground survey of landslides and other ground damage in the region.

In this paper we briefly describe the 22 August 2003 earthquake, its seismotectonic setting, and present a summary of the landsliding and ground damage that it caused, based on a more detailed reconnaissance report completed soon after the event (Hancox *et al.*, 2003). The landslides and liquefaction effects are described and their significance discussed in relation to earthquake-induced landsliding (EIL) in New Zealand and overseas, and other major landslide events in Fiordland.

REGIONAL GEOLOGICAL AND TECTONIC SETTING

The earthquake was located about 10 km off the coast of Secretary Island in central Fiordland, where the terrain is mountainous with relief of up to 1900 m. The area is sparsely populated with the only permanent settlements being Te Anau and Manapouri about 75 km southeast of the earthquake epicentre. The terrain is typically very steep with average slopes of 35–65° or more, having been extensively glaciated during the last ice age. Most of Fiordland is made up of strong rocks such as gneiss, schist, diorite, and granite, which are usually sparsely jointed and unweathered (Figure 1). The region also contains the remains of at least 40 very large (~10⁷–10⁹ m³) prehistoric landslides of post-glacial age (~13,000 years or less), many of which are associated with landslide dammed lakes (Hancox and Perrin, 1994). Thirteen of these landslides have volumes of ~10⁸ m³ or greater, with the largest being Green Lake Landslide (~27 x 10⁹ m³) about 50 km southwest of Te Anau (Figure 1).

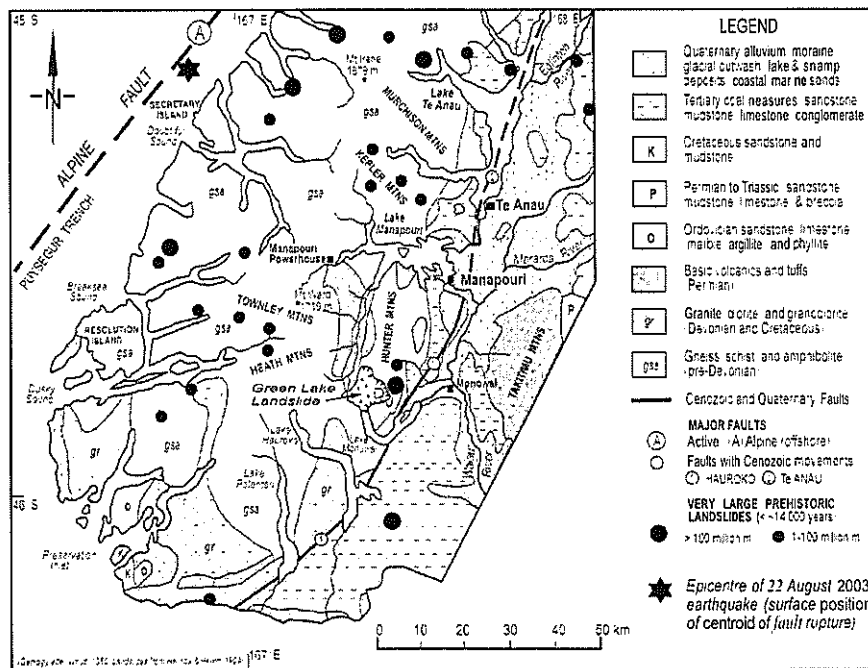


Figure 1. Map showing the location of the 22 August 2003 earthquake epicentre, the regional geology and main physiographic features of southern Fiordland, including some very large prehistoric landslides in the region.

Fiordland is one of the more seismically active parts of New Zealand, having had two earthquakes greater than magnitude 7 (1939, 1960) and seven greater than magnitude 6 in the last 150 years, four of these in the last 15 years (1988, 1989, 1993, 2000). The most significant tectonic feature in the region is the Alpine Fault, which in the Fiordland area is presumed to follow the eastern margin of the Puysegur Trench (Figure 1). A steeply dipping zone of intermediate to shallow depth earthquake activity underlying the Fiordland coast is thought to represent seismicity associated with the east-dipping subduction zone of the Australian plate moving beneath the Pacific plate. 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 at a shallow depth (< 40 km) approximately beneath the surface trace of the Alpine Fault, which is considered to be active in this part of Fiordland (Reyners and Webb, 2002), although no historical earthquakes can be attributed to it. Moderate to large earthquakes in Fiordland over the last 15 years have generally caused minor to moderate landsliding and ground damage (Turnbull and Beanland, 1988; Van Dissen *et al.*, 1994). However, no historical earthquake in Fiordland has caused large bedrock collapses on the scale of the very large prehistoric landslides in the region (Hancox *et al.*, 1997; 2002).

