MICROSEISMICITY IN THE CENTRAL
SOUTHERN ALPS, WESTLAND,
NEW ZEALAND

by

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“Der moderne Mensch kann ins All fliegen, tödliche Krankheiten besiegen und in sekundenschnelle gigantische Datenmengen um den Globus schicken.
Doch wenn die Erde bebt, ist er so machtlos, wie seine Vorfahren ...”
Gewaltige Natur, *Der Spiegel*, 2011

Abstract

Present-day seismicity associated with the central Alpine Fault and the zone of active deformation and rock uplift in the central Southern Alps is reported in this thesis. Robust hypocentre locations and magnitude estimates for ~ 2300 earthquakes have been obtained analysing 18 months of data from the Southern Alps Microearthquake Borehole Array (SAMBA), designed for this study. The earthquakes are distributed between the Alpine Fault and the Main Divide Fault zone and confined to shallow depths (90% of events ≤ 12.2 km). The thickness of the seismogenic zone follows lateral variations in crustal resistivity: earthquake hypocentres are restricted to depths where resistivities exceed $390 \Omega \text{ m}$. Rocks at greater depth are interpreted to be too hot, too fluid-saturated, or too weak to produce detectable earthquakes. A low-seismicity zone extends between the Whataroa and Wanganui rivers at distances 15–30 km southeast of the fault, which is concluded to be a relatively strong, unfractured block that diverts deformation around it. A new magnitude scale is developed incorporating the effects of frequency-dependent attenuation, which enables magnitudes to be calculated consistently for earthquakes of different sizes and frequency contents. Focal mechanism solutions for 379 earthquakes exhibit predominantly strike-slip mechanisms. Inversion of these focal mechanisms to determine the prevailing tectonic stress field reveals a maximum horizontal compressive stress direction of $115 \pm 10^\circ$, consistent with findings from elsewhere in South Island. The 60° angle between the strike of the Alpine Fault and the direction of maximum horizontal compressive stress suggests that the Alpine Fault is poorly oriented in an Andersonian sense. Earthquake swarms of at least 10 events with similar waveforms frequently occur within the region, of which some were remotely triggered by two major South Island earthquakes. Focal mechanisms of the largest event in each swarm ($M_L \leq 2.8$) reveal at least one steeply-dipping nodal plane ($\geq 50^\circ$) and one well-oriented nodal plane in the tectonic stress field. The swarms exhibit a distinctly different inter-event time versus duration pattern from that of typical mainshock-aftershock sequences. The triggered seismicity commences with the passage of the surface waves, continues for ~ 5 and ~ 2 days, and is followed by a quiescence period of approximately equal length. Remotely triggered swarms occur delayed by several hours and their delay and locations are consistent with fluid diffusion from a shallow fluid reservoir. Estimated peak dynamic stresses

(≥ 0.09 MPa) imposed by the surface waves are comparable to observations of triggering thresholds (> 0.01 MPa) elsewhere. The triggered swarms have no apparent differences from the background swarms, and appear to have been clock-advanced. Tectonic tremor in the vicinity of the Alpine Fault coincides with a low-velocity, high-attenuation zone at depth. The tremor occurs at the downdip extension of the Alpine Fault and in the region where bending of the Australian and Pacific plates is largest at depths spanning 12–49 km. Similarities with tremor occurring on the San Andreas Fault near Cholame in terms of tremor duration, depth, spatial extent and amplitude distribution, imply property variations in the lower crust and upper mantle along the strike of the Alpine Fault.

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1. Introduction

1.1. Motivation, thesis objectives and outline

This PhD project is part of the Marsden-funded project “Putting a stethoscope on the Alpine Fault”, which sought to address the following topics

- the accommodation of strain due to plate movement in the mid and upper crust of the central South Island;
- the generation of earthquakes and other strain signals such as slow-slip events and non-volcanic tremor in a fluid-rich, continental transform fault zone; and
- the distribution of strain in both time and space, especially at depths below the brittle-ductile transition zone.

To investigate these topics the Southern Alps Microearthquake Borehole Array (SAMBA) was deployed, a station network designed for the detection of microearthquakes in the central Southern Alps. We note that the terms “array” and “network” are used interchangeably throughout this thesis. Despite this, “array” is more commonly used for small station networks (with inter-station distances <20 km).

In addition to the station installation, service and maintenance of the SAMBA network, an analysis and interpretation of the recorded earthquake data has been performed in the course of this project. Earthquake activity at the microseismicity level ($M_L < 3$) is investigated in order to distinguish between structural and fluid-driven processes that generate earthquakes. Temporal and spatial seismicity patterns and the deformation and rheology associated with them are studied.

This thesis begins with an introduction to the Alpine Fault within the central Southern Alps region, its broader tectonic setting and the seismotectonic activity revealed by recent seismicity studies. An overview of the SAMBA station network, the data processing and the network performance is also provided. Chapter 2 outlines the theoretical background

to the methods applied to the data. Chapter 3 focuses on the analysis of the data and tests performed with the data. A summary and comprehensive discussion of the earthquake locations, magnitudes and focal mechanisms in the prevailing stress field from the first 1.5 years of data is given in Chapter 4 which has been recently published in the *Journal of Geophysical Research* (Boese et al. 2012). This chapter has been modified regarding references to other figures and chapters to better fit the context of this thesis. Repetition of some of the material presented elsewhere in this thesis may occur but this has been kept to a minimum. Chapter 5 characterises swarms and remotely triggered seismicity in the central Southern Alps. This chapter consists of two parts: a general introduction to swarms, observations and suggested mechanisms of triggering and a discussion of the observations of swarms in the study area and their implications. The second part has been written for submission as a publication. Some of the results incorporated were obtained by Katrina Jacobs as outlined in the introduction to the chapter. Chapter 6 summarises unpublished results of different topics that were investigated, specifically deep earthquakes, tremor, tomography, fault-guided waves and strain release by microearthquakes. The tremor analysis was performed together with Aaron Wech. His contributions are described at the beginning of this chapter. The concluding chapter, Chapter 7, summarizes the key findings of the research and discusses their implications. It also provides topics and suggestions for future work with longer records of seismic data from the SAMBA network.

1.2. Tectonic and geological setting of the Alpine Fault

New Zealand is situated on the plate boundary between the Pacific and Australian plates. The Pacific Plate is subducted westwards beneath the Australian Plate along the Tonga-Kermadec Trench and its southern continuation, the Hikurangi Trough, which extends from East Cape of the North Island to the northernmost part of the South Island. Eastward-dipping subduction of the Australian plate occurs at the southern part of South Island along the Puysegur Trench beneath Fiordland and offshore. The Alpine Fault, together with the Marlborough Fault system, links these subduction zones of opposite polarity (Berryman et al. 1992). The Alpine Fault forms the western boundary of the Southern Alps, which are the result of oblique continental collision between two largely submerged continental fragments, the Chatham Rise and Campbell Plateau to the east and the Challenger Plateau to the west (Fig. 1.1) (e.g. Cox and Sutherland 2007).

The New Zealand land mass encompasses different geological terranes of varying age (Mortimer 2004). The basement rocks of the Western Province, extending from Fiordland, along the West Coast to west Nelson (Fig. 1.1) formed from sediment deposits along the margin of the Gondwana Supercontinent during the Palaeozoic. These sediments were metamorphosed and intruded by igneous rocks (Devonian–Cretaceous granites and granitoids) when rifting occurred in the Gondwana break-up phase. The Western Province basement rocks span an age range from Precambrian to late Mesozoic (Cox and Barrell 2007).

Eastern Province rocks, known as the Torlesse Supergroup, which can be found over much of the country from Otago to East Cape, formed from the deposition of sediments (sandstone, mudstone and volcanic rocks) in turbidity currents along the eastern coastline of Gondwana. These sedimentary rocks are Permian to Jurassic in age (Mortimer 2004). In the South Island, they are now being rapidly uplifted along the Alpine Fault. Their metamorphic grade increases from east to west with the once most deeply buried rocks (oligoclase and garnet zone from 20–30 km, Kamp et al. 1989; Cooper 1980; Little et al. 2002; Norris and Cooper 2003) exposed at the surface adjacent to the Alpine Fault. A distinct suite is the Dun Mountain-Maitai Terrane, of originally volcanic rocks that were deposited in discrete pulses of magmatism in the subduction back-arc. These rocks are Carboniferous to Early Cretaceous in age and were highly deformed during the Cretaceous and Late Cenozoic (Mortimer et al. 1999). The schist in different regions on the South Island is referred to as Otago, Alpine and Marlborough Schist despite the lack of formally

Figure 1.1. Regional geological map highlighting the plate boundary between the Australian and Pacific plates through New Zealand (from Officers of the New Zealand Geological Survey 1972 with limited updates by P.J. Forsyth and N. Mortimer (2004) and GNS Science Active Faults Database). Key locations referred to in the text are shown.

defined boundaries between them (Cox and Sutherland 2007).

Eastern and Western Province rocks in South Island are separated by the Alpine Fault and its northern continuation, the Wairau fault. They displace the Dun Mountain Ophiolite Belt by a total of 460 km (e.g. Sutherland 1999). This feature can be accurately traced on the North and South Islands due to its characteristic magnetic properties and is therefore a distinct geologic marker of plate motion. However, the total amount of dextral displacement of the western North Island with respect to the eastern South Island of New Zealand from plate model reconstruction is estimated to be 850 ± 100 km (Molnar et al. 1999). The extent of distributed dextral shear deformation of the Dun Mountain-Maitai Terrane before or during the inception of the Alpine Fault is still debated (Sutherland 1999, and references therein). Mesozoic to late Cenozoic ages for the initiation of deformation on the Alpine Fault have been proposed, but Eocene ages for the localisation of strain on the plate boundary are favoured by Sutherland (1999). Sutherland (1999) and Sutherland et al. (2000) suggested that this major fault evolved from reactivation of a zone of pre-existing fractures, inherited from extensional deformation related to sea-floor spreading south of New Zealand in the Eocene. According to these authors, the Alpine Fault formed during the Miocene when the displacement rates increased significantly in response to a change in the plate motion direction.

Uplift of the Southern Alps started in the late Miocene/early Pliocene when oblique compression of the relative plate motion between the Pacific and Australian Plates increased (e.g. Walcott 1978). Cande and Stock (2004) argued that the onset of convergence occurred in the northern section of the Alpine Fault about 11 Ma in contrast to the southern section where no convergence occurred prior to 11 Ma. Increasing convergence rates between 11 Ma and 6 Ma led to shortening in the central South Island by a maximum amount of 90–110 km (Walcott 1998). The earliest appearance of overthrusts of Haast Schist over Pleistocene gravels is dated at 2.5–3.0 Ma (Rattenbury 1986). These gravels indicate an increase in plate boundary convergence rate rather than the onset of convergent movement on the Alpine Fault (Walcott 1978; Yeats and Berryman 1987). Estimates of the total uplift during the last 5 Myr (the Kaikoura Orogeny) range between 21.5 km (Kamp et al. 1989) and 25–30 km (Cooper 1980; Little et al. 2002; Norris and Cooper 2003).

The Alpine Fault can be divided into three segments (e.g. Sutherland et al. 2000) which differ significantly in their dips and structural styles, and in the widths of deformation adjacent to the fault (Fig. 1.2). In the northernmost part of South Island, slip-partitioning occurs between the Hikurangi subduction zone and several major strike-slip faults (Little and Jones

Figure 1.2. The three segments of the Alpine Fault and its variation in style and dip (modified from Sutherland et al. 2006).

1998), known collectively as the Marlborough Fault System. The faults traverse the northernmost South Island and merge with the Alpine Fault in the southwest. This broad zone of deformation has a width of about 200 km. The four major faults, Wairau, Awatere, Clarence, and Hope Fault, are steeply-dipping, evenly-spaced approximately 30 km apart, and have parallel strike directions (Little and Jones 1998; Grapes et al. 1998). The average regional strike of the Marlborough Faults is 055° in the northernmost South Island but changes to 070° further south (Lamb and Bibby 1989). The Marlborough Faults become progressively younger from the north to the south (Little and Jones 1998, and references therein). The northernmost part of the Alpine Fault, named Wairau Fault, is the oldest (early Miocene age) of the Marlborough faults. It currently has the lowest displacement rates of the Marlborough Fault System, which are estimated to be 4 ± 1 mm/yr (Knuepfer 1992). The Hope Fault is the southernmost and youngest (Pleistocene age) active fault. It has the highest slip-rates of 23 ± 4 mm/a (Langridge et al. 2003). Approximately 60 km south of the Hope Fault, the Porters Pass–Amberley Fault Zone (PPAFZ) may form the incipient continuation of the fault system to the south (Cowan et al. 1996). The PPAFZ consist of several smaller fault segments, including the Porters Pass Fault, a young active, 40 km-long segment, northeast of Lake Coleridge (Cowan 1992, Fig. 1.3 and 1.6). The faults that

Figure 1.3. Map of South Island showing key locations referred to in this introduction. Focal mechanism solutions for the Dusky Sound (Fry et al. 2010), Darfield and Christchurch earthquakes (USGS) are shown. The blue transect shows the locations occupied during magnetotelluric sounding.

ruptured in the Christchurch earthquake sequence 2010/11 approximately 30 km south of the PPAFZ (e.g. Gledhill et al. 2011; Quigley et al. 2012) involved several fault segments, e.g. the east-west striking ($\sim 089^\circ$) Greendale Fault and inferred faults striking $55\text{--}65^\circ$ and 68° for the major Christchurch events](Quigley et al. 2012; Sibson et al. 2011; Beavan et al. 2011; Holden 2011).

Slip rates in the Marlborough Fault System show that about 95% of the plate motion is accommodated by the faults (Holt and Haines 1995). Primarily, dextral strike-slip faulting occurs. The amount of dip-slip is small but increases to the northeast, causing uplift of the coastal Kaikoura ranges (Davey et al. 2007).

With the transition from westward subduction to continental collision, a change from predominantly strike-slip to convergent motion takes place on the central section of the Alpine Fault. This transition zone coincides closely with the region in which the Wairau

and Awatere faults splay off from the Alpine Fault (Davey et al. 2007) and is characterised by the abrupt termination of earthquakes at depths greater than 40 km (Anderson and Webb 1994). A line between Westport and Cheviot (Fig. 1.3) has been suggested to mark the present-day southern termination of the subducting Pacific Plate (Robinson 1991). Reyners and Robertson (2004) argued that subduction of the Pacific Plate extends south to -43° and Reyners et al. (2011) suggested that it extends even further southwards beyond -45° . The crust of the Pacific Plate changes from thick oceanic to thin continental crust, which is inferred to be associated with a change in dehydration conditions within the slab (Reyners and Robertson 2004). The fault dip inferred from seismic reflection studies (Stern et al. 2007) and from the surface dip of the mylonite foliation (Sibson et al. 1981) is estimated to be 60° . Strong reflections at a depth of 33 km are consistent with a subhorizontal fault (Kleffman et al. 1998), indicating a listric geometry with depth.

In the central South Island, the zone of deformation is only 80 km wide (Cox and Barrell 2007). The width of the Alpine schist is narrowest (8 km between Franz Josef and Fox Glacier versus a maximum of 26 km; Little et al. 2005; Cox and Sutherland 2007). According to Norris and Cooper (2001), the central Alpine Fault accommodates 75% of the total plate motion in horizontal movement and up to 100% of the convergence. Both thrust and strike-slip motion are accommodated on a single fault (Norris and Cooper 2001). The surface trace of the Alpine Fault is characterised by serial partitioning into 3–5 km-long segments of strike-slip and reverse segments (Norris and Cooper 1997; Norris and Cooper 2001). Sections of north-east striking ($020\text{--}050^{\circ}$), oblique thrust segments alternate with more easterly ($065\text{--}090^{\circ}$) striking dextral strike-slip segments (Norris and Cooper 1995). Shallow overthrusts of the thrust segments generally extend less than a few kilometres west of the main Alpine Fault trace (Norris et al. 1990). Sandbox-modelling of the fault by Norris and Cooper (2001) suggests that the strike-slip and thrust segments represent the surface structure of a single fault at depth. According to Norris and Cooper (2001), this serial partitioning at the surface results from the perturbation of the stress field by valleys crossing the fault in combination with an oblique-slip ductile fault zone at depth.

In the southern section, between Jackson Bay and Milford Sound (Fig. 1.3), the Alpine Fault is steeply-dipping ($70\text{--}90^{\circ}$) and purely strike-slip in character (Sutherland et al. 2000). About 55–85% of the total plate motion is accommodated on the Alpine Fault. Approximately 10 mm/yr of right-lateral movement is accommodated by active faulting in a 300 km-wide zone southeast of the Alpine Fault (Sutherland et al. 2000). Several faults, notably the Hollyford Fault System and the Moonlight–Siberia Fault System (White 2002), are located to the southeast of the Alpine Fault. The convergent component of plate motion

is partially accommodated by thrust faulting west of the Alpine Fault. A change in the strike direction of the Alpine Fault by 7° counterclockwise occurs south of the Cascade River (Sutherland et al. 2007). The strike-slip motion, inferred from surface displacements, is 23 ± 2 mm/yr (Sutherland 1994; Sutherland et al. 2007). South of Caswell Sound, the fault continues for 230 km offshore and ends near the Puysegur trench (Barnes et al. 2005). The fault trace is continuous but geometrically complex, with right-stepping segments and small pull-apart basins (Barnes et al. 2005). Slip rates determined from displaced submarine, glacial geomorphologic features span 24–33.5 mm/yr (Barnes 2009). This accounts for 90% of the total relative plate motion and indicates a high degree of strain partitioning between the Fiordland subduction zone and the southern Alpine Fault (Barnes 2009).

1.2.1. Characteristics of the central Southern Alps region

The present-day relative plate motion between the Australian and the Pacific plate from the NUVEL-1A model is 37 ± 2 mm/yr at an azimuth of $071 \pm 2^\circ$ (De Mets et al. 1994). The Alpine Fault strikes at an angle of approximately 18° to the plate motion direction, so that the boundary-parallel component amounts to 35.5 ± 1.5 mm/yr and the fault-perpendicular component to 10 ± 1.5 mm/yr at $-43.5^\circ/170.5^\circ$ (Little 2004). Paleoseismic studies suggest that the recent dextral strike-slip component of the fault is constant along the central section of the fault at 27 ± 5 mm/yr, but that dip-slip rates are spatially variable and range between 0 and 10 mm/yr (Norris and Cooper 2001). Little et al. (2005) suggest that enhanced uplift occurs between Franz Josef and Fox Glacier due to a localised steepening by $15\text{--}20^\circ$ of the Alpine Fault's dip. The change in the geometry of the (Australian) footwall at depth is inferred to be the cause of upramping of a backshear array in the hanging wall in an escalator-like fashion in this region (Wightman and Little 2007). The southern end of this transition lies in the vicinity of the Karangarua River (Fig. 1.3). This may cause a localised zone of enhanced deformation associated with elevated seismicity levels along a band of seismicity south of Mt. Cook (Little et al. 2005).

The shape of the Southern Alps orogen is highly asymmetric. It has been described as a two-sided deforming wedge, with the Australian plate behaving as a rigid indenter (Koons 1990). This causes two zones of localised shear deformation, one in the inboard side of the orogen and the other in the outboard side (Koons 1990; Beaumont et al. 1996). The highest elevation within the orogen is reached at the Main Divide. Its position and height result from differences in fluvial erosion between the inboard (eastern) and outboard (western) side of the orogen, driven by the asymmetry in orographic precipitation, and the tectonic advection associated with the convergence (Herman and Braun 2006). Uplift is

highest in the zone between the Alpine Fault and the Main Divide Fault Zone, a >60 km-long backthrust system immediately below and to the east of the Main Divide (Cox et al. 1997; Cox and Findlay 1995) (Fig. 1.4) that separates semi-schists from unmetamorphosed greywacke. The Main Divide Fault Zone exhibits a similar small-scale segmentation to the Alpine Fault, and its overall strike is parallel to the Alpine Fault (Cox and Findlay 1995). The segmentation of the Main Divide Fault Zone consists of oblique-reverse faults of 5–15 km length which dip 40–60° NW and are linked by shorter, 3–5 km-long, more steeply-dipping strike-slip faults forming dextral steps (Cox and Findlay 1995). The Main Divide Fault Zone lies 15–30 km southeast of the Alpine Fault and assuming an Alpine Fault dip of 50° SE likely merges with the Alpine Fault at approximately 10–12.5 km depth (Fig.1.4). All inboard structures are progressively advected into the orogen during uplift along the Alpine Fault ramp and exposed due to the erosion (Cox and Findlay 1995). The Main Divide, as the crest of the Southern Alps, controls the drainage patterns. The landscape features steep hillslopes on the western side of the Main Divide, covered by dense rain forest below 1000 m and a small stripe of coastal plain. East of the Main Divide, the topography is less steep, controlled by differential uplift along large southwest-northeast trending, west-dipping reverse faults. The orogen flattens out progressively towards the east coast (Cox and Sutherland 2007) and extends about 100 km eastwards.

A 20 km-long segment of the central Southern Alps has reached an exhumational steady state according to thermochronology studies (Little et al. 2005). Rock is exhumed more rapidly than it can cool, so high temperatures prevail in originally lower-crustal rocks at shallow depths. According to Craw (1997), who analysed adularia veins at high altitudes (>1400 m above sea level), fluid boiling occurs at 500 ± 150 m below the surface. These adularia-bearing veins formed at temperatures of ca. 320–350°C and ca. 1.5 kbar pressure corresponding to ca. 6 km depth assuming lithostatic conditions. Koons (1987) derived one of the earliest models of the thermal structure of the Southern Alps by modelling the Alpine Fault as a vertical discontinuity, assuming uplift rates of 10 mm/a east of the fault and 4 Myr for exhumation of the Southern Alps. His results indicated a 5 km-wide zone of maximum uplift immediately adjacent to the discontinuity with elevated geotherms. Neither heat generation due to radioactive decay nor shearing was considered, and heat transfer was assumed to occur by conduction only. Modelling results indicated temperatures of 350°C at 5 km depth after 2.5 Myr in a narrow zone of high uplift. After this time, the vertical thermal structure remained constant but thermal diffusion caused the high-temperature region to widen horizontally. Later models by Shi et al. (1996), Batt and Braun (1997) and most recently by Toy et al. (2010) all indicate high temperatures at shallow depths in the vicinity of the Alpine Fault, especially in a narrow zone of 10 km

Figure 1.4. A) Map of the central section of the Alpine Fault and Main Divide Fault Zone (separating semi-schist from greywacke) by Cox and Findlay (1995; their Fig. 1), illustrating the segmentation of both faults and their average orientation. Thin lines indicate mapped faults, dashed lines their inferred traces. The thick black line shows the position of the Main Divide, the crest of the mountain range. B) Inferred intersection of the Alpine Fault and the Main Divide Fault at 10–12 km depth along the cross section between Fox Glacier and Mt. Cook as shown in A). C) Equal angle, lower hemisphere stereonet showing the overall orientation of the segments of the Main Divide Fault Zone (two thick lines) in relation to the Alpine Fault (thin line) and the orientation of slickensides after Norris et al. (1990).