THE 2003 EARTHQUAKE SEQUENCE

The M_w 7.2 earthquake of 22 August and the aftershock sequence was much better recorded than the previous large earthquakes in 1989 and 1993 as a result of hazard monitoring and equipment upgrades made possible by the GeoNet Project. Figure 2 shows the position of the earthquake and the locations of temporary seismographs and strong motion recorders that were deployed, and the distribution of landslides and liquefaction-induced ground damage effects that were observed. Initial data from the deployed instruments and GPS receivers in the epicentral region soon after the earthquake have established that it was not located on the Alpine Fault, but involved thrusting at the shallow part of the underlying subduction interface (Reyners *et al.*, 2003). The earthquake had a long aftershock sequence. Up until the end of September 2003, three recorded events measured between M_L 6.0–6.2 and 18 were in the range M_L 5.0–5.9, with most events occurring near the plate interface. Post-earthquake GPS observations and some of the aftershocks define the mainshock fault plane as being about 18 km wide and 35 km long, and underlying the Fiordland coast from Doubtful Sound to Charles Sound (Figure 2) at a depth of 12–22 km, with a dip of $\sim 30^\circ$ to the southeast (Reyners *et al.*, 2003).

The GeoNet strong-motion network recorded the main shock well, providing 60 strong motion records. The closest record was from Manapouri power station (surface facility) about 44 km from the fault rupture zone (or 55 km from the epicentre), where the highest peak ground acceleration (PGA) of 0.17 g was recorded. A PGA of 0.15 g was recorded at the Te Anau Fire Station. Microzonation effects are clear in some records, with sites on deep, unconsolidated gravels showing clear attenuation of short-period motions (periods < 0.25s) and amplification of longer-period motions (periods > 0.7s) when compared to rock sites. The strong motion records also show that current PGA attenuation relationships for New Zealand subduction interface earthquakes underpredict the ground motions recorded by a factor of about 1.6 on average (Reyners *et al.*, 2003).

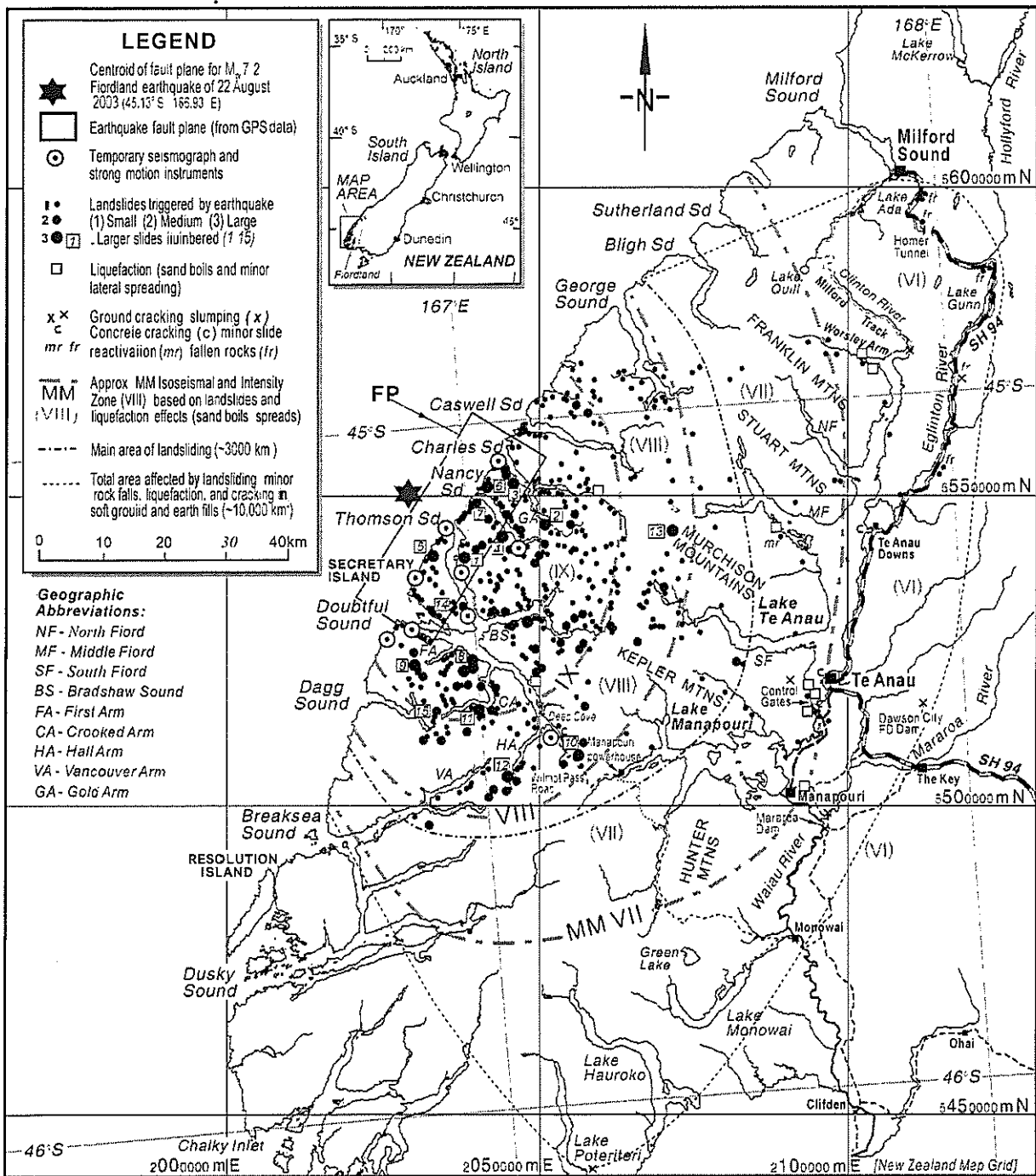


Figure 2. Map showing the location of M_w 7.2 Fiordland earthquake of 22 August 2003, and the distribution of landslides, liquefaction effects, and other ground damage caused by the earthquake. Modified Mercalli (MM) isoseismals based on environmental criteria (landsliding and liquefaction features) and the area of the underlying earthquake fault plane (FP) determined from GPS observations are also shown.

FIELD OBSERVATIONS

Landslides

The authors observed, photographed, and mapped the landslides caused by the earthquake by helicopter and ground surveys throughout the most affected areas west of Lake Te Anau (Figure 2). However, we were unable to fly over all catchments in the area, and we have yet to establish the precise southern limit of the landsliding. More than 400 individual landslides were recorded on a 1:250,000 base map, although this figure probably underestimates the exact number of slides, especially the smaller failures not easily seen from the air. The larger landslides (Figure 2) were later replotted more accurately on to 1:50,000 topographic maps from our aerial photos so that their areas and volumes could be estimated.

The landslides triggered by the earthquake 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 –1000 m or more down slope, and involving regolith (surficial soils and completely weathered rock mass) and shallow bedrock. However, most of the landslides seen were shallow superficial failures involving vegetation, soils and regolith. Landslides with volumes up to $\sim 700,000 \text{ m}^3$ were ranked on a 3-fold scale based on size [numbers in brackets], as follows: (1) Small – $< 1000 \text{ m}^3$ [35-4]; (2) Medium – $1,000\text{--}10,000 \text{ m}^3$ [53]; (3) Large – $> 10,000 \text{ m}^3$ [20-]. There were at least 15 larger landslides with volumes of $\sim 50,000\text{--}700,000 \text{ m}^3$ (Figure 2). These are described in Hancox *et al.* (2003). Notably, no large landslide dams were formed, and there were no very large, deep-seated landslides in bedrock or mountainside collapses on the scale of the prehistoric landslides discussed earlier.