(Shi et al. 1996) to 25 km in width (Toy et al. 2010) immediately east of the fault. These studies emphasise the importance of high erosion rates as a constraint on the thermal models.

Fluid flow at shallow depths in the central Southern Alps is inferred to be driven by topography and dominated by surface waters of meteoric origin that penetrate to depths of 5–6 km (Upton et al. 1995). Evidence for this comes from fluid inclusion studies on calcite and quartz veins (Jenkin et al. 1994; Craw 1997). Jenkin et al. (1994) sampled fluid inclusions from steeply-dipping fissures in the vicinity of the Alpine Fault between the Waiho and Fox River that were deposited by circulating fluids of meteoric origin at depths of 1–6 km assuming hydrostatic fluid pressures. Fluid inclusions originate from fluids composed of water, 15–20% carbon-dioxide and up to 5 wt% NaCl equivalent. Templeton et al. (1998) studied fluid inclusions in the Mackenzie Basin in the eastern, outboard region of the Southern Alps which has lower uplift rates and a lower geothermal gradient. They found evidence for mixing of meteoric fluids with metamorphic hydrothermal fluids at temperatures of 300–350°C at depths greater than 5 km. Near the surface, these fluids mix with basinal waters, characterised by a systematic shift from low $\delta^{18}\text{O}_{\text{calcite}}$ and high $\delta^{13}\text{C}$ to higher $\delta^{18}\text{O}_{\text{calcite}}$ and lower $\delta^{13}\text{C}$ values in post-metamorphic veins. Several hot springs within 10 km of the Alpine Fault north of Franz Josef have oxygen and hydrogen isotope compositions typical of meteoric water and rock-equilibration temperatures of 90–150°C, suggesting circulation to depths shallower than 3 km (Barnes et al. 1978; Jenkin et al. 1994; Allis and Shi 1995; Templeton et al. 1998).

Fluids at mid-crustal depths have been inferred from magnetotelluric studies by Wannamaker et al. (2002) as the cause of high conductivity. Modelling by Vry et al. (2009) showed that rapid exhumation and near-isothermal decompression of typical Alpine schist causes dehydration of the rock, due to a phase transition from epidote to plagioclase. This phase transition occurs in a narrow temperature range ($400^\circ\text{C} < T < 560^\circ\text{C}$) during decompression and generates up to 0.009 moles of fluid per mole of rock. This is possible if the rock has not experienced the same phase transition at higher temperatures and pressures before. Such temperatures are associated with the biotite zone/garnet zone transition that lies approximately midway between the Alpine Fault and the Main Divide Fault Zone (Vry et al. 2009). According to Vry et al. (2009), low volumes of fluid (<1%) interconnected through permeable rock and salinities of 2.5–5 wt% are sufficient to cause the high conductivities observed for the middle to lower crust of the Southern Alps during magnetotelluric sounding by Wannamaker et al. (2002). Wannamaker et al. (2002) inferred that large volumes of fluids, released from metamorphic processes, ascend in a U-shaped conductivity anomaly towards the brittle–ductile transition zone. The centre point of this anomaly lies at 20 km

depth and terminates at 10 km depth with subvertical legs 5–10 km and 60 km southeast of the Alpine Fault. The depth of the top of the conductor is well-resolved (Jiracek et al. 2007) and the conductivity significantly exceeds that of the overlying crust. Wannamaker et al. (2002) proposed that the fluids are trapped in the mylonitised, ductile shear zone, inferred from other studies to lie at these depths. When the uplifted, ductile rocks cross the brittle–ductile transition zone, these fluids are released and ascend, causing a change from lithostatic to hydrostatic fluid pressures in the fault zone.

The high-conductivity region coincides with a zone of low seismic P-wave velocities (reduced by 10%; Smith et al. 1995; Stern et al. 2001). Velocity profiles derived from active source wide-angle reflection and refraction data collected during the SIGHT (South Island Geophysical Transect) project in 1996 and 1998 (Davey et al. 1998; Okaya et al. 2002) show the extent of the low-velocity zone in the Pacific Plate between depths of 6 km (transect 1) and 14 km (transect 2) to the base of the crust at maximum depths of 37–44 km (Scherwath et al. 2003; Van Avendonk et al. 2004). A second low-velocity zone in the Australian plate at 5 to 10 km depth is interpreted to be the result of flexural bending (Fig. 1.5).

A broad, negative, Bouguer gravity anomaly (represented by the –80 mgal contour) extends over a 50×200 km² area (Woodward 1979) about 70 km south of Mt. Cook. The Bouguer gravity anomaly indicates an extensive crustal root of lower density material. The crustal root is inferred to result from lithospheric thickening caused by the plate convergence, and is formed by thickened middle and lower crust (Scherwath et al. 2003). The crust is 15–20 km thick east and west offshore South Island (Scherwath et al. 2003; Van Avendonk et al. 2004) and 27 km at the west coast (Melhuish et al. 2005). Shortening of the crust has increased crustal thicknesses to 40–50 km beneath and slightly east of the Southern Alps (e.g. Davey et al. 2007). The thickness of the crustal root also increases southwards. Gerbault et al. (2002) suggested that southward extrusion of ductile lower crustal material has occurred since significantly more crustal thickening has been inferred in the northern part than in the southern part of the Southern Alps. The area of lowest isostatic gravity anomaly (–30 mgal) extends just southeast of the maximum topography of Southern Alps. Since the crustal root is much larger than needed to balance the Southern Alps isostatically, this suggests a mass excess in the upper mantle (Bourguignon 2009, and references therein). The lithospheric structure is asymmetric in the crust (concentrated in the Pacific plate) but symmetric within the mantle (Stern et al. 2000, Figure 1.5). This has been inferred from P-wave travel-time residuals of teleseismic earthquakes which decrease across the South Island and can be explained by a subvertical high-velocity mantle anomaly

(Stern et al. 2000). Bourguignon (2009) modelled its minimum extent as 110 ± 20 km in width and 70 ± 20 km in thickness assuming a density contrast of 300 kg/m^3 between lower crust and mantle.

More than 15 regional global positioning system (GPS) surveys were conducted in the central and southern South Island between 1994 and 2004 (Wallace et al. 2007). These campaign surveys showed that velocities parallel to the strike of the Alpine Fault are fairly uniform. Two temporary GPS transects, one extending from Lake Hawea to Haast (Pearson et al. 2000) and the other from Christchurch to Hokitika (Pearson et al. 1995, Fig. 1.3), showed that the highest strain rates occur in the vicinity and to the east of the Alpine Fault. Vertical velocities across the Hawea–Haast transect are zero within the 95% confidence limits (Pearson et al. 2000). Beavan et al. (1999) state that errors in vertical velocities obtained in the campaign surveys are too large (>10 mm/yr) to give reliable estimates. A permanent continuous GPS array crossing the Southern Alps between Jacobs River and Waitaha River, the Central South Island (CSI) network, has been recording since 2000 (Beavan et al. 2004; Beavan et al. 2010a). Vertical GPS rates are calculated with respect to a local reference frame. The station OUSD near Dunedin is assumed to be far enough away from the plate boundary that it can be treated as a stable point on the non-deforming Pacific plate with a fixed vertical velocity of 0.0 ± 0.4 mm/yr (e.g. Beavan et al. 2004). Results from analysis of the CSI data yield maximum rates of vertical movement of 6–9 mm/yr relative to the east coast which occur about 20–30 km southeast of the Alpine Fault (Beavan et al. 2010a). Vertical rates have higher uncertainties than horizontal velocities (by a factor of three in the Southern Alps according to Beavan et al. 2007), due to smaller uplift rates, systematic errors (e.g. satellite geometry, atmospheric and ionospheric effects) and are more influenced by various processes such as seasonal cycles, localised erosion and isostatic compensation.

Linear elastic two-dimensional dislocation modelling is commonly used to model GPS data. Beavan et al. (1999) and Pearson et al. (2000) found that single-fault models of the Alpine Fault require much greater locking depths (22 ± 1 km for the central South Island data and 20 ± 2 km for the Hawea–Haast transect) than suggested by the seismicity. In both cases, strike- and dip-slip rates (of 36 ± 0.8 mm/yr and 15.5 ± 1.1 mm/yr or 34 ± 2 mm/yr and 25 ± 9 mm/yr, respectively) differ from geological rates. If a two-fault model is used with a second fault dipping in the opposite direction approximately 80 km southeast of the Alpine Fault, this results in shallower locking depths and/or lower coupling on the Alpine Fault. For the Hawea–Haast data, the best-fitting model has a locking depth of 10 ± 2 km and strike- and dip-slip rates of 23 ± 2 mm/yr and 11 ± 8 mm/yr, respectively (Pearson et al. 2000).

Figure 1.5. Deformation in crust and mantle beneath the central Southern Alps after Davey et al. (2007; their Fig. 14)

Beavan et al. (1999; 2004) found two-fault dislocation models that matched horizontal surface deformation by modelling a second opposite-dipping fault below 30 km depth. This model fits high uplift rates within the mountains but cannot fit observations to the east of the Southern Alps (Beavan et al. 2004). Two-fault models assume a fixed dip for the second structure and are insensitive to whether a single second fault or numerous faults in a broad shear zone accommodate motion (Pearson et al. 2000). Moore et al. (2002) showed that the observed horizontal velocity field can be modelled equally well with a model in which deformation in the upper crust is localised on faults but distributed over a shear zone 30–40 km wide in the lower lithosphere.

Simultaneous inversions for angular velocities of rotating 3-D elastic blocks and the degree of fault coupling has been conducted by Wallace et al. (2007). For the central section of the Alpine Fault between latitudes -43.0° and -43.5° , Wallace et al. (2007) obtained slightly weaker coupling at 1–18 km depths ($\Phi=0.7$ where the coupling coefficient is defined by Wallace et al. (2007) as $\Phi=1-V_c/V$ for creep-rates V_c and long-term slip-rates V) than in surrounding strongly coupled segments ($\Phi=0.85$), indicating a shallower coupling depth in this section.

Geological, geophysical and geodetic studies suggest that only a portion (2/3) of the relative plate motion is accommodated on the central Alpine Fault. The remaining third must be accommodated by structures within and to the east (or west) of the Alpine Fault (Norris and Cooper 2001). Walcott (1978) suggested that distributed deformation occurs on the numerous smaller faults on the eastern side of the Southern Alps. This is difficult to prove, since juvenile faults are abundant in the Southern Alps region and it is not known which of these faults are active. Within 60 km of the Alpine Fault 110 mapped faults of 4–73 km length have been mapped (Cox et al. 2012). The four largest bound the glacial valleys east of the Main Divide and strike north-northeast at angles $85\text{--}120^\circ$ to the maximum compressive stress direction (Cox et al. 2012). The Black Bob Fault and the Haast Ridge Fault form parts of the Main Divide Fault Zone. The Murchison Fault and Liebig Fault bound adjacent high-elevation ranges. The direction of fault slip on these faults obtained from slickenside is predominantly dip-slip, however, minor components of both dextral and sinistral motions are present (Cox et al. 2012). Known active faults in the southeastern foothills of the central Southern Alps (≥ 60 km from the Alpine Fault) include the Ostler Fault and the Lake Heron/Forest Creek/Irishman's Creek fault system (e.g. Barrell and Strong 2010). The Ostler Fault Zone is a ~ 50 km-long, north–south striking, low-angle reverse fault. Its surface trace borders the western margin of the Mackenzie Basin (e.g. Blick et al. 1989). The fault dips $50\text{--}60^\circ\text{W}$ at shallow depths and Ghisetti et al. (2007) inferred it to be a reactivated

normal fault based on its dip (Sibson and Xie 1998). There is little evidence constraining slip-rates on the Ostler fault, but Ghisetti et al. (2007) estimated them to be low (<1.5 mm/a).

Recent large earthquakes in the vicinity of Christchurch (e.g. Gledhill et al. 2011; Quigley et al. 2012) may indicate that the zone of deformation in the central South Island extends further east than previously assumed. Several eastwest-trending faults have been identified in the Canterbury plains by active seismic reflection/refraction studies (Dorn et al. 2010). The only large tectonic structure mapped northwest of the Alpine Fault is the South Westland Fault Zone or Coastal Monocline (Nathan 1986; Cox and Barrell 2007). It consists of a number of continuous southeast-dipping reverse faults and runs offshore between Franz Josef and Fox Glacier.

1.3. Seismicity in the central Southern Alps

Large earthquakes in the South Island have occurred in historic times in the Puysegur subduction zone to the south of the central Southern Alps region, in the Buller region (Doser et al. 1999; Anderson et al. 1993; Dowrick 1991) to the north, near Arthur's Pass (Doser et al. 1999; Berryman and Villamor 2004), in the Marlborough Fault System (Cowan 1991) in the northeast and most recently near Christchurch in the east (Gledhill et al. 2011; Quigley et al. 2012). Strike-slip displacements were observed for the September 1888 Glynn Wye earthquake ($M \geq 7$) on the Hope Fault with 3 m offset on fences crossing the fault (Cowan 1991), and for the March 1929 Arthur's Pass earthquake on the Kakapo/Poulter fault (Doser et al. 1999; Berryman and Villamor 2004). The Darfield earthquake on 4 September 2010 showed maximum dextral displacements of 4.6 m on the previously unknown Greendale fault (Gledhill et al. 2011; Quigley et al. 2012).

Since the beginning of European settlement in Westland in 1840 (Wells et al. 1999), no rupture on the Alpine Fault has been observed (Evison 1971; Anderson and Webb 1994). Only two earthquakes of $M_W \geq 6$ have been reported in the central Southern Alps region: the 1946 M_W 6.7 Lake Coleridge (Doser et al. 1999) and the 1984 M_W 6.1 Godley Valley earthquake (Anderson et al. 1993). Both strike-slip events occurred southeast of the Alpine Fault, at distances of 40 and 55 km from the surface trace (Fig. 1.6 and Table 1.1). These earthquakes have similar strike directions and steeply-dipping fault planes. Anderson et al. (1993) favoured the northeast-striking fault plane undergoing right-lateral strike-slip as the rupture plane of the Godley Valley earthquake, a fault geometry similar to that of the Porter's Pass–Amberley Fault Zone to the northeast (Fig. 1.6). The Godley Valley earthquake was

Figure 1.6. Major faults and focal mechanism solutions for the six largest, more recent events in and adjacent to the study area (dashed rectangle) since 1946 (Table 1.1) in the central South Island of New Zealand. All focal mechanism solutions are lower-hemisphere projections of the focal sphere. Displayed faults are from the Active Faults Database by GNS Science. The inset map shows the tectonic setting of the Alpine Fault.

followed by a $M \sim 5$ aftershock in 1987, 10 km to the east of the epicentre (Little et al. 2005), and the region still exhibits elevated levels of microseismicity (Leitner et al. 2001).

The Marlborough faults have produced three large earthquakes ($M > 7$) since 1840 (Grapes et al. 1998; Cowan 1991; Doser et al. 1999; Berryman and Villamor 2004). Notably seismically active is the region where the Hope Fault merges with the Alpine Fault. Seven earthquakes of $M_W > 5.4$ have occurred in this zone of high deformation since 1888 (e.g. Rynn and Scholz 1978; Árnadóttir et al. 1995)(Table 1.1). These events exhibited a mixture of strike-slip and reverse focal mechanisms (Doser et al. 1999). The two most recent earthquakes in this region were shallow reverse-faulting events (Abercrombie et al. 2000), the M_W 6.7 Arthur's Pass earthquake in 1994 (Robinson and McGinty 2000) followed by the 1995 M_W 6.2 Cass earthquake 30 km to the east (Gledhill et al. 2000). The largest after-

Earthquake name	Date	M _W	Lat [°]	Lon [°]	Depth [km]	Strike [°]	Dip [°]	Rake [°]	Dist [km]	Ref
Glynn Wye	1 September 1888	7.3	-42.63	172.41	15±5	75±20	90±20	-180±20	-	Hi
Arthur's Pass	9 March 1929	7.0	-42.79	171.93	11±5	65±7	90±20	-178±10	-	Do
Lake Coleridge	26 June 1946	6.5	-43.46	171.26	9±4	44±17	89±8	180±13	55.5*	Do
Jackson Bay	13 October 1947	6.2	-44.42	168.48	18±2	26±25	46±18	43±32	22.6	Do
Godley Valley	24 June 1984	6.1	-43.60	170.56	13 ⁺² ₋₃	62 ⁺⁵ ₋₁₂	83 ⁺¹⁰ ₋₅	-154±5	40.2*	An
Wilberforce River	30 March 1992	5.5	-43.04	171.23	5	70	19	134	17.6*	Do
Arthur's Pass	18 June 1994	6.7	-42.98	171.48	5±1	221 ⁺⁵ ₋₁₀	47±5	112 ⁺⁵ ₋₁₀	24.4*	Ab
Aftershock AP	19 June 1994	5.9	-43.16	171.52	8 ⁺¹ ₋₄	253±15	87 ⁺⁵ ₋₁₀	178 ⁺¹⁰ ₋₂₀	42.7	Ab
Unnamed event	29 May 1995	5.5	-42.96	171.50	15	53	67	174	24.0	Do
Cass	24 November 1995	6.2	-42.96	171.83	9	176±10	46±5	44±10	37.0	Gl
Unnamed event	20 October 1998	5.4	-43.83	169.55	12	78	54	-175	14.3*	Dz
Jackson Bay	7 December 2001	5.8	-44.17	168.71	7	48	45	103	10.1*	McG

Table 1.1. Earthquakes of M_W>5.4 and their focal mechanism solutions in the central South Island. Uncertainties are provided when known. For some recent events, the distance (Dist) to the surface trace of the Alpine Fault was determined by projecting the earthquakes on a plane perpendicular to the fault (striking at an azimuth of 145°, the asterisk marks those events plotted in Fig. 1.6). Abbreviated references (Ref) for listed focal mechanism solutions refer to Abercrombie et al. (2000); Anderson et al. (1993); Doser et al. (1999); Dziewonski et al. (1999); Gledhill et al. (2000); Hincapié et al. (2005); McGinty et al. (2005) and references therein.

shocks of the Arthur's Pass earthquake exhibited strike-slip mechanisms (Abercrombie et al. 2000; Doser et al. 1999).

At the southern limit of the central Alpine Fault, two large events occurred at distances of 14 and 3–10 km, respectively, southeast of the surface trace of the Alpine Fault in 1998 and 2001 (Fig. 1.6 and Table 1.1). The 1998 M_W 5.4 earthquake has not yet been studied in detail (but was referred to by Eberhart-Phillips and Bannister 2002). The CMT focal mechanism solution (Dziewonski et al. 1999) gives a strike-slip mechanism but the solution is poorly constrained. The 2001 M_W 5.8 Jackson Bay earthquake (McGinty et al. 2005) was a shallow thrust earthquake with no surface rupture. The aftershock distribution exhibited clustered events 5 km to the southeast of the Alpine Fault. McGinty et al. (2005) obtained a variety of focal mechanism solutions for the aftershocks with predominantly strike-slip faulting. Further south of Jackson Bay, a M_W 6.2 event in 1947 exhibited an oblique-reverse mechanism (Doser et al. 1999) (Table 1.1).

Paleoseismic studies by Adams (1980), Wells et al. (1999) and Wells and Goff (2007) of aggradation terraces, coastal dunes, landslides, trenches across the active fault trace, forest (re-)establishments and tree-ring growth indicate that the two most recent, major events occurred in 1717 and 1630 C.E. with rupture lengths of several hundred kilometres and individual offsets of 8–12 m (Walcott 1998; Sutherland and Norris 1995; Sutherland et al. 2007; Berryman et al. 2012). These studies suggest that failure of the Alpine Fault resulted in large surface displacements along broad sections of the fault (Norris and Cooper 2001).

Seismicity of the South Island was first discussed by Eiby (1971) and Evison (1971). Eiby (1971) distinguished seismicity in the Central Seismic Region from the Main Seismic Region to the northeast and the Fiordland Seismic Region to the southwest, due to pronounced differences in the number and distribution of earthquakes. In 1972, Scholz et al. (1973) conducted a microseismicity study along several parts of the Alpine Fault including a section between Hari Hari and Haast with a network consisting of 10 seismometers. They recognised distinctly lower seismicity rates in this region than to the northeast and southwest. Events occurred at 2–20 km depth and appeared to be spatially continuous to the east of the fault. In a subsequent microseismicity study of the Haast Region in 1973, Caldwell and Frohlich (1975) located 60 of the 122 events recorded over 32 days in a depth range of 4–14 km. Those stations within 5 km of the fault recorded many unlocatable events with S–P travel times smaller than 2.5 s. More than 90% of the locatable earthquakes occurred in a 20 km-wide area about 10 km to the southeast of the fault. Events occurred in groups but did not show a pattern or a defined fault plane and did they have a common focal

mechanism. Caldwell and Frohlich (1975) did not determine magnitudes but estimated them to range between -1 and 3 .

Between 1975 and 1983, when the level of Lake Pukaki was raised by 37 m for hydroelectric power generation, seismicity was monitored by an array of nine short-period seismometers around the lake. Reyners (1988) located 2825 earthquakes of $M_L \geq 0.8$. Of these events only 48 occurred at depths greater than 15 km and 65% of all earthquakes were located between 5 and 9 km. Reyners (1988) reported a completeness magnitude of $M_L = 1.8$. A background seismicity rate of approximately 6.2 events per 30 days was observed. Reyners (1988) obtained a b-value (Section 2.8) of approximately 1 for the background seismicity and observed temporal changes in the b-value in the lake vicinity during the survey period. An increase in seismicity associated with the impounding of the lake was observed and the largest triggered event of M_L 4.6 occurred on 17 December 1978 at shallow depth (< 5 km).