The main area of landsliding in the vicinity of the earthquake epicentre extends over an area $\sim 65 \text{ km}$ long and 40 km wide in the mountains west of Lake Te Anau, and covering an area of $\sim 3000 \text{ km}^2$. Small slides, rock falls, and minor liquefaction-induced ground damage, however, extends over almost $10,000 \text{ km}^2$ (Figure 2). Most landslides were initiated on slopes of $35\text{--}60^\circ$ or greater, with average mnout slopes of $35\text{--}50^\circ$. Regolith slides were mostly first-time failures, although many were on the margins of older slide areas. Where slides involve bedrock, the most common failure mechanisms noted were translational or wedge block failures in strong jointed granitic and gneissic rocks. Bedrock plus regolith slides are common on the outer coast of Secretary Island, and in Nancy and Charles sounds. Bedrock in these areas is shattered (by faulting) and more weathered, and there is more debris on slopes which are being actively undercut by wave action. Not being of glacial origin, these coastal slopes are usually less steep than those in the fiords ($\sim 30\text{--}40^\circ$ compared to $\sim 45\text{--}65^\circ$).

One of the largest bedrock and regolith landslides occurred at Deas Cove in Thompson Sound, where slide debris travelled $\sim 1150 \text{ m}$ down a $35\text{--}40^\circ$ slope to within 50 m of a Department of Conservation hut (Figure 3). In Gold Arm of Charles Sound, a large rock wedge slide in gneissic bedrock caused a $1\text{--}2 \text{ m}$ high tsunami (wave) which damaged a nearby helipad, and shorelines on the opposite (northern) side of the sound (Figure 4).



Figure 3. Aerial photo of the large ($\sim 625,000 \text{ m}^3$) debris slide at Deas Cove in Thompson Sound (Slide 1 in Figure 2). The landslide is $\sim 1150 \text{ m}$ long, and has a vertical fall of 500 m from the source area (sa) to the debris flow (df) at the slide toe. The landslide toe extends to within 50 m of the DOC hut (in small clearing bottom right) but did not damage it. Air photos taken in the 1980's indicate that this large slide in weathered granite and regolith is a reactivated and retrogressively enlarged former landslide area.

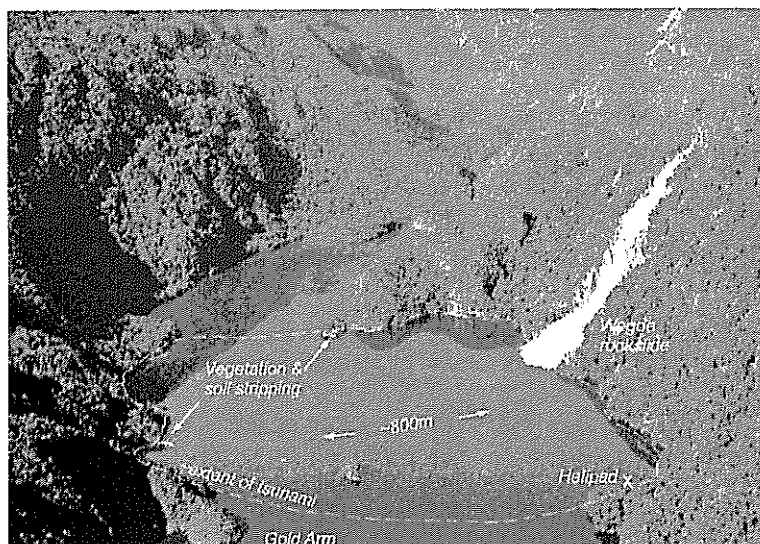


Figure 4. Aerial view of the large ($\sim 200,000 \text{ m}^3$) rock fall/wedge slide in Gold Arm of Charles Sound (Slide 2 in Figure 2). The vertical fall is $\sim 350 \text{ m}$ on a 40° slope. This slide caused a small tsunami (wave) which travelled $\sim 800 \text{ m}$ across the sound, where it damaged vegetation and soil up to $4\text{--}5 \text{ m}$ above high tide level on the northern side, and stripped vegetation off a small island. It also damaged a helipad $\sim 250 \text{ m}$ west of the slide (bottom right). Areas of weathered rock adjacent to the slide, and a pre-existing debris cone at the site, show this is also an old slide area that was reactivated and enlarged by the earthquake.