Eberhart-Phillips (1995) investigated seismicity occurring between March 1991 and April 1994 recorded by the New Zealand National Seismograph Network (NZNSN) in the central South Island. The network was relatively sparse at that time, with an average distance between each earthquake and the nearest station of 40 km and only 2 stations (WVZ and EWZ) in the central Southern Alps. Eberhart-Phillips (1995) reported mainly diffuse seismicity southeast of the Alpine Fault (complete to a magnitude of 2.6) with aftershocks in the area of the Godley Valley earthquake and one swarm further southwest. Although uncertainties in depth were large, an increase in the average depth from 10 km adjacent to the fault to 20 km in 60–70 km distance from the Alpine Fault could be observed.

The Southern Alps Seismic Passive Experiment (SAPSE) comprised 26 broadband and 14 short-period stations deployed across the central South Island for six months (Anderson et al. 1997) in 1995/1996. Average station distances were 30 to 50 km. Leitner et al. (2001) used the data from SAPSE, high-quality location data of the New Zealand National Station Network (NZNSN) and the Lake Pukaki network, and data recorded during a temporary Mount Cook aftershock deployment to conduct a comprehensive study of seismicity along the Alpine Fault. They report a triangular section between the Wanganui River and Franz Josef Glacier where the seismicity of $M_L \leq 2$ is almost absent (Fig. 1.7). According to Leitner et al. (2001), the eastern boundary of the low-seismicity zone coincides with the south-western extension of the Porter's Pass Fault Zone. That study reported 5491 earthquake locations with magnitudes of $2 \leq M_L \leq 4.2$. A simultaneous inversion to determine hypocentre locations and P-wave velocity structure was

Figure 1.7. Seismicity recorded in the central Southern Alps since 1990 by GeoNet (grey, magenta large earthquakes since 1990), Reyners (1988) (light blue), Leitner et al. (2001) (light red) and (O' Keefe 2008) (light green). The stations of the networks which recorded these events are shown in equivalent colours

Depth [km]	Velocity [km/s]		
	Leitner et al. (2001)	O' Keefe (2008)	this study
–2		3.50	5.67
0	6.0	5.67	
8		5.79	5.79
18	6.3	6.28	6.28
35	7.2	7.35	7.35
40	8.0	8.00	8.00

Table 1.2. 1-D velocity models for the central Southern Alps in comparison. Note that not all models have the same depth boundaries, so the velocity column is blank if there is no such depth boundary in the model.

carried out using a subset of 195 earthquakes (Table 1.2). With this amalgamated dataset, 53 well-constrained focal mechanisms throughout a large part of the South Island were computed. Well-constrained focal mechanisms beneath the central Southern Alps exhibited oblique strike-slip and thrust mechanisms, as well as two normal-faulting mechanisms which were inferred by the authors to have been hydrologically triggered. The latter were derived from a swarm of events following severe rainfall and flooding (102 mm of rain in 32 hr) in mid-December 1995 (Leitner et al. 2001). The observed maximum seismogenic depth was rather uniform with 12 ± 2 km except under the highest elevations of the Alps where it was 3–4 km shallower.

In a recent study by O' Keefe (2008) using eight surface seismographs between Fox Glacier and Hari Hari between September 2006 and March 2007, 411 earthquakes with magnitudes $M_L \geq 1.6$ were located. O' Keefe (2008) identified five earthquake swarms, one near Fox Glacier, one near station Whymper Hut, one in the upper Clyde River, one near Whataroa and one near Waitaha. During the survey, seismicity revealed a “horseshoe” pattern with the open side pointing to the fault. Only a few earthquakes occurred close to the fault trace, so no information could be gained about the Alpine Fault's structure. O' Keefe (2008) obtained a b-value of 0.75 ± 0.06 , much lower than in previous studies, and a maximum seismogenic depth of 15 km for most of the survey region but only 10 km for the Fox region. Following the simultaneous inversion for velocity structure (Table 1.2) and hypocentre locations, an average depth error less than 3 km was obtained.

Sporadic subcrustal earthquakes were reported by Eiby (1971), Calhaem et al. (1977), Haines (1979), Reyners (1987) and in several recent studies. Reyners (2005) inferred a subcrustal depth of ~ 50 km for the 8 May 1943, M_W 5.9 Lake Hawea earthquake. Kohler and Eberhart-Phillips (2003) analysed in detail intermediate-depth earthquakes in

the central South Island with magnitudes $M_L \leq 4.0$. Events with well-constrained depths exceeding 30 km recorded by the NZNSN between 1990 and 2000 and by the Lake Pukaki network between 1975 and 1983 were investigated. These results show that deep crustal or mantle events are more frequent at the southern end of the Southern Alps than at the northern end. These events correspond to regions of high P-velocity anomalies or their boundaries. The shallowest of these intermediate-depth events lie at >30 km depth near or below the crustal root. Reyners (1987) suggested, that subcrustal events occur on a 19° NW-dipping plane associated with intra-continental subduction. Kohler and Eberhart-Phillips (2003) concluded that these events are controlled by high shear-strain gradients associated with depressed geotherms and viscous deformation of mantle lithosphere.

1.4. The Southern Alps Microearthquake Borehole Array (SAMBA)

1.4.1. Study area and station networks

In this study, we investigate the area bounded by the three permanent broadband stations WVZ, RPZ and FOZ, operated by GeoNet in the central Southern Alps (Table 1.3 and Fig. 1.8). These stations are spaced 80–100 km apart and have been recording continuously since 1990 following a nation-wide update to digital recording (e.g. Anderson and Webb 1994; Petersen et al. 2011). To supplement the New Zealand national station network in the central section of the Southern Alps, the Southern Alps Microearthquake Borehole Array (SAMBA) comprising eleven short-period seismometers was installed between November 2008 and June 2009. Seven of those instruments are deployed in shallow boreholes of 1–2 m depth, three in deeper boreholes of 46, 61 and 98 m depths, and one surface sensor was added in January 2010 to extend the array to the south (Table 1.4). Along the strike of the Alpine Fault, these stations are approximately evenly spaced 8 km apart on alternating sides of the fault. Perpendicular to the fault, the average distance between SAMBA stations is 16 km. This geometry allows for accurate depth estimates for events located within the network (where the distance between the earthquake hypocentres and the closest station does not exceed 1.5 times the focal depth, see Section 3.4.1 and 4.2.1).

Before March 2009, when the three deep boreholes were drilled, temporary sites near the drilling locations were established (Table 1.4). A “2” appended to the station name is used to indicate a subsequently installed borehole station. A drill rig mounted on a truck was

Figure 1.8. Distribution of the GeoNet broadband stations (black reversed triangles), short-period SAMBA (red triangles) and DFDP10 stations (blue triangles) in the study area. The inset shows the location of the study area in the central Southern Alps.

used to drill the three deep boreholes. A combination of rotary drilling and hammering was applied using compressed air for cooling and to remove the rock chips. For the shallow holes, a portable drill rig was mounted on the rock and drilling was carried out by water-cooled rotary drilling. This enabled the whole rock core to be extracted (albeit in pieces). All boreholes of the SAMBA project have a diameter of approximately 11 cm (4.5 inch), are uncased, and in solid rock except for one which is located in a sediment-filled valley (POCR). The latter has a diameter of approximately 15 cm (6 inch) and is completely steel-cased. In order to suppress air motion near the sensor, the boreholes were filled with sand. A study by Holcomb et al. (1997) showed that this particularly improves the noise level on the horizontal components of the seismogram.

The SAMBA stations are equipped with GeoSpace Technologies HS-1-LT short-period three-component seismometers. Most of the sensors are in small tubes of 26.5 cm length

Name	Location	Lat. [°]	Lon. [°]	Alt. [m]	Ground	Sen- sor	Date installed
WVZ	Waitaha Valley	-43.076	170.736	75	Granite	3ESP	06.09.2003
FOZ	Fox Glacier	-43.566	169.689	10	Greywacke	3ESP	13.10.2004
RPZ	Rata Peaks	-43.719	171.054	412	Rhyolite	3TB	06.06.2001

Table 1.3. Station locations of the three permanent broadband stations operated by GeoNet in the study area. Table modified from Petersen et al. (2011)

and 7.3 cm diameter, have no gimbals and were orientated manually before being lowered into the borehole. The three seismometers in the deeper boreholes have a length of 1 m and a diameter of approximately 10 cm, and contain gimbals. A Mark Products L-4C3D short-period seismometer was temporarily placed on top of each borehole sensor, so that the orientation of the borehole sensor could be determined by comparison of the waveforms on the horizontal components. The same L-4C3D sensor was later used for the additional surface site COVA deployed in the Copland Valley. Each station consists of a Reftek data logger RT130-01, a Reftek GPS clock, a three-component short-period seismometer and a power supply.

Name	Location	Lat. [°]	Lon. [°]	Alt. [m]	Depth [m]	Ground
COSA	Cook Saddle	-43.448	170.060	377	1.40	(+5 m of gravel)
EORO	End of Road	-43.426	170.169	233	1.35	schist
MTFO	Mount Fox	-43.512	170.003	1216	2.00	schist (bed)
WHAT2	Whataroa Quarry	-43.279	170.360	95	61.00	granite (bed)
WHYM	Whymper Hut	-43.441	170.371	906	1.30	landslide
POCR2	Potter's Creek	-43.352	170.223	185	46.00	sediments
LABE	De la Beche	-43.546	170.245	1590	2.15	greywacke (bed)
GOVA	Godley Valley	-43.639	170.503	814	1.95	greywacke (bed)
FRAN	Carnavan's Knob	-43.380	170.160	124	98.00	granite (bed)
REYN	Reynold's Creek	-43.353	170.391	1310	1.20	semi-schist (bed)
COVA	Copland Valley	-43.613	169.968	1477	~1	schist (bed)
Name	Time	Lat. [°]	Lon. [°]	Alt. [m]	Depth [m]	Ground
WHAT	11/2008–03/2009	-43.279	170.361	106	2.40	granite (bed)
POCR	12/2008–03/2009	-43.348	170.177	95	0.40	sediments
MARI	12/2008–02/2009	-43.727	170.575	771	1.00	soil

Table 1.4. Final station locations (top) and temporary station locations before March 2009 (bottom) of the SAMBA network. The depth gives the total borehole depth. Note that the sensor at COSA was covered by at least 5 m of additional gravel from a landslide nearby. Stations that are thought to be located in bedrock (bed) are marked.

A second temporary seismometer array deployed as part of the Deep Fault Drilling Project (DFDP10) operated temporarily northeast of SAMBA (Fig. 1.8). This array comprised twelve short-period Mark Products L-22D3D seismometers recording from 20 January to 1 May 2010. Most of these instruments were either buried approximately 0.4 m deep in soil or sediments or directly placed on bedrock and covered with sand and stones (Table 1.5). The DFDP10 stations were initially operated on normal gain ($\times 1$) but this was changed to high gain ($\times 32$) at the first station service (Julian day 074/075 2010). All stations consisted of a Reftek data logger RT130-01, a Reftek GPS clock, a three-component short-period seismometer and a power supply.

Name	Location	Latitude [°]	Longitude [°]	Altitude [m]	Ground
BLO	Blue Lookout	−43.244	170.787	1607	bedrock
BON	Mt Bonar	−43.083	170.651	1031	soil
DRC	Dry Creek	−43.241	170.452	128	soil
ERE	Erewhon Station	−43.510	170.855	600	soil
GCK	Gaunt Creek	−43.306	170.311	114	soil
GHU	Godley Hut NZAC	−43.480	170.509	1108	moraine
NOL	Nolans Hut	−43.349	170.484	181	soil
ONE	One One Road	−43.114	170.492	62	soil
POE	Poerua Valley	−43.195	170.520	103	soil
VBV	Veil Bivouac	−43.385	170.680	1019	moraine
WHB	Whataroa Bridge	−43.295	170.412	97	sediments
WNQ	Wanganui Quarry	−43.182	170.631	180	sediments

Table 1.5. Station locations of the DFDP10 array between January and May 2010

All stations operate with continuous GPS recording. The response functions of all the different sensors used in this study are given in Appendix A. The polarities of the SAMBA and DFDP10 stations have been checked as described in Appendix B. No polarity reversals were found for the SAMBA borehole stations. However, the polarity of the Mark Products sensors used for station COVA and all stations of the DFDP10 network were found to be reversed for the vertical component. Also, the instrument response of GeoNet station RPZ was identified as wrong in the sense that it reverses the correct polarity (Appendix A).

1.4.2. Data acquisition, processing and quality control

All the short-period sensors used in the SAMBA and DFDP10 arrays have a natural frequency of 2 Hz, except for the sensor in the deepest borehole (FRAN) which has a natural

frequency of 4.5 Hz and the surface sensor (COVA; 1 Hz). The SAMBA stations record at a sample rate of 200 samples per second on high gain. The DFD10 stations recorded at 100 samples per second except for one instrument (BLO) which erroneously recorded at 40 samples per second for the second half of the recording period. Data obtained from GeoNet are also recorded at 100 samples per second.

The continuous Reftek raw data were processed by converting to miniseed format using the Reftek tool *rt_mseed* and to sac format using *ms2sac*. Miniseed data from all short-period stations and the data from the three GeoNet broadband stations were combined into single 15 minute-long miniseed files for inspection using the earthquake analysis software SEISAN (Havskov and Ottemöller 2000). The continuous data are scanned for events using a triggering routine that compares short-term and long-term averages in 0.5 s and 10 s data windows, respectively. When the signal-to-noise ratio exceeds a specified threshold of 3.0 at four stations, event windows of 60 s length are extracted for manual inspection. The detection threshold of the triggering routine is chosen to be low in order to detect microearthquakes recorded by only a few stations. The arrival-times of the P- and S-waves of all recorded earthquakes have been identified manually using SEISAN (Havskov and Ottemöller 2000). For wave arrival-times recorded by three or more stations, hypocentre locations are determined with the SEISAN tool HYP, a modified version of HYPOCENTER (Lienert et al. 1986). These preliminary hypocentres are then relocated using the non-linear location program NonLinLoc 5.0 of Lomax et al. (2000) to better constrain the posterior uncertainties in the hypocentre parameters (Section 4.2.1).

Continuous recordings were checked for GPS absolute timing uncertainties by scanning State of Health files from the Reftek data loggers. For short time spans, the program *logpeek* was used to check how long the GPS timing uncertainty lasted. Events which occurred during uncertain timing were assigned a weight of 9, so that relative arrival times $t_s - t_p$ instead of absolute arrival times were used. For time spans longer than several days, the timing uncertainty was determined from analysis of the residuals (Section 3.1.1). These longer periods of GPS timing uncertainties are discussed in Section 3.1.3 and 3.1.4.

The location quality was checked by comparing hypocentre locations of those earthquakes located by GeoNet with locations obtained using SAMBA (Section 3.3.1). The hypocentre locations before and after weighting (Section 3.1.2) were also compared. If the new location shifted more than 5 km upwards (which in almost all cases corresponded to moving the earthquake to the surface), the arrival-times of those events were reviewed and repicked.

1.4.3. Performance of the SAMBA stations

In regions of high background noise, the installation of seismometers in boreholes can significantly improve the quality of the recorded data (Bormann 2002). A borehole reduces environmental influences on the sensor such as temperature and pressure fluctuations. Turcotte and Schubert (2002) derived the following formula for the amplitude of time-dependent temperature fluctuations in the subsurface, assuming that the surface temperature is periodic in time $T_S = T_0 + \Delta T \cos \omega t$:

$$T = T_0 + \Delta T \exp(-z/d) \cos(\omega t - z/d), \quad (1.1)$$

where T_0 is the average surface temperature, ΔT is the variance of the temperature, z describes the depth, $d = \sqrt{2\kappa/\omega}$ the skin depth, κ is the thermal diffusivity, $\omega = 2\pi f$ is the angular frequency and f the frequency of the temperature variation. Equation 1.1 is plotted for periods of 12 hours, half a year and one year in Fig. 1.9. The thermal diffusivities of different rocks at room temperature have been documented in many studies (e.g. Hanley et al. 1978; Whittington et al. 2009) and range between 1.3 and 1.8 mm²/s for schist (Whittington et al. 2009). At approximately 1 m depth, daily temperature fluctuations no longer affect the sensor. The influence of seasonal temperature cycles on the instrument vanish at depths of 10 to 15 m (Fig. 1.9).

The high-frequency content of the seismic waves is better recorded in a borehole because the attenuation which is highest in the near-surface layers is reduced. Abercrombie (1998) showed that the uppermost hundred metres of the crust are most attenuative ($Q \sim 10$; Section 2.6.1), independent of the rock-type.

It is common to analyse acceleration spectra of ambient seismic noise to assess the site quality (e.g. Bormann 2002). The power spectral density

$$P(\omega) = \int_{-\infty}^{\infty} p(\tau) \exp(i\omega\tau) d\tau \quad (1.2)$$

is the Fourier transform of the auto-correlation function of the noise $p(\tau)$ (Bormann 1998), ω is the angular frequency and τ a time interval. The acceleration power spectral density $P_a(\omega)$ is obtained from the velocity power spectral density $P_v(\omega)$ by

$$P_a(\omega) = \omega^2 P_v(\omega) \quad (1.3)$$

$$\rightarrow P_a(f) = 4\pi^2 f^2 P_v(f) \quad (1.4)$$

Figure 1.9. Temperature decay with depth for temperature variations with different periods. Equation 1.1 is plotted for $T_0=0^\circ\text{C}$, $\Delta T=15^\circ\text{C}$ and $\omega t=\text{const}$.

and expressed in units of decibel [dB]:

$$P_a = 10 \log(P_a / 1(\text{m/s}^2)^2 \text{s}) \quad (1.5)$$

Typical seismic noise spectra are dominated by marine microseismic noise in the 0.05–0.5 Hz band (Havskov and Alguacil 2004). This noise reflects the periods of the water waves (10–16 s) and mainly originates from storms and breaking surf along shorelines (Havskov and Alguacil 2004). At higher frequencies, wind and cultural noise dominate. Traffic and machinery usually generate noise at high frequencies (>2–4 Hz) (e.g. Bormann 1998). Seasonal and diurnal variations for microseismic and cultural noise at a given site can differ by one to two orders of magnitude in amplitude (Havskov and Alguacil 2004).

To assess the performance of the SAMBA stations and to quantify the improvement obtained by installing sensors in shallow boreholes, acceleration power spectral densities are calculated in the frequency range 0.01 to 80 Hz. The traces have been corrected for instrument response and the spectra have been smoothed five times. The same time periods as identified by Petersen et al. (2011) for “low” and “high” noise conditions were analysed. The period between 8:00 and 18:00 UTC of 27 June 2009 (New Zealand night time) was characterised by “low wind noise” (wind velocities <10 km/h; Petersen et al. 2011), no teleseismic events within the previous 24 hours and no regional earthquakes of $M \geq 3$. In contrast, “high” noise conditions prevailed on 29 June 2009 between 8:00 and 12:00 UTC

(mean wind speed 22.3 km/h). Petersen et al. (2011) determined spectra for the whole ten- and four-hour periods. Since noise spectra can be straightforwardly obtained with SEISAN for periods ≤ 35 min but not for longer time windows, 35 min of each hour in the given time periods were used to calculate spectra. These spectra were then stacked and the average and standard deviation determined. The results are shown in Fig. 1.10 (a), (b) and (c). A comparison of the results from Petersen et al. (2011) and those obtained in this study for station FOZ (Fig. 1.11) shows only minor differences. Therefore, the average spectra obtained in this study are inferred to be comparable with those obtained for longer time periods.

Reference models for the noise power spectra of a station are the Peterson New High and New Low Noise Models (NHNM and NLNM; Peterson 1993) which were obtained from globally recorded ambient seismic noise in different frequency bands on the vertical component. All SAMBA stations show a log-linear trend for frequencies below 0.1 Hz. It is not clear whether this is an artefact due to the limited sensitivity of the short-period sensors in this frequency range or generally indicates high regional noise levels below 0.05 Hz. All stations exhibit relatively high noise levels in the microseism band (frequencies of 0.1–0.5 Hz), likely reflecting their proximity to the Tasman Sea. For frequencies above 1 Hz, the SAMBA stations EORO, GOVA and FRAN are closest to the NLNM. The remaining stations show similar levels of -140 dB. Stations COSA, MTFO, WHYM, REYN and LABE feature a local maximum in the 3–15 Hz frequency range. Suggestions for sources for the noise in this frequency range are given in Table E.1. At POGR, the noise is elevated at around 0.4 Hz. Since this station is the only station located in a deep, sediment-filled valley, this may be a resonance frequency (corresponding to sediment layer thicknesses of 340–450 m using the approach of Parolai et al. 2002).

GeoNet station FOZ was identified by Petersen et al. (2011) as a “good” station with low noise levels over the entire frequency band under both “low” and “high” noise conditions (Fig. 1.11). The spectra of WVZ and RPZ are comparable to that of FOZ under “low” noise conditions but are elevated under “high” noise conditions.

For the time period of “high” noise conditions, the acceleration power spectral density in the frequency range 0.1–1 Hz is elevated for all SAMBA stations. This is most distinct at station COSA and the GeoNet stations RPZ and WVZ. Station LABE was not recording on that day, so no comparison is possible for this station.

A comparison of the spectra from borehole and surface sensors was obtained for the

Figure 1.10. (a) Power spectral density for “low” (left) and “high” (right) noise conditions for the shallow SAMBA stations. Each panel shows spectra for all three channels (Z,N,E) with the title specifying date and station name. See (c) for more detail.

Figure 1.10. (b) Power spectral density for “low” (left) and “high” (right) noise conditions for the shallow SAMBA stations. Each panel shows spectra for all three channels (Z,N,E) with the title specifying date and station name. See (c) for more detail.

Figure 1.10. (a–b) Power spectral density for “low” (left; 27 August 2009) and “high” (right; 29 August 2009) noise conditions for the shallow SAMBA stations. The average and the standard deviation of the stacked noise PSD resulting from stacking of several 35 min-long data windows are shown for the vertical (Z, red), the north (N, blue) and the east-component (E, green). SH specifies the short-period seismometer recording on high gain. Dashed and dotted lines show the New High and Low Noise Model of Peterson (1993). (c) The same for the deep borehole sensor FRAN (for which due to the unknown orientation channel 1 (blue) and 2 (green) are shown instead) and the shallow station LABE. The latter was not operating on the 29 August 2009 and cannot be assessed for high noise conditions on that date. Note that stations WHAT and POCR are shown in Fig. F.1 and 1.12.

Figure 1.11. Power spectral density for “low” (left) and “high” (right) noise conditions at the GeoNet station FOZ (RPZ and WVZ shown in Fig. F.2). For this station a comparison with results by Petersen et al. (2011) is shown (lower panel) which indicates that the stacked 35 min-long data windows are comparable with 6 hr-long spectra obtained by Petersen et al. (2011).

borehole sensors at station WHAT and POCR. This shows that the surface sensors have elevated noise levels over the entire frequency range. In particular, for frequencies of 0.5–1.5 Hz at POCR, the background noise is comparable to the NHNM. Also the variation for frequencies above 10 Hz is larger (Fig. 1.12).

Between November 2008 and June 2009, problems were encountered with the Reftek data logger firmware at the SAMBA stations. Data acquisition failures occurred after random periods of time. Consequently, not all stations were recording data at the same time (Fig. C.1 for more detail). Fortunately, throughout this period at least four stations were recording simultaneously.

Almost every SAMBA station, which is equipped with a solar panel, recorded regular spikes in the data generated by the solar controller when charging the batteries (pulse-modulation-width charging). Different solar controllers were tested, the sensor and power cables were separated and additional shielding added, and various filters applied. Despite these measures, the spikes in the data could not be eliminated. Solar controllers from five different suppliers were tested. The spikes could only be avoided with two types of solar controllers, which were both expensive, difficult to obtain and not designed for use in the field (see Table D.1 for more detail). Most of the SAMBA stations are now equipped with ProStar 15 and 30 solar controllers, which generate double-spikes every six minutes. These noise spikes are annoying but only influence the signal if they coincide with an earthquake.

Despite the difficulties encountered with the setup and the operation of the stations in their remote locations in the central Southern Alps, the overall performance of the SAMBA array is good. The primary objective of recording earthquakes down to magnitude $M_L \sim 0$ on and in the vicinity of the Alpine Fault has been obtained. This has enabled the detection of as many earthquakes within one year as were recorded by the permanent national station network within 13 years. Experience with drilling shallow boreholes and recording with borehole seismometers has been gained. Initial problems with the Reftek data loggers are now solved, and only one station (REYN) could not be maintained further after nine months of operation. The remaining stations are all still in operation. Stations MTFO, WHYM and GOVA have been performing particularly well and recording continuously since installation. The more problematic stations are EORO, LABE and COVA. These stations have difficult access, little sun and/or a lot of snow in winter, and were therefore non-operational for parts of the year. Individual station performance of the remaining stations is assessed in Table E.1. Table I.1 provides an overview over the source dimensions and the expected recording range of earthquakes regarding the distance to the source (and the station spacing).

Figure 1.12. Power spectral density for a pair of borehole (left) and surface sensors (right) at the same station at “low” noise conditions. Each panel shows spectra for all three channels (Z,1,2 due to the unknown orientation) with the title specifying date and station name.

With a recording time of over 3.5 years and continuing, the SAMBA stations have enabled the longest microseismicity study in the region of high topography in central South Island to date. Several large earthquakes were recorded from around the globe in 2009–2012 (Table 5.3), including the two most recent large events in New Zealand, the M_W 7.8 Fiordland earthquake of 15 July 2009 and the M_W 7.0 Christchurch earthquake of 3 September 2010 and their aftershock-sequences.

2. Theoretical considerations

The primary characteristics with which we commonly describe an earthquake are hypocentre coordinates (latitude, longitude and depth), origin time and energy release. Secondary parameters include the rupture mechanism and the driving stress field, which describe the nature of the earthquake and its tectonic setting. How these quantities are estimated from seismic observations is summarised in this chapter. Related topics such as weighting of the phase arrivals, uncertainty estimates of the hypocentre location and determination of the velocity model using seismic tomography are also explained.

2.1. Earthquake location

In the following, the term “earthquake location” refers to the process of determining the hypocentre parameters. Earthquake location involves two parts; the forward problem of computing theoretical travel-times from a given velocity model and the inverse problem of determining the unknown parameters representing the hypocentre. If we assume the simple case of a homogeneous medium with a constant wave speed v , the arrival times t of the earthquake waves at i observation points x, y, z can be calculated from the equation

$$t_{\text{cal}}^i = T + \frac{1}{v} \sqrt{(X - x_i)^2 + (Y - y_i)^2 + (Z - z_i)^2} \quad (2.1)$$

with the earthquake origin time T and the hypocentre coordinates $\mathbf{P}=(X, Y, Z)$. This forward calculation can be performed for all possible hypocentres (e.g. nodes on a grid spanning a specific region), or an initial hypocentre $\mathbf{P}_0=(X_0, Y_0, Z_0)$ is assumed (for example, the coordinates of the station with the earliest arrival times) to approximate the true hypocentre. The calculated arrival-time is then compared with the observed arrival-time at each station in order to determine the actual hypocentre coordinates. The best hypocentre is that which minimises the differences between observed and calculated arrival times (the so-called arrival-time residuals):

$$r_i = \Delta t_i = t_{\text{obs}}^i - t_{\text{cal}}^i \quad (2.2)$$

Equation 2.1 clearly shows that non-linearity is an intrinsic characteristic of the earthquake location problem (Lomax et al. 2009).

2.1.1. Geiger's location method

If the initial hypocentre guess is close to the actual hypocentre, then the arrival-time residuals can be mathematically expressed by a first-order Taylor series about $\mathbf{P}_0$,

$$r_i = t_{\text{obs}}^i(\mathbf{P}) - \left(t_{\text{cal}}^i(\mathbf{P}_0) + \frac{\partial t_i(\mathbf{P}_0)}{\partial \mathbf{P}}(\mathbf{P} - \mathbf{P}_0) + \dots \right) + (T - T_0) \quad (2.3a)$$

$$= \frac{\partial t_i}{\partial x_i} \Delta x + \frac{\partial t_i}{\partial y_i} \Delta y + \frac{\partial t_i}{\partial z_i} \Delta z + \Delta t_0 \quad (2.3b)$$

in which only the linear terms are retained (since the corrections are small). The term Δt_0 describes the correction needed to adjust the origin time estimate. For many stations, this yields a linear system of equations of the form

$$\mathbf{r} = \mathbf{G}\Delta\mathbf{p}, \quad (2.4)$$

where $\mathbf{r}$ is the residual vector, $\mathbf{G}$ is a matrix of partial derivatives of the arrival-time to each station with respect to the hypocentre coordinates and an additional column of ones corresponding to the origin time correction term. The vector $\Delta\mathbf{p}$ contains the hypocentre and origin time corrections. The earthquake hypocentre solution can be iteratively found by adding the small changes $\Delta\mathbf{P}$ to successive hypocentre estimates. This inversion method originates from Geiger (1910). At least four observations of arrival-times for the recording stations are required to determine both the earthquake hypocentre and the origin time. Often many more stations record the earthquake, so the linear system is overdetermined. The most common technique applied for solving the system of equations is least squares with or without damping; however, several other methods exist (e.g. Buland 1976; Thurber 1985).

A linear approximation of the arrival-time residuals is only adequate for weakly non-linear problems (e.g. Tarantola and Valette 1982b). This is the case if the stations are not too far away from the earthquake hypocentre and the seismic velocities are slowly varying with position. If so, the matrix containing the partial derivatives $\mathbf{G}$ in equation 2.4 can be inverted because it is of full rank. However, when all stations are far away from the earthquake, then the observations are insufficient to fully constrain the spatial location. In this ill-conditioned case, linear inversion becomes unstable (a small change in the input produces a large, physically implausible change in the output; Tarantola and Valette 1982a). To account

for this, the depth is commonly fixed to a reasonable value during standard earthquake hypocentre determination with sparse station coverage. The mathematical explanation for instability of linear inversion is that the hypocentre location and the origin times are not independent variables but the least squares technique requires that all variables are strictly independent or can be made so (James et al. 1969).

2.1.2. Probabilistic non-linear location methods

Probabilistic non-linear locations of earthquakes most accurately describe the hypocentre uncertainty and can be used with complicated, 3-D velocity structures. Probabilistic methods make use of all available *a priori* information about the unknown parameters (as, for example, by specifying bounds), and about the distribution of the data (e.g. by assuming a Gaussian distribution). The following approach was described by Tarantola and Valette (1982b) to solve a non-linear inverse problem for a physical system Ξ parameterisable with a finite set of parameters $\mathbf{X}=\{X_1,...,X_m\}$ in $\mathbb{R}^m$. Any specific set of parameters is called a model $\mathbf{x}=\{x_1,...,x_m\}$ and any degree of knowledge about the parameters can be described by a density function $f(\mathbf{x})$. Two states of information, which have been obtained independently and can be represented by corresponding density functions $f_i(\mathbf{x})$ and $f_j(\mathbf{x})$, can be combined to produce a new state of information. The conjunction or the final state of information $f(\mathbf{x})$ can be expressed as

$$f(\mathbf{x}) = \frac{f_i(\mathbf{x})f_j(\mathbf{x})}{\mu(\mathbf{x})}, \quad (2.5)$$

where $\mu(\mathbf{x})$ is the null information (or neutral element), representing the state of total ignorance that is often described with a uniform distribution.

Tarantola and Valette (1982b) transformed this relationship to incorporate probability density functions for the data and model parameters and derived a probabilistic formulation of the non-linear earthquake location problem. Three density functions are necessary to find the *a posteriori* density function of the hypocentre solution:

- $\rho(\mathbf{d},\mathbf{p})$, which contains a combination of *a priori* information about the data $\mathbf{d}$ and the model parameters $\mathbf{p}$;
- $\theta(\mathbf{d},\mathbf{p})$, which contains the theory and errors caused by it, and;
- $\mu(\mathbf{d},\mathbf{p})$, which represents the null information.

The data used to determine the spatial position and the origin time of an earthquake are the phase arrival-times at observing stations. *A priori* information about the distribution of

arrival-time residuals and arrival time uncertainty can be specified by assuming Gaussian statistics. The velocity model is predefined and its influence on the hypocentre location is only accounted for by assuming uncertainties in the arrival times obtained by the theoretical forward computation of arrival-times. *A priori* information about the spatial location of the earthquake could be a depth or a lateral constraint (e.g. that the earthquake did not occur above the ground or outside a certain region). The null information on the spatial location is that the earthquake occurred in all regions of equal volume with the same probability. Generally, no *a priori* information about an earthquake's origin time, which is independent of the data, is available. So, a uniform *a priori* probability density function is assumed for the origin time. The theoretical relationship between calculated arrival times and the hypocentre position can be derived from wave propagation theory and the velocity model.

The resulting *a posteriori* probability density function corresponding to equation 2.5 is

$$\sigma(\mathbf{d}, \mathbf{p}) = \frac{\rho(\mathbf{d}, \mathbf{p}) \theta(\mathbf{d}, \mathbf{p})}{\mu(\mathbf{d}, \mathbf{p})}. \quad (2.6)$$

Integration over the data space leads to the probability density function in the model space:

$$\sigma_p(\mathbf{p}) = \int \sigma(\mathbf{d}, \mathbf{p}) d\mathbf{d} \quad (2.7)$$

A detailed description about how the *a priori* information is expressed as probability density functions was presented by Moser et al. (1992). After analytical integration over the data (Tarantola and Valette 1982b), the *a posteriori* density function is

$$\sigma_p(\mathbf{P}, T) = K \rho_p(\mathbf{P}, T) \exp \left[-\frac{1}{2} \left(\mathbf{t}^{\text{obs}} - \mathbf{t}^{\text{calc}} \right)^T (\mathbf{C}_t + \mathbf{C}_T)^{-1} \left(\mathbf{t}^{\text{obs}} - \mathbf{t}^{\text{calc}} \right) \right], \quad (2.8)$$

where K is a normalisation factor, $\rho_p(\mathbf{P}, T)$ is the *a priori* probability density function of the parameters, and $\mathbf{C}_t$ and $\mathbf{C}_T$ are the variance-covariance matrices of the observations and the velocity model, respectively (Section 2.2.2).

2.2. Arrival-time residuals and uncertainty assessment

The unknown spatial error of an earthquake's location is estimated by determining the uncertainty calculated using the travel-time residuals at all stations. Travel-time residuals typically result from measurement errors, misidentification of the phases, uncertainties in

the absolute time as recorded at the station (Stein and Wysession 2003) and shortcomings in the velocity model employed.

The lower limit of uncertainty in arrival time readings, with typical values of 5 to 20 ms, is governed by the sampling rate of the digitiser (Fremont and Malone 1987). The reading accuracy itself depends on the background noise level and the sharpness of the signal. S-phase arrivals are prone to phase misidentification because the phase onset is obscured by the P-coda. For the same reason, S-phase picks usually have larger uncertainties.

Systematic differences from the true phase arrival time can be expected when phases are identified using an automated process, whereas a random distribution around the true arrival time of an impulsive phase can be assumed when determined by eye. Also, any deviations of the true velocity along the whole ray path from the given model velocity are reflected in the travel-time residuals (e.g. Slunga et al. 1995). However, only the mean velocity effect can be seen, since velocity deviations can compensate for each other. Systematic errors due to unmodelled velocity structures, which are generally unknown, have been addressed in this study by a comprehensive analysis of the residuals (Section 3.1.1) and by a tomography study (Section 6.3).

It is standard to report the root-mean-squared residual σ_{RMS} as a estimate for the uncertainty of the observation. This parameter is obtained from the weighted sum of the travel-time residuals n at all stations:

$$\sigma_{\text{RMS}} = \sqrt{\frac{1}{n} \sum_{i=1}^n (w_i r_i)^2}. \quad (2.9)$$

Here w is the weight of the phase (for which $\sqrt{\frac{1}{n} \sum_{i=1}^n (w_i)^2} = 1$), and r the travel-time residual for the respective phase. If the travel-time residuals at different stations are of similar size, the root-mean-squared residual σ_{RMS} represents the typical travel-time residual for each station. In this case, σ_{RMS}^2 is approximately equal to the variance of the observations.

If not specified otherwise, S-waves will have larger *de facto* weights than P-picks in the travel-time inversion due to their lower velocity. As seen in equation 2.9, a single large residual not only influences the hypocentre location but also significantly alters σ_{RMS} . To prevent this, it is important to downweight large residuals of the travel-time arrivals (Section 3.1.2) which result from phase misidentification or timing errors.

2.2.1. Uncertainty in absolute earthquake locations

Linear inversion often inaccurately describes the uncertainty of the absolute hypocentre location. In particular, a single-point hypocentre solution, with a confidence ellipsoid at the 95% level, does not describe the uncertainty of the hypocentre location in all directions equally (e.g. Lomax et al. 2009). Due to the station configuration relative to the earthquake hypocentre, the uncertainty ellipsoids can be elongated in some directions compared to others. Under unfavourable circumstances, these ellipsoids underestimate the uncertainty of the hypocentre (e.g. Deichmann 1992).

2.2.2. Uncertainty assessment with probabilistic non-linear location methods

It is apparent that non-linear location methods mitigate errors introduced by the non-linearity of the hypocentre estimation problem. Moreover, there is generally no need to calculate partial derivatives of the travel-times relative to the hypocentre as required for linearised inversion methods. Probabilistic earthquake hypocentre determination thus allows us to describe the location accuracy more thoroughly by giving the full specification of the uncertainty volume (e.g. Moser et al. 1992). The uncertainty of the hypocentre is then represented by the spatial probability density function, which often shows remarkable differences in form and volume compared to an idealised ellipsoid (particularly for ill-conditioned problems; Lomax et al. 2000). Husen and Smith (2004) found that for ill-conditioned problems the difference between the hypocentre obtained with linearised location methods and the maximum likelihood hypocentre solution is large (outside the uncertainty estimate), implying that linear methods are insufficient in this case. For this reason, they used this difference as a measure of the location accuracy.

Equation 2.8 contains the uncertainties of the observations and the velocity model in the form of the combined variance-covariance matrices $\mathbf{C}_t$ and $\mathbf{C}_T$. This specific case only results if both uncertainty distributions are assumed to be Gaussian. The variance matrix of the observations $\mathbf{C}_t$ is defined as

$$(C_t)_{ij} = \begin{cases} (\sigma_i^2)^{\text{reading}} & \text{if } i = j \\ 0 & \text{otherwise} \end{cases} \quad (2.10)$$

where $(\sigma^2)^{\text{reading}}$ is the variance of the reading error (Moser et al. 1992). The theoretical model-error matrix contains the variances of the model on the diagonal and the covariances

on the off-diagonal entries.

In the non-linear location method by Lomax et al. (2000) applied in this study, the model-error covariance terms are calculated as

$$(C_T)_{ij} = (\sigma^2)^{\text{model}} \exp\left(-\frac{D_{ij}^2}{2L_{\text{corr}}^2}\right) \quad (2.11)$$

where σ^{model} accounts for an estimate of the unknown model error in the travel-time residual, L_{corr} is an estimate of the characteristic length scale of the velocity anomaly between two stations, and D describes the distance between two stations i and j . This matrix governs the extent to which a correlation in the uncertainties due to modelling exists between two stations. The value of σ^{model} is varied in proportion to the travel-time, so that a distant station would have a greater uncertainty due to the assumed velocity model than a nearby station. These parameters directly influence the size of the uncertainty volume and have been thoroughly evaluated as described in Section 4.1.1.

To evaluate σ^{model} , the estimates of σ_{RMS} and σ^{reading} are used. Assuming that the variance of an observation σ^{total} results from the variance of the phase reading and the variance due to unmodelled velocity structure, then:

$$\sigma_{\text{total}}^2 = (\sigma^2)^{\text{reading}} + (\sigma^2)^{\text{model}} \quad (2.12a)$$

$$\Leftrightarrow (\sigma^2)^{\text{model}} = \sigma_{\text{total}}^2 - (\sigma^2)^{\text{reading}} \quad (2.12b)$$

The estimate of the variance of an observation σ_{RMS}^2 is here used as a proxy for σ_{total}^2 . σ_{RMS}^2 is a fair representation of σ_{total}^2 for large numbers of phase picks. A mean value of σ_{RMS} can be derived from the typical location error (Section 3.1.1). Two independent estimates for the variance of the reading error σ^{reading} are obtained. First, we assume that the variance of the travel-time residuals from the mean is solely due to reading errors, whereas the mean value itself represents the error of the velocity model σ^{model} (Chapter 3.1.1). The second estimate for σ^{reading} comes from the variance of the time lag gained from cross-correlation of similar earthquakes in swarms (Section 4.2.1).

Nelson and Vidale (1990) found out that a small fraction of the absolute uncertainty arises from the finite-difference calculation of the travel times. They compared differences of the hypocentres for forward and backward calculations using the algorithm of Podvin and Lecomte (1991) and obtained average location uncertainties of 0.2 km in the horizontal plane and 0.3 km in depth due to differences in the computations. Husen et al.

(2003) estimated absolute location uncertainties using the probabilistic non-linear location program NonLinLoc (Lomax et al. 2000) by locating 34 quarry blasts and shots distributed throughout Switzerland. They obtained maximum epicentral uncertainties of ± 2 km and ± 5 km in focal depth with a typical station spacing of 30–50 km.

Lomax et al. (2009) performed simulations to test absolute uncertainties of earthquakes located on a vertical fault plane. This fault caused a sharp velocity contrast of 20% between adjacent media. Assuming uncertainties of ± 0.1 s and ± 0.2 s for synthetic P- and S-wave arrivals, respectively, they observed shifts in hypocentre locations of up to 5 km in depth and 2 km horizontally. Within the network, the true hypocentre locations were generally within the probabilistically determined location volume. In contrast, location uncertainties became large and irregular outside of the station array.

Wittlinger et al. (1993) performed synthetic location tests for a structurally complex setting of a gas reservoir with strongly heterogeneous media. They showed that hypocentre locations for vertical lineaments of events outside the station network became strongly clustered and were shifted by up to 4 km horizontally and 6 km in depth when located with the linearised location program HYPO71. Using a non-linear location method, the general pattern of the synthetic vertical lineaments outside the network could be correctly retrieved.

2.2.3. Influence of station configuration on location uncertainty

The shape of the uncertainty volume directly reflects the station distribution of the network. In general, an elongated epicentral uncertainty ellipse indicates a sparse station distribution in one direction relative to the hypocentre (Havskov and Ottemöller 2010). For an event lying outside the station network, the estimated hypocentre is shifted relative to the true hypocentre towards or away from the centre of the network depending on the assumed velocity structure (as seen in tests reported by Lomax et al. 2009). To prevent mislocation of earthquake hypocentres, testing of uncertainties regarding absolute location is essential. The hypocentral uncertainties for events outside the SAMBA network were tested by locating earthquakes in the same area, once outside the SAMBA network and later within the DFDP10 network (Section 3.4.2). This enables us to estimate the amount of shifting due to the station distribution relative to the earthquake.

Using different subsets of the station network to locate an earthquake can provide an estimate of how stable the event location is (James et al. 1969). Location and relocation of earthquakes also depend critically on the velocity model applied. This can be tested by locating the same set of earthquakes using different velocity models. The influence of anisotropy on the hypocentre location may also be non-negligible in some settings. The effect of velocity variations on the hypocentre location for the Southern Alps region is discussed in Section 6.3.1.

2.3. Weighting

A Gaussian uncertainty distribution is often assumed for arrival time phase picks (e.g. Pavlis 1986). This is valid for impulsive onsets of phase arrivals because picking uncertainties are randomly distributed around the true value (Buland 1976). Emergent arrivals, in contrast, are likely picked too late, depending on the signal to noise ratio, meaning that the uncertainty distribution tends to be asymmetric in time (Anderson 1982; Pavlis 1986). Travel-time uncertainties caused by an inadequate velocity model are often treated as Gaussian errors, simply because these model uncertainties are unknown (e.g. Pujol 2000). If the earthquake location is well-constrained (that is, when there are a suitable number of observations available from a range of distances and azimuths), the model uncertainty usually dominates the total location uncertainty. Otherwise, a bad phase pick can lead to a large location uncertainty. For well-located events, model uncertainties are systematic. A travel-time residual of 0.1 s resulting from an inappropriate velocity model corresponds to a typical hypocentre shift between 500 and 650 m for P-waves (e.g. Slunga et al. 1995) depending on the velocities.