There was also clear evidence of aspect-controlled landsliding on some slopes. The multiple rock and debris slides in the outer parts of Nancy Sound (Figure 5) showed preferential landsliding on north to northwest facing slopes. Preliminary field mapping shows that this was general across the area most affected by landsliding, with approximately 62% of the landslides occurring on north and west-facing (** E-SSW) slopes and fewer slides on east and south-facing slopes (Figure 6).

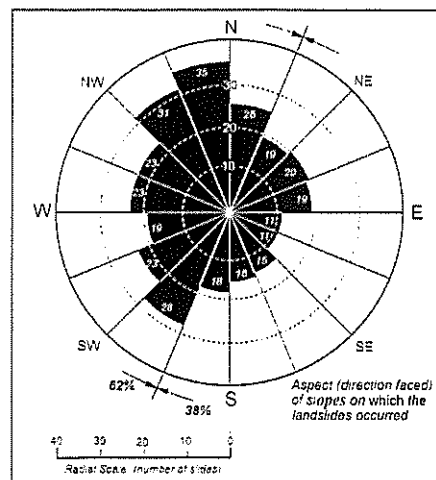
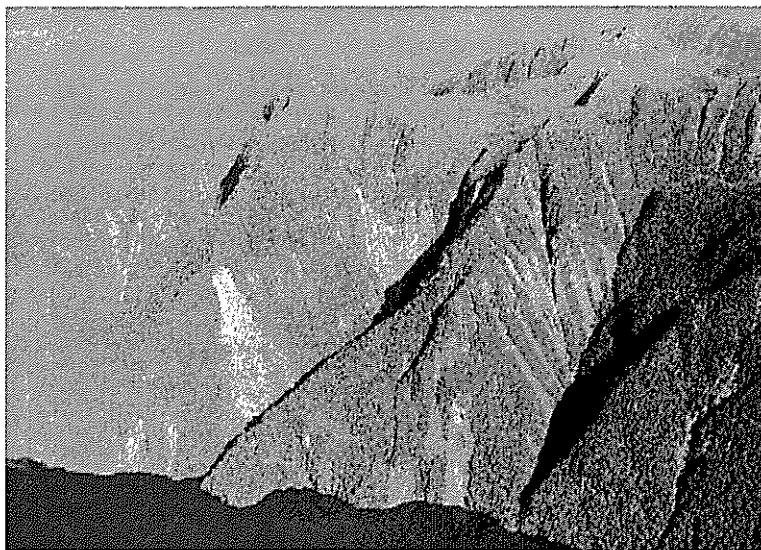


Figure 6. Rose diagram showing landslide directions relative to slope aspect.

Figure 5. Aerial photo showing large rock fall debris slide (centre left—Slide 4 in Figure 2) and multiple shallow narrow debris slides on steep (45-50°) slopes of side streams on the south side of Nancy Sound. Such multiple small slides were common in the epicentral area. The preferential landsliding seen here on northwest-facing slopes was general across the landslide affected area (see Figure 6).

Liquefaction

Soil liquefaction effects (sand boils and small-scale lateral spreading) were observed in many places, particularly at the southwest end of Lake Te Anau (Figure 2). Although widespread, these effects were mostly localised in areas of saturated fine sandy alluvium and were generally minor compared to other large earthquakes in New Zealand (e.g., Napier, 1931, Inangahua 1968, and Edgcumbe 1987, – Hancox *et al.* 1997). A spectacular liquefaction feature was observed on the southwest shore of Lake Te Anau, about 400 m west of the Te Anau Control Structure (Control Gates in Figure 2). At this location a liquefaction-induced lateral spread and shallow rotational slump uplifted an area of lake shore about 20 m long and 15 m wide (Figure 7). This failure is interesting because of the combination of rotational and lateral spreading failure modes that are demonstrated, although it did not cause any significant or long-lasting damage to the lake shore. The coarse alluvium in which failure occurred suggests that finer and more susceptible sandy materials underlie the beach gravel deposits at this location. A similar feature with an unusual cluster of underwater sand boils occurred on the right bank of the Waiau River –100 m downstream from the Control Structure.