Manual weights are often assigned depending on the sharpness of the phase onset and do not necessarily reflect its uncertainty. Assuming that the phase picks are randomly distributed around the true value, so that the distribution can be fitted by a Gaussian curve, a measure of the reading uncertainty can be calculated from the distance to the mean value (as applied in Section 3.1.2).

2.3.1. Jeffreys' weighting

Jeffreys (1973) proposed a weighting procedure for downweighting extreme travel-time residuals so that Gaussian statistics can be applied to the remaining residuals. This weight-

ing scheme uses the ratio μ of the number of extreme values relative to all values and the standard deviation σ_0 of the population without extreme values to reduce the influence of outliers on the location:

$$w(r) = \frac{1}{1 + \mu \exp\left(\frac{r^2}{2\sigma_0^2}\right)} \quad (2.13)$$

where w is the weight assigned to the travel-time residual r . To calculate σ_0 , outliers were defined as those travel-time residuals greater than two standard deviations σ of all travel-time residuals.

One problem with Jeffreys' weighting is that it introduces bias if the travel-time residuals do not have common variances (Section 3.1.1). Therefore, Jeffreys' function (equation 2.13) is effective for confining the influence of outliers but should not be used for weighting residuals unless they have common variances. The weighting scheme determined in this study is a combination of both Jeffreys' and systematic weighting (Section 3.1.2).

2.4. Cross-correlation

The cross-correlation function of two aperiodic functions $x(t)$ and $y(t)$ is defined as

$$C_{xy}(t) = \int_{-\infty}^{\infty} x(\tau)y(\tau+t)d\tau \quad (2.14)$$

(e.g. Havskov and Ottemöller 2010). Cross-correlation describes the process of shifting the function $y(\tau)$ relative to $x(\tau)$ by t for $-\infty < t < \infty$, multiplying the two functions and integrating the product. This functional computes the area under the curve which both functions have in common while moving one function along the t -axis. In other words, the cross-correlation function is a measure of the similarity of the two functions. If the two functions $y(\tau)$ and $x(\tau)$ are strictly identical the correlation is called the auto-correlation. Often the cross-correlation function is normalised by the auto-correlation functions

$$C_{xy}^N(t) = \frac{\int_{-\infty}^{\infty} x(\tau)y(\tau+t)d\tau}{\sqrt{C_{xx}(t)C_{yy}(t)}} \quad (2.15)$$

In this case, a cross-correlation coefficient for a particular time $t = t_0$ can be interpreted with regard to three key values:

$$C_{xy}^N(t_0) = \begin{cases} 0 & \text{the two functions are dissimilar,} \\ 1 & \text{the two functions are identical at that time,} \\ -1 & \text{the two curves are identical but one is flipped } (y(\tau) = -x(\tau)) \end{cases}$$

Cross-correlation is now used routinely to identify clusters of similar earthquakes (e.g. Nadeau et al. 1995), to determine the relative time for which the two events are most similar (lag-time), to detect low-frequency events within tremor (Shelly et al. 2007; Brown et al. 2008) or aftershocks immediately following the mainshock (Peng and Zhao 2009) and for high-precision earthquake relocation (e.g. Schaff et al. 2002; Du et al. 2004). Comparison of seismic waveforms over long periods of time can also be used to investigate temporal changes in the properties of a structure over time (e.g. Li et al. 2003).

The absolute value of the cross-correlation coefficient depends on filtering and the choice of time-window-length, which shifts the cross-correlation coefficients up and down by a constant amount. The lower the high-frequency content and the shorter the window length, the higher the cross-correlation coefficient (e.g. Nakahara 2004; Schaff et al. 2004).

Earthquakes with highly similar waveforms (cross-correlation coefficients larger than ~ 0.9) are called repeating earthquakes or multiplets, and are associated with repeated rupture of the same asperity (e.g. Nadeau et al. 1995; Gans et al. 2003). Waveform similarity decreases with increasing inter-event distance (Nakahara 2004; Schaff et al. 2004). For inter-event distances on the order of several kilometres, the correlation coefficients diminish rapidly (e.g. Hutchings and Wu 1990). Several authors (e.g. Nadeau et al. 1995) showed that high cross-correlation coefficients (>0.9) are associated with separation distances of only a few tens or hundreds of metres. This corresponds to source separations smaller or equal to one-quarter of the wavelength of the dominant wave frequency (Geller and Mueller 1980). However, the cross-correlation coefficient for events recorded at a common station depends strongly on the geological heterogeneities in the vicinity of the recording site (Hutchings and Wu 1990), the degree of velocity structure heterogeneity along the ray path (Schaff et al. 2004), and on the heterogeneous structure in the earthquake source region. Nakahara (2004) used the event separation distance above which cross-correlation coefficients fall below 0.6 to characterise the damage zones of two large earthquakes in Japan. He found that the structure around the source region of the Awaji earthquake is more heterogeneous (smaller correlation distances) than in the Kobe region.

2.5. Earthquake relocation

Relative locations of earthquakes can be determined more accurately than absolute locations (e.g. Schaff et al. 2002). The reason for this is that differential travel-times of two closely spaced events, which are recorded by the same distant station, can be precisely calculated using cross-correlation of the waveforms. Location uncertainties due to unmodelled velocity structure are diminished because the ray paths of these events are similar along most of the path and differ only in the source region (Waldhauser and Ellsworth 2000). Differential travel-times resolvable by cross-correlation are on the order of milliseconds and therefore at least ten times better than the corresponding reading uncertainties (cf. Section 2.2.1). This is why relative locations of spatially clustered events can resolve the fine-scale structure of the cluster, while the uncertainty in the absolute locations of individual events is too large to resolve the separation between these events.

A first step in relative earthquake location is to locate similar events relative to one master event which has a well-determined absolute location (e.g. Pujol 2000). This is known as the master-event technique. A more sophisticated method is to locate all possible pairs of similar events relative to each other. The spatial location of an individual event is then constrained by its distance to all other surrounding events.

It has been argued by Menke and Schaff (2004) that relative relocation of earthquakes can, in principle, be used to determine absolute locations, provided that the station coverage and data quantity are excellent. Several authors have investigated how relative locations can constrain absolute locations. Slunga et al. (1995) reported that systematic arrival time differences between similar events at a 6 km distant station changed by 1 ms if a pair of events was moved by 200 m while retaining their relative locations, but this finding is clearly velocity dependent.

2.5.1. Uncertainty assessment of relative earthquake locations

The most commonly used program for relative earthquake locations is the double-difference algorithm hypoDD of Waldhauser and Ellsworth (2000). Dunn (2004) tested whether systematic and random errors can be identified using hypoDD assuming a synthetic data set of five earthquakes aligned north-south and separated by 1 km. Systematic travel-time errors at a few stations become apparent using hypoDD. Random errors (ranging between

0 and 0.7 s) added to the theoretical travel-times affected both the epicentre and the depth. Epicentre locations were shifted approximately 1 km, whereas the effect on the depth of the earthquakes was on the order of 2.5 km. However, the overall linear feature could be resolved and event distances between the earthquakes remained approximately constant.

During the relocation process, the weighted sum of the squared double-difference residuals for pairs of earthquakes at each station is minimized. This can be accomplished using either the singular value decomposition (SVD) or the conjugate gradients method (LSQR). SVD reliably resolves relative location errors but can only be used for small datasets. Large datasets can be solved efficiently with by LSQR but a location error assessment is unreliable (Waldhauser and Ellsworth 2000).

2.6. Amplitude decay

The amplitude of a seismic wave diminishes as the wave propagates due to the redistribution of energy. With distance Δ from the wave's source the amplitude A of a seismic wave decays as

$$A(\Delta) = GA_0\Delta^{-\alpha}\exp(-\gamma\Delta) \quad (2.16)$$

where α is the geometric spreading coefficient, and γ is the anelastic attenuation coefficient (describing the absorption of seismic energy) for a specific region (Stein and Wysession 2003). G is the instrument gain and A_0 is the amplitude of the wave at the source ($\Delta=0$). Note that equation 2.16 neglects the effect of scattering. The anelastic attenuation parameter γ is known to be frequency dependent: $\gamma = \gamma(f)$. This frequency dependence originates from the relationship between γ and the quality factor Q (e.g. Kim 1998),

$$\gamma(f) = \frac{\pi f}{Q(f)v}, \quad (2.17)$$

where v is the velocity for P or S-waves, $Q(f) = Q_0 f^v$ is a function of f containing the quality factor Q_0 at a reference frequency $f=1$ Hz and v is the frequency exponent.

The local magnitude $M_L = \log A_0 + \text{const.}$ is obtained by taking the logarithm of equation 2.16 and rearranging, which results in equation 2.23 (Section 2.7).

2.6.1. Geometrical spreading and attenuation

Geometrical spreading of the wave describes the amplitude decay of the expanding wavefront in space, which causes increasing amplitude reduction with distance from the wave's source (Stein and Wysession 2003). The geometrical spreading coefficient α reflects the exponent of the amplitude decay with distance. This value is usually constrained based on theoretical considerations. The effects on the amplitude of wave-expansion in space can be shown by starting with the total mean energy density of a plane wave:

$$\bar{E} = \frac{1}{2} \rho \omega^2 A^2, \quad (2.18)$$

where ρ is the density of the medium, $\omega = 2\pi f$ the radial frequency or frequency f , and A the wave amplitude. The energy flux through an areal element is given by:

$$\begin{aligned} \text{Energy flux per areal element} &= \text{energy density} \times \text{velocity} \times \text{areal element} \\ d\bar{E}^{\text{flux}} &= \frac{1}{2} \rho \omega^2 A^2 \times c \times ds \end{aligned} \quad (2.19)$$

Since the energy flux through two areas $d\bar{E}_1^{\text{flux}} = d\bar{E}_2^{\text{flux}}$ in a homogeneous medium must be the same for a spreading wave (ρ, ω, c all constant), the amplitude ratio corresponding to different areas is

$$\frac{A_2}{A_1} = \sqrt{\frac{ds_1}{ds_2}} \approx \frac{r_1}{r_2} \quad (2.20)$$

The area of the expanding spherical wavefront increases with $4\pi r^2$, where r is the radius of the sphere. Therefore, for body waves travelling directly from the source to the receiver, the amplitude decay is proportional to $1/r$, and $\alpha=1$. For distances greater than 100 km, the first waves commonly recorded at the receiver are refracted waves, which show a different amplitude decay with $\alpha \sim 2$ (Stein and Wysession 2003). Surface wave decay is proportional to the circumference increase of a growing circle in two dimensions which results in $\alpha=0.5$ (Stein and Wysession 2003).

The second contribution to the amplitude decay results from attenuation, which can significantly decrease the amplitude. Crustal attenuation results from both intrinsic and scattering attenuation. Which mechanism dominates and what factors govern attenuation remain debated and seem to vary with tectonic settings (e.g. Frankel 1991; Adams and Abercrombie 1998). Aki and Richards (1980) suggested that attenuation is higher in regions of tectonic activity. Frankel (1991) showed that anelastic scattering is greater in southern California than in New York or South Africa, but noted that regional differences in attenuation could rather reflect the geologic complexity than tectonic activity.

In general, local lithological variations, inhomogeneities, the state and amount of fractures, crustal fluids and temperature variations affect attenuation in the crust (Hauksson and Shearer 2006). The effect of temperature becomes prominent at temperatures above the solidus of the rock (e.g. Mavko 1980). Consequently, magma and partial melts are highly attenuative. In non-volcanic settings, however, where temperatures in the crust are below the solidus temperature, the effect on attenuation is much smaller than the effect of major tectonic structures on attenuation (e.g. Hauksson and Shearer 2006). Thus, attenuation in the crust is related to some degree to the permeability and the amount of fracturing of the rock, its fluid content and fluid pressure (Frankel 1991).

Geometrical spreading and scattering attenuation are elastic processes because the energy of the wave is redistributed but conserved. Scattering describes the redistribution of seismic energy at inhomogeneities, which shifts energy from the direct wave into the coda (Frankel 1991). Intrinsic attenuation, in contrast, is an anelastic process associated with mechanisms that convert seismic energy into heat. The wave's energy is transformed to heat by internal friction along grain boundaries (Stein and Wysession 2003). Therefore, attenuation increases with every oscillation of the wave. The longer the path through the crust, the more depleted the spectrum of high frequencies. For direct P- and S-waves, the quality factor Q_P (inverse of the attenuation coefficient) for P-waves is generally two to three times higher than Q_S . Knopoff (1971) derived $Q_P \sim 2.25 Q_S$ for a Poissonian solid under the assumption that little energy is lost in compression (bulk modulus is large). However, Hauksson and Shearer (2006) observed $Q_S/Q_P \approx 1.3$ and related this to fluid saturation of the rock. The near-surface structure of the crust strongly influences the attenuation of the seismic wave. A high percentage of crustal attenuation can be attributed to the uppermost few kilometres (< 3 km; Abercrombie 1998). Q near the surface is low (in the range of 10 to 100), independent of rock type and frequency (Abercrombie 1998, and references therein).

2.7. Earthquake magnitudes

Earthquake magnitude is a logarithmic measure of earthquake size. Magnitudes can be calculated from the duration of the coda (length of decay of the scattered waves of a seismic phase), maximum amplitude or amplitude of a phase at a certain frequency.

Richter's 1935 original definition of the local magnitude M_L included a station cor-

rection term S (accounting for the instrument gain and the site amplification due to the local geology) and a distance correction term $-\log(D_0)$. This term was defined in such a way that a magnitude 3 event would produce an amplitude of 1 mm on a Wood-Anderson torsion seismograph at 100 km distance from the source:

$$M_L = \log(A(\Delta)) - \log(D_0(\Delta)) + S \quad (2.21)$$

The amplitude A was measured as half the peak-to-peak displacement value determined from the maximum amplitude on the horizontal components of a Wood-Anderson seismograph. To separate the effects of geometrical spreading (proportional to $\log(\Delta)$) and scattering and anelastic attenuation (proportional to Δ) along the path to the receiver, the distance correction term $-\log(D_0)$ can be replaced by

$$-\log(D_0(\Delta)) = \alpha \log(\Delta) + \gamma \log(e)\Delta. \quad (2.22)$$

Thus the local magnitude is

$$M_L = \log(A(\Delta)) + \alpha \log(\Delta) + \gamma \log(e)\Delta + S \quad (2.23)$$

where α is the geometrical spreading coefficient and γ the anelastic attenuation factor for a specific region. This formula is commonly used for shallow local events recorded by stations at distances of ≤ 1000 km.

To calibrate the local magnitude M_L in a certain region, the coefficients α and γ have to be determined using regression analysis (e.g. Hutton and Boore 1987; Uski and Tuppurainen 1996). For the Wellington region, Robinson (1987) derived the relationship

$$M = \log(A(\Delta)) + 1.0 \log(\Delta) + 0.0029\Delta + S \quad (2.24)$$

which is currently used by GeoNet to estimate magnitudes all over New Zealand. O' Keefe (2008) determined a magnitude scale for the central Southern Alps region using 411 microearthquakes recorded in late 2006:

$$M = \log(A) + 1.0 \log(\Delta) + 0.0073\Delta + S \quad (2.25)$$

This result indicates higher attenuation in the Southern Alps than accounted for in the magnitude determinations by GeoNet. A new magnitude scale for the Southern Alps region is derived in Section 3.2.

2.8. Gutenberg-Richter relationship and b-value

The cumulative number of earthquakes $N(M)$ for any magnitude M larger than the magnitude of catalogue completeness M_c obeys an exponential distribution with increasing magnitude with an exponential decay parameter λ :

$$N(M) = N_0 e^{-\lambda(M-M_c)} \quad (2.26a)$$

$$\log N = \underbrace{\log N_0}_a - \underbrace{\lambda \log e}_{b} (M - M_c) \quad (2.26b)$$

$$\log N = a - b (M - M_c) \quad (2.26c)$$

The magnitude of completeness or cut-off magnitude is the lowest magnitude for which all events are detected in time and space (e.g. Woessner and Wiemer 2005). Equation 2.26c is known as the magnitude–frequency relationship and was first determined empirically by Gutenberg and Richter for earthquakes in California (Gutenberg and Richter 1944). The constant $a = \log N_0$ is specific to a certain region and corresponds to the logarithm of the number of earthquakes at the reference magnitude M_c . For this reason, a is also called the rate of occurrence or the productivity of a region. The value $b = \lambda \log e$ represents the slope of the relationship between the logarithm of the cumulative number of earthquakes N and their magnitudes $M \geq M_c$. In other words, the b-value describes the expected number of earthquakes of low relative to high magnitudes. A b-value close to unity represents a 10-fold increase in the number of earthquakes with every decrease of one magnitude unit. A b-value lower than one indicates that a higher number of larger events can be expected compared to the reference model with $b=1$. For example, if 10000 events with $M \geq 4$ occurred and the b-value of the Gutenberg-Richter distribution is $b=1$, than 10 earthquakes of $M \geq 7$ can be expected, whereas for $b=0.9$ this results in 20 expected events with $M \geq 7$. Low b-values have been described as indicators for locked fault segments or high stress areas (e.g. Wiemer and Wyss 1997), or have been associated with the maturity of the fault system (Stirling et al. 1996).

To determine the b-value accurately, fitting of the slope of the Gutenberg-Richter relationship with linear least squares is insufficient because the b-value depends on the cut-off magnitude M_c and the errors of the magnitudes M . Therefore, the maximum likelihood estimate of b has to be determined instead (Aki 1965). The maximum likelihood function L for a set of n magnitudes M_j above the cut-off magnitude M_c which follow an exponential

distribution as seen in equation 2.26a is

$$L = \prod_{j=1}^n \lambda e^{-\lambda(M_j - M_c)} \quad (2.27a)$$

$$= \lambda^n e^{-\lambda[(M_1 - M_c) + (M_2 - M_c) + \dots]} \quad (2.27b)$$

$$\log L = n \log \lambda - \lambda \log e \sum_{j=1}^n (M_j - M_c) \quad (2.27c)$$

At the maximum, $\partial \log L / \partial \lambda = 0$ and hence

$$0 = \frac{n \log e}{\lambda} - \log e \sum_{j=1}^n (M_j - M_c) \quad (2.28a)$$

$$\frac{n}{\lambda} = \sum_{j=1}^n (M_j - M_c) \quad (2.28b)$$

$$\lambda = \left(\frac{1}{n} \sum_{j=1}^n (M_j - M_c) \right)^{-1}. \quad (2.28c)$$

Since $b = \lambda \log e$ it follows

$$b = \frac{\log e}{\bar{M} - M_c} \quad (2.29)$$

with the mean magnitude $\bar{M}$ for $M \geq M_c$. The uncertainty of the b-value reduces with the uncertainty of the mean magnitude. The larger the number of events n with a magnitude, the smaller the standard deviation of the mean $\sigma_n = \sigma / \sqrt{n}$. Therefore

$$\sigma_b = \frac{b}{\sqrt{n}} \quad (2.30)$$

To avoid errors in b-value calculation, a sufficiently large data set of well-located earthquakes are required (more than 2000 events are needed to calculate b to within an accuracy 0.05 at the 98% confidence level, Felzer 2006).

2.9. Seismic tomography

Seismic tomography aims to resolve 3-D velocity structure based on the assumption that the model uncertainty dominates the location uncertainty of well-recorded events. This is often reflected by the station correction terms. Consequently, a systematic analysis of travel-time residuals can provide information about the velocity structure between the earthquake source and the recording stations. Negative residuals indicate higher velocities and positive

residuals lower velocities in relation to the reference velocity model. Lateral variations that are not considered in the assumed velocity model can be studied using this approach. Those velocity anomalies can be located if many, preferably crossing ray paths between different sources and receivers are available.

In seismology, tomography studies are performed to image 2-D or 3-D structures of the ground using earthquake waves. Based on Fermat's principle that the ray path represents the path of minimum travel-time for direct waves, the travel-time τ of a seismic wave results from the integrated velocity $v(s)$ of the wave along a ray path s through an isotropic medium

$$\tau = \int ds/v(s) \quad (2.31)$$

or alternatively expressed using the slowness $u(s) = 1/v(s)$:

$$\tau = \int u(s)ds \quad (2.32)$$

The arrival-time t of a body-wave generated at the source k and recorded by the receiver i is therefore

$$t_i^k = T^k + \int_k^i u ds \quad (2.33)$$

with the origin time T . The arrival-time residual r is calculated as the difference between observed and calculated arrival-times and is attributed to velocity perturbations along the ray-path:

$$r_i = t_i^{\text{obs}} - t_i^{\text{cal}} \quad (2.34)$$

Linearisation of equation 2.34 around an initial hypocentre and velocity model using a truncated Taylor series (as in equation 2.3a) yields

$$r_i^k = \sum_{l=1}^3 \frac{\partial t_i^k}{\partial x_l^l} \Delta x_l^l + \Delta T^k + \int_k^i \delta u ds \quad (2.35)$$

For arrival-time tomography, the volume between the source and the receivers is divided into equal-sized blocks with a velocity perturbation Δu_j for each block,

$$u_j = l_j \Delta u_j \quad (2.36)$$

with the path l_j in block j . For a single source–receiver configuration the velocity perturbation $\Delta u(s)$ can occur anywhere along the ray path. When there are enough ray paths and different source–receiver configurations, those blocks with different velocities can be

identified. However, this is only possible if a certain number of rays propagate through each of the blocks. Also, the rays have to sample the same block on a number of ray paths in different directions, so that the structure can be resolved.

Using the discretisation of the ray path into ray segments within a block,

$$\int_k^i \delta u ds = \sum_{j=1} G_{ij} \Delta u_j, \quad (2.37)$$

equation 2.35 can be expressed in matrix form as

$$\mathbf{r} = \mathbf{G}\Delta\mathbf{p} + \mathbf{H}\Delta\mathbf{m}, \quad (2.38)$$

where $\mathbf{G}$ and $\mathbf{H}$ contain the partial derivatives of the hypocentre and model part and $\Delta\mathbf{p}$ and $\Delta\mathbf{m}$ are the vectors of combined hypocentre/origin time and velocity model adjustments (Kradolfer 2003). The simultaneous inversion for hypocentre locations and velocity structure is a mixed-determined problem (Kradolfer 2003). This means some of the unknown model parameters are overdetermined while others are underdetermined and cannot be uniquely estimated. Therefore, no unique solution exists because too little information is available for some of the model parameters. Damping, smoothing and *a priori* constraints are needed to find a solution of the model parameters. If the solution is underdamped, the residual vector is minimised but not the solution vector. In the overdamped case the over- and underdetermined model parameters are minimised but the solution does not reflect the smallest residual vector. The solution of a mixed-determined problem using damped least squares solution is

$$\Delta\mathbf{m}^{\text{est}} = (\mathbf{G}^T\mathbf{G} + \theta\mathbf{I})^{-1}\mathbf{G}^T\Delta\mathbf{d} \quad (2.39)$$

(e.g. Kradolfer 2003), which has to be solved iteratively.

In double-difference seismic tomography (Zhang 2003) as applied in this study, the observations of pairs of events are considered, and equation 2.35 becomes

$$r_i^k - r_i^h = \sum_{l=1}^3 \frac{\partial t_i^k}{\partial x_i^l} \Delta x_i^l + \Delta T^k + \int_k^i \delta u ds - \sum_{l=1}^3 \frac{\partial t_i^h}{\partial x_i^l} \Delta x_i^l - \Delta T^h - \int_h^i \delta u ds \quad (2.40)$$

This can be simplified if the events are sufficiently close in space so that the path from the sources k and h are similar except within the source region

$$dr_i^{kh} = r_i^k - r_i^h \quad (2.41a)$$

$$= \sum_{l=1}^3 \frac{\partial t_i^k}{\partial x_l^1} \Delta x_l^1 - \sum_{l=1}^3 \frac{\partial t_i^h}{\partial x_l^1} \Delta x_l^1 + \Delta T^k - \Delta T^h \quad (2.41b)$$

$$= (T_k^i - T_h^i)^{obs} + (T_k^i - T_h^i)^{cal} \quad (2.41c)$$

where dr_i^{kh} is the double difference between observed and calculated differential arrival times. The travel-time difference outside the source region is small compared to the difference inside the source region.

A joint inversion for absolute and relative locations as well as the velocity structure can be obtained using the double-difference seismic tomography program *tomoDD* of Zhang and Thurber (2003). It incorporates a special weighting scheme which includes distance weighting so that relative locations are only taken into account for closely spaced events and are suppressed for distant earthquakes. The program also explicitly accounts for path anomaly biases between event pairs (Zhang 2003).

2.10. Focal mechanisms

A focal mechanism characterises the geometry of the rupture on the fault plane of an earthquake. Three angles are used to characterise the mechanism and the orientation of the fault plane:

- the strike angle ξ describes the orientation of a horizontal line in the fault plane measured clockwise from north $0^\circ \leq \xi \leq 360^\circ$
- the dip angle δ describes the inclination of the fault plane from the horizontal $0^\circ \leq \delta < 90^\circ$ (where the dip direction is always measured to the right hand side looking along the strike)
- the rake angle λ specifies the direction of the hanging-wall movement with respect to the footwall $-180^\circ \leq \lambda \leq 180^\circ$ (measured in the plane of the fault from the strike direction, where positive values indicate upwards movement of the hanging-wall as during reverse faulting and negative values represent downward movement during normal faulting)

The polarities of the waves radiated by the earthquake and recorded at the stations are commonly used to determine focal mechanism solutions (Stein and Wysession 2003). If the first motion of the P-wave at a station is upwards the earthquake has caused compression of the source material in the direction to the receiver. If, in contrast, the polarity is downwards, the P-wave directed towards the receiver resulted from dilatation of the material around the source. S-waves show different polarisation characteristics and, in principle, can be used in the same way to determine the focal mechanism. However, the S-arrival is often emergent and the polarity obscured. Also, if shear-wave splitting occurred this has to be corrected for before the polarity of the S-wave can be studied (e.g. Zollo and Bernard 1989). The directions of the highest amplitude of compressional and tensional P-wave radiation can be derived from the compressional and dilatational quadrants and are described with the P-axis and T-axis.

Different mechanisms of faulting cause distinct radiation patterns of the waves emitted from the earthquake source (Fig. 2.1). The take-off angles of the ray paths from the source to the receivers are plotted on the focal sphere, a hypothetical unit sphere enclosing the source (Stein and Wysession 2003). Take-off angles on the upper hemisphere result from rays recorded near the source, and are projected onto the lower hemisphere. Most commonly, a focal mechanism solution is depicted on a lower-hemisphere projection of the focal sphere.

Figure 2.1. Principal stress directions, focal mechanism solutions and the orientation of faults after Anderson (1951). Compressional quadrants of the focal mechanism solutions are filled in black.

The focal mechanism solution determined from the far-field radiation alone contains an intrinsic ambiguity regarding the fault plane because the rupture plane is indistinguishable from the auxiliary plane. Other data, such as mapped faults, radiation patterns as a function

of frequency of the seismic waves or aftershock distributions must be used to interpret the fault plane (McKenzie 1969).

Mathematically, the far-field radiation pattern of shear-failure on a fault can be modelled using a double couple of equivalent body forces (Stein and Wysession 2003). A double couple describes two force pairs a small distance apart which point in opposite directions, so that the total moment vanishes. The nine components of the seismic moment tensor $\mathbf{M}$ represent all double couples of different orientation. Thus, the seismic moment tensor describing shear failure on a fault in a coordinate system aligned with the fault plane and its normal is always symmetric and contains two components of opposite polarity.

Uncertainty in the focal mechanism solution is caused by several factors of which some can be tested and eliminated, but others cannot. The focal mechanism solution is constrained by

- the number of available polarity readings and the coverage of the focal sphere, as the number of recording stations is always limited and the stations can be distributed unevenly around the earthquake hypocentre;
- the influence of falsely wired seismometers on the solution (this can be checked by comparing first motions of a teleseismic event recorded by all stations, since the same first motion is expected for all stations on a small aperture array in a large distance, Fig. B.1);
- the signal to noise ratio of the recorded waveforms; and
- the uncertainty in the take-off angle which results from the uncertainty in the hypocentre location and the inaccuracy of the velocity model.

Since most earthquakes are recorded by only a few stations, the spatial coverage on the focal sphere is often insufficient to determine the mechanism. Several different methods exist which use additional information to increase the number of observations per earthquake. Regional moment tensor inversion uses the whole waveform recorded by broadband stations and fits it with theoretical waveforms. This method is typically applicable for earthquakes above a magnitude threshold of $M_L \sim 3.5$. For the New Zealand region, focal mechanisms for earthquakes of $M_L \geq 4$ are determined as part of the routine earthquake analysis by GeoNet (Ristau 2008). Amplitude ratios of P- and S-waves can be used to determine the focal mechanism because of the radiation pattern of P- and S-waves from the source (e.g. Hardebeck and Shearer 2003, and references therein). P-wave amplitudes have maximum wave amplitudes near the P- and T-axes and minimum amplitudes near the nodal planes.

Maximum S-wave amplitudes are observed at the nodal planes (Fig. 2.2). Also, P- and S-wave amplitude ratio methods have the advantage that no corrections for magnitude and geometrical spreading are necessary.

Figure 2.2. P- and S-wave amplitude patterns for a double-couple source. Note that the slip vector lies in the fault plane whereas the fault normal vector lies in the auxiliary plane.

Probabilistic approaches based on Bayes' theorem can also be used for focal mechanism parameter inversion (Walsh et al. 2009). To characterise the uncertainty thoroughly, the resulting Bayesian posterior probability density function (PDF) of the focal mechanism parameters is parameterised by a generalised Matrix-Fisher distribution (Arnold and Townend 2007; Walsh et al. 2009). The Matrix-Fisher distribution describes the distribution of vectors on a sphere. Three matrices, M , U and D characterise the distribution. M controls the

orientation of the distribution, the spin matrix U twists the shape of the distribution at the nodes and the concentration matrix D describes the density of the distribution of vectors in one direction. A scalar Matrix-Fisher distribution allows for the simplest description of the distribution. Arnold and Townend (2007) used the scalar concentration parameter κ from the scalar Matrix-Fisher distribution to determine the standard deviation σ_{Θ} of the focal mechanism parameters

$$\sigma_{\Theta} = \exp(3.9155 - 0.5659 \log(\kappa)) \quad (2.42)$$

assuming equal errors in strike, dip and rake. However, the uncertainty of the focal mechanism solution is best represented graphically by the distribution of the P- and T-axes. The tighter the contours, the better the focal mechanism of the earthquake is constrained (Section 3.5).

2.11. Stress estimates

Assuming that individual focal mechanisms represent a uniform state of stress and are driven by the same stress field, regional stress estimates can be derived by inverting a set of independent focal mechanism observations as described by Gephart and Forsyth (1984), Michael (1984) and Arnold and Townend (2007). A focal mechanism solution enables the calculation of the orientation of the maximum, intermediate and minimum compressive stresses and a ratio of the magnitudes of the stresses for the tectonic stress field in a small crustal volume around the earthquake (Arnold and Townend 2007). Although six components are needed to fully describe the stress tensor, these four parameters suffice to calculate the direction but not the magnitude of maximum horizontal compressive stress (Lund and Townend 2007). Due to the free surface of the earth, which supports no shear stresses, one of the three principal stress axes is subvertical and this vertical stress increases linearly with depth ($\sigma_v = \rho gh$). However, topographic stresses can regionally disturb the stress field (e.g. Liu and Zoback 1992). If one of the principal stresses is not strictly vertical, the direction of the maximum horizontal compressive stress can differ by up to tens of degrees from the direction of the prevailing larger subhorizontal compressive stress (Lund and Townend 2007). Depending on the stress ratio $R = (\sigma_1 - \sigma_2) / (\sigma_1 - \sigma_3)$, small deviations of the direction of the principal stress axis from the vertical and horizontal, respectively, can be associated with substantial changes in the direction of the maximum horizontal compressive stress, especially for stress states in dominantly compressional or extensional tectonics (Lund and Townend 2007).

2.12. Mohr-Coulomb stress analysis

According to Anderson's theory of faulting, failure on new or preexisting faults is governed by the normal σ_n and shear stress τ acting on the fault plane, the coefficient of internal friction μ as well as the cohesive strength τ_0 of the rock. The cohesive strength is the inherent strength of the fault and corresponds to the minimum shear stress required for failure on new faults. It is zero for preexisting faults, but finite and positive if, for example, hydrothermal cementation has sealed the fault. This is expressed by the Coulomb frictional failure criterion:

$$|\tau| = \tau_0 - \mu \sigma_n \quad (2.43)$$

To account for fluid pressure this equation is modified to

$$|\tau| = \tau_0 - \mu(\sigma_n - P_f) \quad (2.44)$$

where P_f is the fluid pressure (Hubbert and Rubey 1959). The higher the fluid pressure for a given normal stress σ_n , the lower the shear stress τ the failure surface can sustain before failing.

A 2-D Mohr circle (with the intermediate principal stress σ_2 lying in the fault plane) provides a means of determining the normal- and shear-stresses on a fault plane of given orientation in the prevailing stress field. All possible combinations of normal and shear stresses given by

$$\sigma_n = \frac{(\sigma_1 + \sigma_3)}{2} + \frac{(\sigma_1 - \sigma_3)}{2} \cos(2\theta) \quad (2.45a)$$

$$\tau = \frac{(\sigma_1 - \sigma_3)}{2} \sin(2\theta) \quad (2.45b)$$

lie on a Mohr circle, where θ is the angle between σ_1 and the fault plane (Fig. 2.3). Failure occurs when the straight line that expresses the frictional failure criterion (equation 2.44) touches or intersects the circle. The formation of new conjugate faults occurs at angles of $\theta^* = 1/\tan(1/\mu) = 30^\circ$ (for $\mu = 0.6$) to σ_1 . Faults slipping with non-optimal orientation $\theta \neq \theta^*$ can be interpreted as having a low coefficient of friction $\mu' < \mu$ or increased pore-fluid pressure P'_f (Fig. 2.3 b) (e.g. Townend and Zoback 2004). If faults and fractures preexist, failure occurs on those favourably oriented structures first because $\tau_0 \sim 0$ and fault reactivation is possible for $\theta < 2\theta^*$ (Sibson 1985). Sibson (1989) showed that severely misoriented faults ($\theta > 2\theta^*$) can only be reactivated in fluid-overpressured regimes where $P_f \geq \sigma_3$.

Figure 2.3. Figure 1 of Townend (2006) illustrating the Coulomb frictional failure criterion (equation 2.44) and the angle under which failure occurs using a Mohr circle. (a) New faults form at an angle θ° to the maximum horizontal compressive stress σ_1 whereas (b) preexisting faults can be reactivated at a different angle $\theta > \theta^\circ$ in the case of reduced friction $\mu' < \mu$ or increased pore-fluid pressure P'_f .

In the tensile field $\sigma < 0$ (extensional stress), the failure envelope is given by the Griffith failure criterion

$$\tau^2 = 4T_0\sigma_n + T_0. \quad (2.46)$$

The tensile strength of rock is in the order of 1–4 MPa, about ten to hundred times lower than compressional strength. Tensile failure is possible at shallow depths (small confining pressures which allows tensile cracks to open; e.g. Frohlich 1994) or high fluid pressures (usually near-lithostatic; e.g. Sibson 1994).

3. Data analysis

3.1. Travel-time residuals and weighting

Outliers in the travel-time residuals can significantly affect an earthquake's estimated location if they are not downweighted, as demonstrated by Gomberg et al. (1990). This is because the location program attempts to fit all observed travel-times with the same weight equally well, so the effect of one outlier is distributed amongst the travel-time residuals of all other stations, shifting the hypocentre location away from its true position. Weighting of the travel-time residuals by the data analyst, as implemented in many location programs is important, but outliers are seldom recognised as such during the phase identification and weighting process. Anderson (1982) described the problem as follows: "... a good location cannot be determined until outliers have been identified, and outliers cannot be identified until a good location has been determined". Nevertheless, outliers can be identified and corrected for when all the travel-time residuals at one station are compared. Therefore, the analysis of the travel-time residuals at each station helps to identify mispicked phases, inaccuracies in the assumed velocity model and unstable hypocentre locations. Moreover, the residual analysis can be used to derive a reliable, objective weighting procedure.

3.1.1. Travel-time residual analysis

The travel-time residual analysis was first performed for events recorded by the SAMBA network between 24 March 2009 and 31 December 2009 (Fig. 3.1). This analysis was later repeated for earthquakes recorded by the DFDP10 array between 1 January 2010 and 31 April 2010.

Assuming that, on average, phases are correctly picked, the mean and the standard deviation of the travel-time residuals of all events at one station can be used to find outliers and trends in the dataset (Fig. G.1 (a) and (b) for the combined GeoNet and SAMBA stations and (c) and (d) for the DFDP10 network). Events with large residuals (≥ 2 standard

deviations) were double-checked.

Deviations of the mean of the travel-time residuals for one station from zero reflect inaccuracies in the applied 1-D velocity model. For example, the negative mean values for stations WHAT, COVA and WVZ (Fig. 3.2 a) show that calculated arrival times are on average greater than the observed arrival times, and thus that the velocity model is too slow. Since these three stations are located close to the Alpine Fault at either end of the combined SAMBA/GeoNet network, the seismic waves have the longest paths along the strike of the Alpine Fault. These negative residuals could, for example, indicate anisotropy with increased velocities parallel to the Alpine Fault. Mean positive residuals are observed for station POCR and ONE (Fig. 3.2 (a) and (b)), which are both situated in sediment-filled valleys. As the velocity model assumes bedrock at the surface, the velocity model is too fast in these cases. Assuming consolidated sediments at shallow depths with velocities of $V_P=2.2\text{--}2.7$ km/s (Davey 2010), the mean travel-time residual of 0.17 s for P-waves at POCR can be used to estimate the thickness of the sediment layer. For vertical incidence at the station, the estimated thickness of sediments $d=\Delta t(v_{\text{bedrock}}-v_{\text{sediments}})$ is 500–600 m beneath POCR. This range is slightly larger than the 320–450 m obtained in Section 1.4.3 from the resonance frequency. However, the lower range may reflect the fraction of less consolidated sediments from the total sediment thickness. An equivalent calculation for station ONE suggests sediment thicknesses of 2.1–2.5 km. For comparison, Davey (2010) obtained a 2.5–3 km-thick sediment cover in the coastal plain of the Whataroa valley from refraction/wide-angle reflection data. Onshore drillholes Harihari-1 (Sircombe and Kamp 1998) and Waiho-1 (Sutherland 1996) closer to the coast recorded 2.5 km and 3.6 km, respectively, of sediments above basement. The average basement depth northeast of the South Westland Fault Zone is about 3 km (Sircombe and Kamp 1998).

Assuming that the majority of the earthquakes are well located and few outliers bias the travel-time residuals, the standard deviation of the travel-time residuals at one station reflects the typical reading uncertainty. A comparison of the travel-time residuals for impulsive and emergent P- and S-waves (Table 3.2) at all stations shows that for the majority of stations

Figure 3.1. RMS residual of all events between 24 March 2009 and 31 December 2009 versus event number with its mean (blue) and standard deviation (red).

Figure 3.2. (a) Histograms of the residuals of SAMBA stations for the period November 2008 to December 2009. Note the different time-scales for the stations COSA and COVA. See Fig. 3.2 (b) for explanations of the symbols.

Figure 3.2. (b) Histograms of the residuals as in (a) for DFDP10 stations for the period January 2010 to April 2010. Note the different time-scale for station ONE which has a large delay. Blue and red crosses indicate the values of Jeffreys' weighting functions for P- and S-waves as calculated according to equation 2.13 for the first iteration described in the text, light blue and orange crosses are used accordingly for the second iteration.

the standard deviations for different phase onsets are similar. Consequently, impulsive and emergent phase arrivals can be treated as one population. However, the stations themselves do not have common variances. This means, Jeffreys' weighting function (Jeffreys 1973, equation 2.13) cannot be used on the collected residuals, since the weight for a particular residual is determined by the width of the initial residual distribution at each station. If two stations have the same percentage of outliers but not the same variances and Jeffreys' weighting is applied (equation 2.13), travel-time residuals at the station with the smaller variance contribute less to the location due to larger weights (although deviations from the mean value are smaller) than those at the other station exhibiting greater spread of the travel-time residuals. For the SAMBA stations, this was observed for the S-phase residuals at station WHYM ($\sigma_S=0.084$ s), which is located close to the majority of the events. The assigned weights to the residuals were systematically higher than at station MTFO ($\sigma_S=0.132$ s). This results in increased depth uncertainties, because the depth is particularly constrained by those stations near the hypocentre. For this reason, Jeffreys' weighting has to be modified to obtain a more consistent weighting scheme.

3.1.2. Weighting of the travel-time residuals

Jeffreys' intent in deriving the weighting scheme was to exclude those travel-time residuals in the tails of the weighting function (residuals that would contribute less than 5% in the location process), which is reasonable and can be consistently used for every station. To determine the number of outliers μ_0 and the standard deviation without outliers σ_0 required for Jeffreys' weighting, the following procedure is applied. First the mean and standard deviation of the travel-time residuals (including extreme values) and the number of outliers, defined as those values greater than two standard deviations of all travel-time residuals, are determined. Then the mean and standard deviation are recalculated excluding the outliers (1st iteration). These values are used to recalculate the number of outliers. In the second iteration the new number of outliers are removed from the travel-time residuals and the final values μ_0 and σ_0 are determined (mean and standard deviation of population without extreme values after two iterations; Table 3.1 (a) and (b)). These values are used to weight all residuals.

The new systematic weighting procedure implemented in the location process weights phases according to their deviation from the mean of all travel-time residuals and downweights outliers determined with Jeffreys' weighting. A weight of 75% was applied to residuals deviating from the mean travel-time residual by more than 0.1 s but less than 0.2 s. Half the weight (50%) was assigned to residuals differing from the mean by ≥ 0.2 s; 25% to residuals ≥ 0.4 s,

Station	Number of phases	Number of outliers in		Mean [s] (initial)	Standard deviation [s] (initial)	Mean [s] (1st iter.)	Standard deviation [s] (1st iter.)	Mean [s] (2nd iter.)	Standard deviation [s] (2nd iter.)
FRAN P	729	50	93	0.033	0.091	0.034	0.062	0.032	0.052
FRAN S	707	51	95	-0.013	0.152	0.003	0.114	0.016	0.096
REYN P	563	31	61	0.049	0.126	0.035	0.100	0.026	0.089
REYN S	498	22	51	0.027	0.173	0.017	0.119	0.010	0.101
WHAT P	685	17	60	0.021	0.144	0.018	0.119	0.013	0.104
WHAT S	618	12	27	-0.085	0.177	-0.089	0.144	-0.091	0.137
LABE P	655	51	99	-0.042	0.108	-0.031	0.070	-0.028	0.054
LABE S	683	31	79	-0.095	0.211	-0.067	0.125	-0.045	0.091
MTFO P	610	29	72	-0.024	0.127	-0.014	0.101	-0.008	0.085
MTFO S	622	50	98	-0.069	0.285	-0.040	0.178	-0.016	0.132
EORO P	780	51	103	-0.045	0.092	-0.035	0.066	-0.032	0.056
EORO S	675	53	100	-0.156	0.169	-0.129	0.110	-0.113	0.079
COSA P	346	18	38	0.038	0.112	0.036	0.084	0.036	0.073
COSA S	365	20	37	0.095	0.225	0.088	0.162	0.090	0.144
POCR P	592	36	65	0.158	0.092	0.163	0.072	0.165	0.065
POCR S	488	20	44	0.133	0.160	0.129	0.121	0.129	0.108
WHYM P	840	64	113	-0.015	0.096	-0.017	0.074	-0.018	0.064
WHYM S	803	39	93	0.010	0.166	0.024	0.105	0.034	0.084
GOVA P	777	58	125	-0.027	0.108	-0.032	0.068	-0.031	0.053
GOVA S	846	37	100	0.014	0.137	0.024	0.085	0.030	0.066
FOZ P	273	14	29	0.021	0.206	0.006	0.106	0.001	0.079
FOZ S	348	21	43	-0.031	0.201	-0.048	0.138	-0.057	0.115
RPZ P	227	9	24	-0.039	0.157	-0.052	0.101	-0.055	0.081
RPZ S	284	10	31	0.083	0.140	0.074	0.101	0.072	0.084
WVZ P	273	24	49	-0.059	0.160	-0.050	0.107	-0.048	0.084
WVZ S	265	19	35	-0.087	0.198	-0.068	0.144	-0.051	0.119

Table 3.1. (a) Pre-weighting number of phases, number of outliers, mean and standard deviations of the travel-time residuals for all P- and S-phase arrival-times recorded by the SAMBA stations between November 2008 and December 2009.