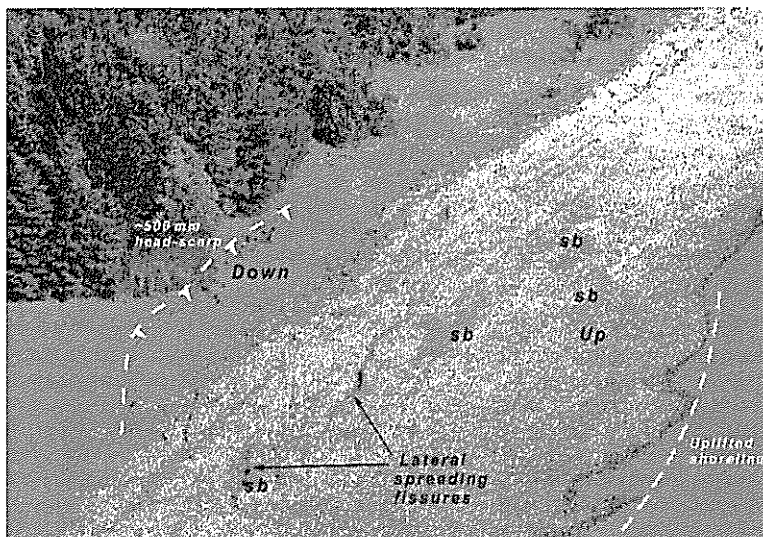


Figure 7. Shallow rotational slump and lateral spreading failure, with sand boils (*sb*) and fissuring on the shore of Lake Te Anau about 400 m west of the lake Control Structure. The affected area is –20 m long and 15 m wide, and has a 500 mm head scarp (in grassed area left) and several extensional fissures (300–400 mm wide and up to –10 m long) in beach gravels, from which fine sand and water was ejected. The back of the slumped area dropped –500 mm, while the slide toe at the lake edge was raised a similar amount, forming a low ridge on which sand boils formed.

Other liquefaction and soft sediment deformation effects observed included spreading and slumping of road edges along Hillside Road where the road crosses a swampy area ~2.5 km east of Manapouri. Very minor liquefaction effects were observed elsewhere around the shores of Lake Te Anau. At Glade House wharf, near the southern end of the Milford Track, there was some subsidence of sandy gravel around piles and pull-apart cracking (lateral spreading) 1–3 m back from the lake edge. Other small delta collapses were noted in Middle Fiord and Worsley Arm of Lake Te Anau (Figure 2).

Damage to structures, infrastructure, and other earthquake effects

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. The most spectacular, but relatively minor, damage occurred at the Lake Te Anau Control Structure. The embankment fill of this structure is retained by a series of concrete slabs, that moved outward by up to 50 mm, vertically relative to each other by up to 75 mm, and had concrete chips knocked off in several areas. Damage to buildings and other structures in the Te Anau–Manapouri area was also minor, and included cracking of chimneys, fireplaces, masonry and foundations in Te Anau, and cracks and breaks in concrete at the Te Anau Marina. Small (mm-wide) cracks were also observed in abutment earth fills at the Mararoa Dam ~10 km southeast of Manapouri, and at the Dawson City flood detention dam 15 km east of Te Anau (Figure 2). 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 these falls were quickly cleared. Other important infrastructure components (e.g., Manapouri powerhouse, electricity transmission lines, communication links) in the Te Anau – Manapouri area were largely unaffected by the earthquake.