Station	Number of phases	Number of outliers in 1st/2nd iter.	Mean [s] (initial)	Standard deviation [s] (initial)	Mean [s] (1st iter.)	Standard deviation [s] (1st iter.)	Mean [s] (2nd iter.)	Standard deviation [s] (2nd iter.)
FRAN P	143	6	-0.027	0.107	-0.027	0.094	-0.024	0.090
FRAN S	117	8	-0.033	0.176	-0.041	0.147	-0.039	0.140
REYN P	46	2	0.051	0.112	0.048	0.086	0.044	0.082
REYN S	40	3	0.070	0.132	0.062	0.104	0.068	0.087
WHAT P	244	8	-0.007	0.145	0.012	0.095	0.020	0.080
WHAT S	201	10	-0.080	0.157	-0.093	0.130	-0.103	0.120
LABE P	243	13	-0.150	0.188	-0.138	0.166	-0.135	0.152
LABE S	195	16	-0.178	0.215	-0.160	0.177	-0.147	0.156
MTFO P	188	11	-0.064	0.154	-0.048	0.133	-0.031	0.116
MTFO S	140	8	-0.073	0.186	-0.047	0.151	-0.032	0.128
EORO P	22	0	-0.060	0.077	-0.060	0.077	-0.060	0.077
EORO S	14	0	-0.089	0.074	-0.089	0.074	-0.089	0.074
COSA P	149	7	0.083	0.120	0.077	0.089	0.077	0.076
COSA S	128	6	0.111	0.137	0.118	0.116	0.124	0.107
POCR P	161	12	0.138	0.104	0.150	0.082	0.158	0.069
POCR S	107	8	0.116	0.158	0.116	0.132	0.121	0.118
WHYM P	214	12	-0.040	0.124	-0.027	0.106	-0.017	0.096
WHYM S	150	6	-0.026	0.151	-0.014	0.130	-0.011	0.120
GOVA P	279	18	-0.068	0.137	-0.063	0.115	-0.054	0.099
GOVA S	268	15	-0.034	0.161	-0.012	0.132	-0.005	0.121
COVA P	60	4	-0.116	0.163	-0.107	0.126	-0.078	0.093
COVA S	43	2	-0.296	0.239	-0.297	0.214	-0.308	0.204
FOZ P	108	11	0.029	0.159	0.031	0.108	0.021	0.090
FOZ S	116	3	-0.032	0.191	-0.036	0.181	-0.042	0.170
RPZ P	87	5	0.022	0.155	0.006	0.115	-0.005	0.091
RPZ S	136	8	0.172	0.147	0.167	0.121	0.163	0.109
WVZ P	191	7	-0.136	0.147	-0.143	0.130	-0.145	0.121
WVZ S	154	5	-0.223	0.177	-0.213	0.155	-0.202	0.145

Table 3.1. (b) Pre-weighting number of phases, number of outliers, mean and standard deviations of the travel-time residuals for all P- and S-phase arrival-times recorded by the combined SAMBA and DFDPI0 stations between January 2010 and April 2010.

and 0% for ≥ 0.8 s (summarised in Table 3.3).

Station	$\mu_{IP} \pm \sigma_{IP}$ [s]	$\mu_{EP} \pm \sigma_{EP}$ [s]	$\mu_{IS} \pm \sigma_{IS}$ [s]	$\mu_{ES} \pm \sigma_{ES}$ [s]	IP	EP	IS	ES
FRAN	0.03 ± 0.06	0.04 ± 0.09	0.02 ± 0.11	-0.01 ± 0.15	421	401	315	471
REYN	0.11 ± 0.08	0.04 ± 0.11	-0.00 ± 0.09	0.01 ± 0.11	8	610	88	446
WHAT	0.02 ± 0.13	0.02 ± 0.15	-0.10 ± 0.18	-0.07 ± 0.18	419	330	185	449
LABE	-0.03 ± 0.07	-0.04 ± 0.14	-0.06 ± 0.16	-0.06 ± 0.13	522	232	460	278
MTFO	-0.02 ± 0.10	-0.00 ± 0.11	-0.01 ± 0.16	-0.04 ± 0.19	322	357	177	507
EORO	-0.03 ± 0.07	-0.04 ± 0.08	-0.10 ± 0.08	-0.13 ± 0.12	406	478	245	479
COSA	0.03 ± 0.12	0.04 ± 0.09	0.16 ± 0.15	0.13 ± 0.93	75	304	67	331
POCR	0.16 ± 0.09	0.16 ± 0.09	0.14 ± 0.25	0.12 ± 0.14	345	313	71	444
WHYM	-0.01 ± 0.08	-0.00 ± 0.11	0.01 ± 0.10	0.02 ± 0.12	671	288	358	529
GOVA	-0.03 ± 0.08	-0.00 ± 0.10	0.03 ± 0.08	0.01 ± 0.31	699	195	614	333
FOZ	0.01 ± 0.10	0.01 ± 0.10	-0.02 ± 0.18	-0.04 ± 0.16	96	200	108	267
RPZ	-0.06 ± 0.09	-0.04 ± 0.11	0.04 ± 0.09	0.05 ± 0.10	130	117	156	152
WVZ	-0.08 ± 0.14	-0.08 ± 0.14	-0.05 ± 0.12	-0.13 ± 0.22	236	84	196	88

Table 3.2. Post-weighting mean and standard deviations of the travel-time residuals for impulsive and emergent P- and S-phase arrival-times for all stations. The mean μ represents the model-error σ^{model} of the 1-D velocity model for one station. The standard deviation σ indicates the mean reading error σ^{reading} . The numbers give total numbers of P- and S-phases recorded at the station.

Weight	Contribution	Absolute deviation from the mean [s]
0	100%	< 0.1
1	75%	≥ 0.1
2	50%	≥ 0.2
3	25%	≥ 0.4
4	0%	≥ 0.8

Table 3.3. Weight assigned to the observed phase arrival-time according to its deviation from the mean phase arrival time at each station.

Figure 3.3 shows the final weights applied to P- and S-phases for all events between November 2008 and April 2010. Few phases are omitted entirely (weight=4) at each station, and the majority of phases have a full weight (weight=0). P-picks seldom have a weight larger than 2, however some S-phases (especially at stations MTFO, REYN and POCR) were assigned with only half or less than the full weight.

Figure 3.3. (a) Overview of weights assigned to P (blue) and S (red) phases for the SAMBA stations with the number of outliers listed in Table 3.1 (a) and (b)

Figure 3.3. (b) Overview of weights assigned to P (blue) and S (red) phases for the DFDP10 stations applying the weighting scheme (Table 3.3). Note that the DFDP10 stations operated only for the last few months of the time analysed.

3.1.3. Time corrections of residuals for station COSA

During a 77-day period between 25 June and 10 September 2009, station COSA near Cook's Saddle lost satellite contact, resulting in GPS clock uncertainties. Several metres of gravel (from a landslide nearby) had been dumped on top of the sensor and buried the GPS antenna approximately 20 cm deep. During this interval, no signal from the GPS satellites was received, resulting in missing time stamps from the external clock and the internal data logger clock began to drift. The uncorrected travel-time residuals are shown in Fig. 3.4 (a).

The uncorrected travel-time residuals show a systematic pattern. The total accumulated time difference between the internal Reftek clock and the external GPS clock for the 77-day period was 26 s. A linear drift correction (blue crosses in Fig. 3.4 b) does not fully account for the differences as shown in (c): the linear correction introduces systematic bias to events within one group with the same clock differences (e.g. events 776 to 825 in (a) and (c)). Applying the same correction value to selected groups of events results in a step function as shown by the black line in (d). The same correction was applied to the S-phase (green line in d) when no P-phase could be identified for the same event. If a P-phase exists, the location program HYPOCENTER (Lienert et al. 1986) used within SEISAN automatically corrects the S-phase. These corrections produce a normal distribution of the reduced travel-time residuals as shown in Fig. 3.4 (e) (cf. Fig. 3.2), which can be treated with the weighting scheme as described in Section 3.1.2. This observation implies that the clock drift of the Reftek data loggers (at least the 915x-series units used in this study) occurs in discrete pulses rather than continuously as assumed when using a linear drift correction. These pulses may be related to large temperature changes, but more observations would be needed to clarify the cause.

3.1.4. Time corrections of residuals for station DRC

DFDP10 station DRC lost satellite contact between Julian days 042 and 117 of 2010 (14 March–27 April). However, no apparent drift can be detected from the distribution of the residuals before the station service on day 074. During the servicing, the GPS problem was noted but could not be fixed. After Julian day 074, a constant time delay of 24.878 s was determined which persisted until day 117 on which satellite contact was reestablished and a time difference of 24 s between internal and external GPS clock was recorded by the Reftek.

Figure 3.4. (a) Uncorrected travel-time residuals for station COSA near Cook's Saddle. Black dots indicate a weight assigned to the absolute arrival time, (b) travel-time residuals of P arrivals (dark blue) that are fitted with a linear drift correction (light blue) between the known beginning and the end time of the GPS clock uncertainties, (c) corrected travel-time residuals using the linear drift correction, (d) travel-time residuals of P (blue) and S arrivals (red) fitted with step-functions for P (black line) and S-residuals (green line) (e) corrected travel-time residuals using the step-functions in (d)

3.1.5. Station terms from the residuals

Figure 3.5 shows the average travel-time residuals (also called the station terms) of all stations that result from using the velocity model in Table 1.2. The majority of the station terms on the western side of the Alpine Fault and those on the eastern side close to the fault show positive average travel-time residuals. Those stations distributed within the “High” Alps have negative station terms on average. This systematic difference reflects lower velocities of the sediment-filled plains west of and within the damage zone of the Alpine Fault and higher velocities for the bedrock with little sediment cover in the Southern Alps. A tomographic inversion to resolve these differences in the velocity model is performed in Section 6.3.

Figure 3.5. Station residuals for P- (circle) and S-waves (triangles) as determined using NonLinLoc for all events recorded between November 2008 and April 2010.

3.2. A local magnitude scale for the Southern Alps

3.2.1. Geometrical spreading and attenuation parameter

In routine, New Zealand-wide magnitude determination performed by GeoNet, an attenuation coefficient of $\gamma=6.7\times10^{-3}\text{ km}^{-1}$ is used. This standard value was derived by Robinson (1987) from 1552 local events between 1978 and 1985 of magnitude 1.6–5.2, and depths of 0.5–90.0 km that occurred in the Wellington region. There, the subduction zone interface lies at 20–25 km depth (Ansell and Bannister 1996), so most local earthquakes occur in the crust and the subducting slab. Seismic waves of the deep earthquakes cross the crust vertically and are therefore less attenuated than waves which propagate laterally in the crust. Station WEL in Wellington, which has been equipped with a Wood-Anderson seismometer since January 1931 serves as a reference station (Haines 1981). Station correction terms for the GeoNet stations are adjusted in such a way that the reference seismograph in Wellington would have a zero station correction term¹.

If the standard attenuation term is used for magnitude calculations from the SAMBA data (according to equation 2.23) the majority of the magnitude residuals are positive for hypocentral distances larger than 80 km as shown in Fig. 3.6. For this reason, a new local magnitude scale for the central Southern Alps region has had to be derived. The magnitudes determined in this study are calibrated using magnitudes obtained by GeoNet (e.g. Ristau 2009). Magnitude estimates are available from GeoNet for a total of 147 earthquakes ($1.6\leq M_L\leq 4.2$), enabling the new required magnitudes to be related and compared to the existing magnitude scale. Figure 3.7 shows all amplitude readings at the SAMBA stations for the period November 2008 to December 2009 and the GeoNet magnitudes for events recorded by both networks. Amplitude-readings are obtained from half the peak-to-peak displacement on one of the horizontal components of a pseudo-Wood-Anderson seismograph, simulated in SEISAN (version 8.2.1, October 2008, and 8.3, May 2010) (Havskov and Ottemöller 2000). It is assumed that the gain of the Wood-Anderson instrument is 2080 ± 60 , as reported by Uhrhammer and Collins (1990).

All amplitudes of the SAMBA stations are first inverted for the geometrical spreading coefficient α (equation 2.16), ignoring the effects of attenuation (Fig. 3.8, top). The obtained value of $\alpha=1.7$ indicates that normal body-wave expansion proportional to r^{-1} prevails, but that attenuation has a non-negligible effect on the amplitude decay. In the

¹Information obtained from <http://www.geonet.org.nz/resources/earthquake/hypocentre-derivation.html>, last accessed February 2012

Figure 3.6. Magnitude residuals versus hypocentral distance for all earthquakes recorded by the SAMBA stations between November 2008 and December 2009. The same attenuation parameter $\gamma=6.7 \times 10^{-3} \text{ km}^{-1}$ as used by GeoNet in routine magnitude determination has been applied.

Figure 3.7. The logarithm of all available amplitude readings (blue) for earthquakes between November 2008 and December 2009 and calibration magnitudes (red) of events which were recorded by several GeoNet stations in relation to hypocentral distance. In order to derive the correct magnitude of the earthquakes, the amplitude decay with distance from the source due to geometrical spreading and attenuation has to be determined.

Figure 3.8. Distance correction term (left) and magnitude residuals (right) versus hypocentral distance for different inversion procedures as described in the text.

following, $\alpha=1$ is assumed and all inversions are performed for the anelastic attenuation parameter.

Inversion of the amplitude data for a frequency-independent anelastic attenuation coefficient yields $\gamma=0.0148 \text{ km}^{-1}$. This value is twice the standard attenuation coefficient used by GeoNet, but smaller than the value obtained by O' Keefe (2008). However, the result is unstable when amplitude data with different maximum hypocentral distances are inverted, in which case an increase of γ with decreasing hypocentral distance is observed. Simultaneous inversion of the amplitude data for two distance ranges, separated by the crossover distance Δ_c , indicates high γ -values for $\Delta < \Delta_c$ and low values for $\Delta \geq \Delta_c$ (Fig. 3.8, middle). Although this inversion fits the amplitude data best, the two attenuation coefficients cause an unphysical offset of approximately half a magnitude unit in the attenuation correction curve at Δ_c . Similarly, a simultaneous inversion of the amplitude data separated for the SAMBA and GeoNet stations, indicates a low attenuation parameter for the GeoNet and a high attenuation parameter for the SAMBA stations (which are on average closer to the hypocentres in question).

In a different approach, the attenuation coefficients in an upper surface layer γ_u and a halfspace below γ_l is inverted for simultaneously. This enables testing whether a highly attenuative surface layer exists, which has a proportionally larger effect on small station–event distances than on large hypocentral source–receiver distances. The inversion is performed for surface layer thicknesses of 1–8 km (and is shown for a thickness of 2 km in Fig. 3.8), revealing plausible values for all layer thicknesses and no particular improvement of the magnitude residual misfit at larger distances.

Those different inversion trials all produce predominantly positive magnitude residuals for hypocentral distances $\Delta \geq 60 \text{ km}$, indicating that the applied attenuation curve overestimates the actual attenuation. Magnitude residuals which are normally distributed about zero can only be obtained if an additional correction term proportional to Δ^2 is added in the inversion for frequency-independent γ .

However, when the same amplitude data are inverted for a frequency-dependent attenuation factor $\gamma=\gamma_0 f$ the data can be fit for all hypocentral distances. This inversion is performed because the period of the wave corresponding to the maximum amplitude value shows a clear increase with hypocentral distance (Fig. 3.9, top). The resulting $\gamma_0=1.89 \times 10^{-3} \text{ s/km}$ explains the decrease of the attenuation function with distance. Earthquake waves from larger distances are depleted of their high-frequency content ($f>10 \text{ Hz}$)

and are therefore less attenuated. High-frequency waves from nearby earthquakes, on the contrary, are highly attenuated. For the hypocentral distances considered in this study, which peak at 20 km and few amplitude readings for distances ≥ 70 km, many more microearthquakes at short distances are inverted for attenuation.

Magnitude inversion for the period November 2008 to December 2009 for the SAMBA stations results in $\gamma_0 = 1.89 \pm 0.02 \times 10^{-3}$ s/km (Section 4.2.2). When the same inversion method was applied to the data from the combined SAMBA and DFDP10 arrays for the period January to April 2010, the same frequency dependence of the attenuation coefficient was observed with a slightly larger value of $\gamma_0 = 2.37 \pm 0.01 \times 10^{-3}$ s/km.

Given the hypocentral ranges encountered in this study, the magnitude correction term corresponding to geometrical spreading $\alpha \log(\Delta)$ varies between 0 and 2 magnitude units. The attenuation term $0.4343 \gamma \Delta$, in contrast, ranges between 0 and 1 for frequencies of 10 Hz, and 0 and 0.55 magnitude units for wave frequencies of 5 Hz. In other words, the influence of attenuation can amount to half that of the geometrical spreading.

3.2.2. Magnitude station correction terms

Magnitude station correction terms reflect a systematic deviation of the magnitude at one station from the magnitude of an earthquake due to unmodelled subsurface structure. Stations above thick sediment layers exhibit larger magnitudes than stations on basement rocks due to amplification. The station correction terms obtained in this study overall reflect the depth at which the sensor is deployed and the site geology. The two deepest borehole stations in bedrock FRAN and WHAT have the lowest correction terms, followed by the third borehole station in sediments POCH and the GeoNet station RPZ, which is situated in a 50 m deep borehole in rhyolite (Petersen et al. 2011). The remaining GeoNet stations (FOZ, WVZ) and the SAMBA stations GOVA show lower than average values, indicating bedrock at these sites. Similarly, the lowest station correction terms from the DFDP10 network were found for station BLO, BON and GHU, where field observations suggest these sites are close to bedrock.

The station correction terms determined here for the GeoNet stations (FOZ, WVZ and RPZ) do not agree with values (0.34–0.52) used by GeoNet for these stations (K. Fe-naughty, pers. comm., 2009). This may result from the distribution of earthquakes relative to the stations in this study versus the distribution of all earthquakes in New Zealand relative to these stations as used by GeoNet. However, since the magnitudes obtained

Figure 3.9. Period of the maximum Wood-Anderson displacement amplitude (top), attenuation functions (centre) and magnitude residuals (bottom) versus hypocentral distance for the period January to April 2010 for amplitude readings from the combined SAMBA and DFDP10 arrays. The same plot for the period November 2008 to December 2009, when only the SAMBA network was operational is shown in Fig. 4.5

in this study are not significantly different from the magnitudes determined by GeoNet, whereas the attenuation coefficients are, this has to have an effect on the station correction terms (equation 2.23). Any systematic differences in the magnitudes will also be reflected in the station correction terms. The correction term attributed to attenuation in this study will be larger (on average) due to the frequency-dependent attenuation coefficient and the high-frequency content of nearby earthquakes than the same term used by GeoNet, and thus the station term by which the whole magnitude is reduced will also be larger in this study. This theoretically explains why the station terms derived here are one magnitude unit larger than those used by GeoNet.

A discussion about the errors of the magnitudes determined in this study due to the calibration described above is given in Section 4.2.2. One observation, however, that has not been mentioned is that low frequencies are amplified on simulated Wood-Anderson seismograms as a result of filtering. As a result, the displacement for a wave with lower frequency than that wave with the maximum displacement on the unfiltered trace appears larger than the actual maximum displacement on this trace. However, these large displacements of the low-frequency waves are not restricted to the earthquake signal. Therefore, peak amplitudes were only picked if the maximum amplitude corresponded to that on the unfiltered trace.

A comparison of the magnitudes of the events listed in the GeoNet and the SAMBA catalogue is shown in Figures 3.11 and 3.12. This shows that few earthquakes within the study area and the majority of earthquakes at the boundaries of the array are used for magnitude calibration. More discussion on the comparison between locations of the same events will be given in Section 3.3.1. The difference between the magnitude derived in this study and by GeoNet is small but significant. For events of $M_L \geq 3$, the resulting SAMBA magnitudes exhibit lower values than the GeoNet magnitudes (c.f discussion of magnitude uncertainties in Section 4.2.2).

Figure 3.10. Station correction terms in magnitude units for the combined SAMBA and DFDP10 networks. Symbols are the same for stations belonging to one network. The bars that link the events indicate the depth differences between hypocentre locations obtained in this study and by GeoNet. Yellow bars show values greater than 10 km, whereas green bars mark smaller hypocentral differences.

Figure 3.11. SAMBA magnitudes versus GeoNet magnitude for events listed in both catalogues. The black dotted line indicates equal values. The black solid line represents the best fitting regression line and its 95% confidence interval (green lines). The red lines reflect the spread in the magnitude values within the 95% interval.

Figure 3.12. Epicentres of events listed in the GeoNet and the SAMBA catalogue and used for magnitude calibration in this study

3.3. Hypocentre comparisons and stability

3.3.1. Events from the GeoNet catalogue

A comparison of the hypocentres of 245 events recorded by both the GeoNet network and the SAMBA array is shown in Fig. 3.12. The mean difference in latitude is $0.01 \pm 0.06^\circ$ ($\pm$ standard deviation of the data), $0.00 \pm 0.08^\circ$ in longitude, 2 ± 6.4 km in depth and in RMS residual is 0.02 ± 0.16 s. The bars that link events in Fig. 3.12 indicate the depth differences between hypocentre locations obtained in this study and by GeoNet. Yellow bars show values greater than 10 km, whereas green bars mark smaller hypocentral differences. Most events with large differences shifted towards the centre of the SAMBA array, reflecting the slightly faster ($+0.17$ km/s) P-wave velocity in the uppermost 8 km used in this study (Table 1.2) in comparison to the standard New Zealand velocity model used by GeoNet (Table 3.4).

Depth [km]	V _P [km/s]	V _S -Velocity [km/s]
0.0	5.5	3.3
12.0	6.5	3.7
33.0	8.1	4.6

Table 3.4. Standard 1-D velocity model used by GeoNet for hypocentre location New Zealand-wide.