DISCUSSION

Our studies after the earthquake have clearly shown that superficial landsliding was indeed widespread, but the shaking intensity was evidently below the threshold required for very large bedrock failures and mountain collapses in Fiordland. As expected, the landslides triggered by the 22 August earthquake decreased in number and size with distance from the epicentre. Figure 2 shows that the greatest density of landsliding, and indeed many of the larger landslides, occurred within 25 km of the area of the mainshock fault plane determined from GPS observations. Near the epicentre a few small 3rd-order alpine catchments of about 5 km² in area had up to 20% of the slopes affected. High-density landsliding also occurred on the steep coastal slopes nearest the epicentre. In general, however, the landslide density in the entire affected area was low, with the largest percentage of the affected area having only isolated, individual slides. In alpine and many coastal catchments with the highest density of slides, there was a distinct locational preference for the northerly facing slopes, with 62% of the landslides occurring on north and west (N–SSW) facing slopes (Figure 5 and 6).

Slope aspect preference has been noted in a number of multiple landslide events in the past (Crozier *et al.*, 1980) with several explanations offered for this. These can include directional rainfall, or in the case of earthquake-induced landsliding the directional transmission of earthquake waves with respect to topography. However, the latter does not appear to be the case for the August 2003 earthquake as preferential sliding on northerly slopes is general across the affected area. Another possibility is that 'sunny' slopes undergo a greater frequency and range of thermal expansion and wetting and drying cycles, leading to a weakening of joints and more rapid regolith production on sunnier north-facing slopes than on shady slopes. Sunny slopes may also have thicker vegetation, which on steep slopes could make them more prone to failure during earthquakes because the soil mass cannot support the vegetation cover when strongly shaken. Further studies of site conditions and landslide distribution relationships using GIS techniques are planned to further explore this issue.

Overall, the landsliding is far more significant than what occurred during the 1993 M_w 6.8 earthquake in the same area. Based on environmental criteria used for assessing earthquake intensities (INQUA, 2003; Hancox *et al.*, 1997; 2002), the number and size of landslides triggered by the August 2003 earthquake suggest that the Modified Mercalli (MM) shaking intensity in the epicentral area was about MM IX, while at Deep Cove it was MM VIII, and at Te Anau and Manapouri between MM VI and MM VII (Figure 2). This is generally consistent with intensities indicated by damage to buildings and structures in the Te Anau area, and also the PGA values recorded by strong motion instruments at Deep Cove (0.17 g) and Te Anau (0.15 g). The main area of landsliding (~3000 km²) fits well on the magnitude/area curve for worldwide data (Keefer, 1984), but is slightly above the mean regression line for historical earthquakes in New Zealand (Hancox *et al.*, 2002) – presumably reflecting the very steep terrain in Fiordland. Although the landslide effects were widespread, the slope failures were mainly superficial. There were no deep-seated landslides similar to the 40 very large (~10⁷–10⁹ m³) prehistoric (post-glacial) landslides identified in Fiordland (Hancox and Perrin, 1994). A considerably larger earthquake than that of 22 August 2003, probability events centred on the Alpine Fault, is thought to be required to trigger such very large bedrock collapses.

CONCLUSIONS

- (1) The M_w 7.2 earthquake in Fiordland on 22 August 2003 caused widespread superficial landsliding over more than 3000 km² across the unpopulated mountains 50–70 km west of Te Anau. Minor liquefaction effects (sand boils and small-scale lateral spreading) were also widespread but non-damaging. More than 400 landslides with volumes ranging up to $\sim 700,000$ m³ were triggered, principally on very steep slopes within 25 km of the earthquake fault plane where shaking intensity is estimated to have reached MM IX. There were no deep-seated failures like the very large prehistoric bedrock slides and mountainside collapses previously identified in Fiordland.
- (2) Landslides and rock falls caused only minor damage to the few roads in the area, with a cutting collapse closing the Wilmot Pass road, and a number of small rock and debris falls on SH 94 to Milford Sound. Lateral spreading caused minor collapses of road edges on Hillside Road east of Manapouri. Other infrastructure components in the region (e.g., Manapouri powerhouse and electricity transmission lines, communications) were largely unaffected.
- (3) Our initial landslide reconnaissance mapping has shown some interesting relationships between landslide development and slope aspect and steepness, with preferential failure on north to southwest-facing slopes clearly indicated. Further studies using GIS are planned to examine this relationship in more detail.

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