3.3.2. Quarry blasts

A total of seven quarry blasts from a quarry in Whataroa were recorded by the SAMBA stations and the combined SAMBA/DFDP10 arrays (Table 3.5). These quarry blasts are characterised by impulsive P- and weak S-waves (Fig. 3.13) (e.g. Allmann et al. 2008, and references therein). Similar amounts of energy were released as by M_L 0.7–1.9 earthquakes. The quarry is located at $-43.283^\circ/170.361^\circ$, approximately 0.4 km southeast of station WHAT. Unfortunately, the operating company would not provide origin times for these shots. Nevertheless, the events could be used to test the location and the polarity of the stations (Appendix B). When the depth was initially unconstrained, these events located at 5–8 km depth at epicentral coordinates matching the quarry location well. However, in this case the residuals for the closest station WHAT were high (>0.68 s too early). Fixing the depth to the surface resulted in a shift of the epicentre by 0.005 – 0.02° (0.4–1.6 km) north. The residuals for WHAT diminished to absolute values of 0.2–0.4 s but still indicate that the

velocity model is too slow.

Husen et al. (1999) observed the same phenomenon for seven mine blasts recorded by a 35-station network in the vicinity of Antofagasta, Chile. They interpreted the consistently large offset in depth as the result of an inadequate approximation of the local near-surface velocities in the vicinity of the mine. An inversion for the velocities in the uppermost two layers shifted all events to 1–2 km depth, which reflects their absolute location error. Only few quarry blasts in Whataroa were recorded by the majority of stations (13 combined SAMBA/GeoNet network and additional 12 DFDP10 stations in 2010), so we do not invert for the velocities in the uppermost layers separately as Husen et al. (1999) did but instead these blasts are considered in the tomography described in Section 6.3.

Date	Time	Error [s]	Lat. [°]	Error [km]	Lon. [°]	Error [km]	Stn	RMS [s]	Gap [°]
22.06.2009	04:06:06	0.51	−43.279	1.3	170.361	1.6	11	0.2	177
04.01.2010	01:08:25	0.35	−43.245	1.5	170.373	1.5	10	0.1	298
12.01.2010	03:37:06	0.40	−43.251	2.5	170.361	2.1	6	0.2	297
22.01.2010	01:29:28	0.42	−43.280	1.9	170.364	1.4	10	0.2	123
27.01.2010	03:16:02	0.55	−43.276	1.3	170.357	1.3	13	0.2	155
01.02.2010	04:24:00	0.48	−43.273	1.5	170.355	1.4	13	0.2	184
23.02.2010	22:25:08	0.40	−43.282	1.0	170.355	1.4	14	0.2	154

Table 3.5. Quarry blasts in Whataroa recorded by the SAMBA network or SAMBA and DFDP10 arrays (with the number of stations Stn that recorded events) that were located using a fixed depth of 0 km.

3.4. Station configuration

3.4.1. Earthquakes outside the station network

To test the influence of the station network on the calculated hypocentres, we locate events that occurred outside the SAMBA array but inside the DFDP10 array using travel-time observations from the combined station networks and from the SAMBA stations only. Those earthquakes that were recorded by more than 20 stations between January and April 2010 are shown in Figure 3.14. The PDF hypocentre solutions shown in red were calculated from the travel-time observations at the SAMBA stations only. Black lines indicate the shift from the original hypocentres (using all stations for locating the event) to the relocated hypocentres (when only the SAMBA stations are considered). The epicentral coordinates are stable and

Figure 3.13. Distance versus reduced travel-time for the quarry shot (equivalent M_L 1.5) of 23 February 2010 recorded by the SAMBA and DFDP10 stations. The travel-time was reduced by the time resulting from the velocity of the uppermost layer (5.67 km/s) over the distance. The distance (in km) is given by the number after the station name and component shown. Note that the delay at ONE is likely due to slow velocities associated with the sediments (cf. large station term in Fig. 3.5). WVZ and FOZ are the most distant stations for waves propagation parallel to the Alpine Fault and are both located on the Australian plate.

horizontal changes are less than 1 km which is within the uncertainty of the hypocentres. However, this stability is due to the high number of stations that recorded the earthquake (Section 3.4.2). The depth, in contrast, changes markedly because there is no station within a distance of 1.5 times the focal depth. The hypocentres are moved to the surface (positive vertical shifts; Fig. 3.14 b). In most cases, however, the original location is contained in the uncertainty volume.

3.4.2. Number of recording stations

For a few events recorded by all SAMBA and the three surrounding GeoNet stations, the stability of the hypocentres was tested. The most distant stations to the centre of the SAMBA array were systematically taken out during the location process. This was repeated in six steps, starting with thirteen and ending with three stations (Fig. 3.15). The uncertainty in each hypocentre directly reflects the number of stations that recorded the earthquake and the station distribution. Its volume decreases markedly with increasing number of stations that recorded the event. Despite a poor azimuthal coverage of the recording stations relative to the event, the number of stations that record the event is significant (e.g. 10 versus 7 stations) in improving the hypocentre location.

Overall, the epicentral coordinates of the maximum likelihood hypocentre (latitude and longitude in Fig. 3.16) change little during the location steps with different numbers of recording stations. The root mean squared residual σ_{RMS} depends on the number of stations at which the earthquake is recorded. As expected, the σ_{RMS} value increases with the number of recording stations (Havskov and Ottemöller 2010). Fluctuations in depth are most pronounced and are on the order of several kilometres. Note that one particular phase arrival time can disturb the hypocentre solution (e.g. 5 versus 4 recording stations) as illustrated by Fig. 3.16. When the travel-time residuals become incompatible, the event is moved to the surface (depth=-2 km due to topography). This is usually accompanied by a change in the RMS value.

3.4.3. Hypocentre locations before and after weighting

A comparison of the hypocentres determined from the consistently weighted phases with those resulting from the original manually weighted phases reveals only minor differences for the majority of events but significant shifts for some individual events. All hypocentre locations that changed by more than 10 km either vertically or horizontally were re-checked

Figure 3.14. Hypocentre locations (maximum likelihood hypocentres, black dots) and their uncertainty represented by the probability density function (red) of each earthquake in map view (rotated by 55° so that the Alpine Fault is shown by the solid black almost vertical line) and on a profile parallel to the X-axis below. Short black lines represent shifting of the hypocentres from their original position (including DFDP10 station) to the new locations (excluding DFDP10) in order to test the stability of the hypocentres outside the SAMBA network. Horizontal and vertical fluctuation of the hypocentre coordinates are shown in the histograms.

Figure 3.15. Probability density functions of the hypocentre solution for different numbers of stations that recorded the earthquake. In total 44 events were recorded by all SAMBA and GeoNet stations. For details about the plots see Fig. 3.14.

Figure 3.16. Fluctuation of hypocentre coordinates and RMS values due to the number of travel-time observations used for locating the earthquakes. Which stations were considered in the different steps is shown in Fig. 3.15. The different coloured lines in each panel correspond to individual earthquakes.

manually for inconsistent travel-time residuals. Usually, in such cases, one mispicked phase could be identified that was downweighted when applying the weighting scheme. On average, the final epicentres shifted less than the mean horizontal uncertainty of 1 ± 1.3 km (as shown in Fig. 3.17 and 3.18). In depth, the majority of the events shifted by less than ± 1.8 km. The tails of the distribution of vertical shifts indicate extreme changes of as much as 10 km upwards (positive values, usually events with inconsistent picks that were moved to the surface) and downwards (negative depth shifts, where the weighting eliminated inconsistencies in the phase picks). These occurred predominantly at the boundaries of the array. All other vertical movements lie within the error volume of the hypocentre uncertainties (3 ± 2.9 km). Changes in RMS-values show the improvement of the average travel-time residual due to weighting (as shown in Fig. 3.18, compare also Fig. 4.4). Our weighting procedure effectively removes the influence of the station term from the weight, so that stations with larger station terms, which were usually downweighted manually, get comparable weights to those stations with small station terms.

3.4.4. Hypocentre relocations

A total of 1993 events in the catalogue of absolute hypocentre locations were used for relocation with hypoDD Waldhauser and Ellsworth (2000). These events were filtered between 3–15 Hz and cross-correlated when the earthquake was recorded by the same station. A cross-correlation threshold of 0.5 for at least 5 station pairs had to be exceeded for the event to be listed in the cross-correlation catalogue. This resulted in a catalogue of 8137 pairs of earthquakes. Strong links (>8) between the events in the two catalogues are required for relocation, so that the events' spatial positions can be determined relative to those nearby events. Earthquake relocations are determined in an iterative process, starting with a full weight on the catalogue double-difference values. In five iterations, the catalogue differential-travel-times were systematically downweighted and the weights of the cross-correlation differential-travel-times were increased. The specifications for the iterations in the relocation process are shown in Table 3.6.

Initially, only catalogue P-wave differential travel-times are used in the relocation process so that the initial position of the events is determined by the absolute hypocentre location. Next, catalogue S-waves differential travel-times with half the weight of the P-wave differential travel-times are included. In the third iteration step, the cross-correlation differential travel-times are added and the maximum separation distances and the standard deviation of the catalogue differential travel-times are lowered. The maximum separation distance between absolute hypocentre locations is set here to 3 km (Maurer and Deichmann

Figure 3.17. Hypocentre locations after the weighting scheme was applied (blue stars and uncertainty in red) and their changes (black lines) from the previous location without systematic weighting. Shown are all events between November 2008 and April 2010 on map view and projected on cross sections along the X- and Y axis through the origin of the coordinate system.

Figure 3.18. Horizontal and vertical location shifts between the hypocentres obtained without and with the weighting scheme. Note that the vertical scale extends to counts of 1500 for all three plots and has been truncated to make the tails of each histogram visible.

1995) which corresponds to approximately twice the average location error of the events (average horizontal uncertainty of 1 ± 1.3 km), and two standard deviations as the residual threshold to be consistent with findings in Section 3.1.1.

Iteration	Cross-correlation dt weights		Res. thres. in σ	Max. event sep.	Catalogue dt weights		Res. thres. hold in σ	Max. event sep.	Damping
	P	S			P	S			
1	0.001	—	—	—	1	0.001	—	—	95
2	0.01	—	—	—	1	0.5	7	6	95
3	0.1	0.001	7	7	1	0.5	3	3	90
4	1	0.001	5	5	0.5	0.01	2	3	90
5	1	0.5	3	3	0.01	-9	2	3	60
6	1	0.5	2	0.5	0.001	-9	2	3	60

Table 3.6. Input parameters for hypoDD relocations. The residual threshold (column 3 for cross-correlation and 6 for catalogue data, respectively) is expressed in terms of number of standard deviations σ , so that a value of 3 means all residuals within 3σ of the mean residual are included. Note that the maximum event separation (column 4 for cross-correlation and 7 for catalogue data, respectively) is in kilometres.

In the following iteration, weights for the cross-correlation differential travel-times are increased and the S-phase differential travel-times from the cross-correlation catalogue are added. In the fifth and six iteration steps, the maximum separation distance and the standard deviation of the cross-correlation differential travel-times are systematically reduced. The catalogue differential travel-times are downweighted to constrain the spatial position as a relative location between earthquakes. The final threshold for the maximum separation distance between relative hypocentre locations is 500 m, significantly smaller than the overall absolute location error.

During the relocation process, the weighted square sum of these double-difference residuals for pairs of earthquakes at each station is minimized. This can be accomplished using either the singular value decomposition (SVD) or the conjugate gradients method (LSQR). SVD is applied to look at the fine-scale structure of several clusters of events otherwise LSQR is used.

In total 936 events were relocated and approximately 20 earthquakes were removed as “airquakes” (defined as events with depths above sealevel in hypoDD so topography is not taken into consideration). Conditioning numbers for the largest cluster of events start at high values of 140 but improve significantly once the airquakes are removed (<100). The final two iterations have stable values between 60 and 70. For fewer than 11 links between

events, the majority of earthquakes are combined to one large cluster. The larger the number of links, the more tightly clustered are the events and the smaller the overall number of relocated events per cluster. In principle, however, the clusters remain the same but get split into several smaller ones. The overall number of events that are relocated drops by one-third to 664.

Figure 3.19 shows the relocated hypocentres for eight links. Also shown are the differences between the absolute catalogue locations and the relocated hypocentres (black bars). Individual clusters with more than eight events are shown in colour. Notably those events northeast of the SAMBA array which could be correlated with events that occurred in early 2010 within the DFDP10 array (when the depth was better constrained) shifted downwards by more than a few kilometres. Unfortunately, however, the hypocentre relocations do not reveal linear or planar fault structures in more detail. Nevertheless, they do show that earthquakes occur in a defined depth range and are not as scattered throughout the seismogenic zone as apparent in the absolute event locations (cf. Fig. 3.17). Several of the clusters identified in the relocation coincide with swarms identified in Chapter 5.

3.5. Focal mechanism comparisons

To test our focal mechanism program, we first tried to reproduce focal mechanism solutions obtained by Leitner et al. (2001) for the Southern Alps region. In order to get take-off angles for the events studied by Leitner et al. (2001), we use the station coordinates of the GeoNet and SAPSE networks, the 1-D velocity model applied by Leitner et al. (2001) and the travel-time observations to relocate the events with NonLinLoc. A total of 18 mechanisms for earthquakes that occurred between November 1995 and April 1996 could be compared (Fig. 3.20).

For the majority of events, we obtain a similar focal mechanism but our results differ in detail from those solutions by Leitner et al. (2001). The main reason for these differences is that the relocated hypocentres differ (especially in depth) from the original locations, resulting in different distributions of P-wave polarities on the focal sphere. Some of the original travel-time observations had to be disregarded because no information about the station location could be obtained. Additionally, some of the instruments showed a reverse polarisation compared to the majority of P-wave polarisations, but we had no information about which SAPSE stations had reversed Z-components. However, few reversals (<20%) were accounted for in the focal mechanism determination.

Figure 3.19. Double-difference hypocentre relocations of events between November 2008 and April 2010 on map view and as cross sections along the X- and Y axis through the origin of the coordinate system. Clusters with more than eight events that are not in the largest cluster (with >700 events) are shown in colour. Black bars indicate hypocentre changes resulting from the relocation of the events.

Figure 3.20. Focal mechanism solutions of quality A (light grey) and B (dark grey) by Leitner et al. (2001) in comparison to the solutions obtained in this study (dark and light green). The number on top of the mechanism gives the minimum rotation angle (Arnold and Townend 2007) required to map one mechanism onto the other. Note that the hypocentre locations of the original and reproduced focal mechanism are not the same due to different location methods. All focal mechanism solutions shown are lower hemisphere projections with the coloured quadrants representing compressional first arrivals.

The minimum rotation angle or angular difference between the two focal mechanism solutions is determined using

$$\alpha_R = \cos^{-1} \left(\frac{\text{tr}(\mathbf{R}_1^T \mathbf{R}_2) - 1}{2} \right) \quad (3.1)$$

by Arnold and Townend (2007). $\mathbf{R}_1$ and $\mathbf{R}_2$ are rotation matrices corresponding to the two focal mechanism solutions. The rotation angles between Leitner et al.'s (2001) and our solutions range between 15.8° and 88.6° (Fig. 3.20).

The focal mechanism solutions obtained in this study were also compared with those of Ristau (2008), who determined focal mechanisms for earthquakes of $M_W \geq 3.6$ after December 2004 using regional moment tensor inversion. Only three solutions are listed for the time period addressed in this study (Table 3.7). The first event (8 May 2009) and the most recent event (14 August 2009) are in the SAMBA catalogue, but the focal mechanism solutions differ markedly from those of Ristau (2008). A comparison of the focal mechanism solutions and the distribution of P-wave polarities for these two events are shown in Fig. 3.21. The difference between the mechanisms obtained here and those of Ristau (2008) is attributed to the compressional first motion at the NNW rim of the stereogram, which constrains the NNW strike of one of the fault planes. This compressional polarity was recorded by station RPZ which has a correct polarity but a wrong instrument response (cf. Appendix A). Since one of our focal mechanism solutions is well-constrained by the distribution of first motions on the focal sphere, we suggest that Ristau's solution is influenced by the erroneous instrument response of station RPZ.

Date	Time	Lat [°]	Lon [°]	Strike [°]	Dip [°]	Rake [°]	Strike [°]	Dip [°]	Rake [°]	Mag
08.05.2009	00:34	-43.15	170.96	332	86	5	242	85	176	3.7*
	00:34:41	-43.15	170.93	151	48	-44	280	78	336	4.1
27.06.2009	18:32	-42.60	169.91	163	71	26	64	65	159	4.3*
14.08.2009	16:24	-43.25	170.79	343	73	24	246	67	162	3.7*
	16:24:09	-43.25	170.78	152	38	-29	186	73	48	4.1

Table 3.7. Comparison of hypocentre locations and focal mechanism parameters for events analysed by Ristau (2008) and in this study. Note that the magnitudes determined by Ristau (2008) (marked by *) are moment magnitudes M_W whereas in this study we have determined local magnitudes M_L .

A total of 209 focal mechanism solutions for earthquakes of $M_L > 0.44$ were obtained for the period November 2008 to December 2009 when the SAMBA network was operating and 165 solutions for January to April 2010 from the combined SAMBA/DFDP10 networks

Figure 3.21. Focal mechanism solution of two events that were analysed by Ristau (2008) (grey) and in this study (green, with uncertainties of 30.3° and 31.1°, respectively, as shown by the scale). The distribution of the first motions on the stereogram is plotted beneath for detail.

(Table K.1). In the latter period a maximum of 24 stations (including the three GeoNet stations) were recording simultaneously compared to 13 stations in the earlier time period (Fig. C.1). Therefore, the solutions obtained with the larger number of recording stations were chosen as reference mechanisms and compared with those solutions obtained earlier. Focal mechanism solutions were grouped by region (as shown in Fig. 4.7) and compared by sorting their hypocentre locations (from upper left to lower right in each region) and are listed in Fig. 3.22, H.1– H.3. This comparison shows that the distribution of P-wave polarities on the focal sphere is similar for groups of events and that only the number of polarities varies as indicated by the number at the upper right of the focal mechanism. The strike, dip and rake of the focal mechanism solution of similar events differ insignificantly except when the number of polarity picks is low (≤ 8). Then in some cases the solution exhibit large differences (minimum rotation angles $\geq 90^\circ$). This is especially the case when one polarity pick is wrong. For this reason, we computed the uncertainty of the focal mechanism and the minimum rotation angle for events with different numbers of P-polarities in Table 3.8.

The uncertainty of those mechanisms with the most P-wave polarity picks (≥ 14) ranges between $21\text{--}24^\circ$. It increases to an average of 30° for less picks, but does not exceed 35° for events with only 7 polarity picks. The minimum rotation angles between the best-constrained focal mechanisms (≥ 13 polarity picks) range between 1.5 and 91.9° , indicating different faulting types. In other words, we can discriminate individual mechanisms when the rotation angles are much larger than the uncertainties of the focal mechanism solutions. The mean minimum rotation angle between all focal mechanism solutions is 60.1° which is twice the mean uncertainty. This suggests that the mechanisms are different and that these can be distinguished from each other for earthquakes with polarity picks ≥ 13 .

The distribution of the P- and T-axis contours on the focal sphere (Fig. 3.22, H.1–H.3) reflect the similarity of the solutions better than the focal mechanism parameters due to many different mechanisms and the rather large uncertainties of the solutions. This comparison shows that composite focal mechanisms do not exhibit more information than the focal mechanism with the largest number of polarity picks. Sorting of the events in one region with depth, did not reveal any systematic changes in the focal mechanisms, as reported elsewhere (e.g. Imanishi et al. 2011).

Figure 3.22. (a) P-wave polarities and focal mechanism solutions for earthquakes within region 2. The number on the left specifies the event number in the cluster as listed in Table K.1. The number on the right states the number of P-wave polarities. All focal mechanism solutions shown are lower hemisphere projections with polarity picks in blue representing compressional first arrivals.

Figure 3.22. (b) Probability density contours of P- (orange) and T-axes (green) of the focal mechanism solution for earthquakes within region 2.

Num. P-pol.	Num. eq.	Mechanism uncertainty				Rotation angle			
		min [°]	max [°]	mean [°]	std. dev. [°]	min. [°]	max. [°]	mean [°]	std. dev. [°]
Region 4									
≥ 13	10	20.3	27.3	24.5	2.1	10.8	74.6	35.1	11.3
10–13	23	25.8	29.3	27.6	1.1	1.6	108.2	48.9	16.6
8–10	33	26.2	31.6	29.4	1.2	3.5	107.3	52.2	15.9
7	31	28.9	32.3	30.2	0.9	0.4	107.9	48.8	15.7
Region 6									
≥ 13	6	24.3	30.6	27.2	2.3	1.5	91.9	47.3	23.8
10–13	24	25.8	30.0	27.9	1.5	5.2	102.4	48.9	13.8
8–10	22	27.9	32.3	29.8	1.3	2.1	100.7	53.2	16.9
7	26	29.1	34.9	30.7	1.2	2.3	107.7	55.2	15.7
Region 2									
≥ 13	4	22.6	26.9	25.1	1.9	34.0	63.6	32.4	16.2
10–13	21	25.2	30.7	27.5	1.5	2.9	101.4	49.9	13.4
8–10	40	27.0	35.4	29.0	1.7	2.2	109.2	50.8	13.1
7	35	28.4	31.7	29.8	0.9	2.4	109.0	45.0	12.1
Region 1									
10	2	27.1	28.6	27.9	1.0	75.7	75.7	37.9	53.6
9	5	27.6	31.7	29.1	1.6	16.9	105.2	50.0	21.1
8	5	27.7	30.9	29.4	1.5	18.0	76.7	39.0	17.2
7	8	28.5	33.8	30.2	1.9	19.6	109.4	54.5	19.6

Table 3.8. Uncertainty of the focal mechanism parameters and rotation angles (for each minimum, maximum, mean and standard deviation of the mean) for focal mechanism solutions obtained for different numbers of P-polarity picks for regions shown in Fig. 4.7.

4. Microseismicity and stress in the vicinity of the Alpine Fault, central Southern Alps

This chapter has been published in the *Journal of Geophysical Research* (Boese et al. 2012) and summarises the results of the data from November 2008 to December 2009. The published version has been altered to fit the context of this thesis and to avoid repetitions. The introduction has been omitted because it summarised information provided in Chapter 1.2. The two last sections have been added and are not included in the publication. These revisit two topics discussed only briefly in the paper—the normal and shear stresses on the Alpine Fault plane and the seismicity gap in the area between the Whataroa and Wanganui rivers.

4.1. Methods

4.1.1. Earthquake hypocentre and uncertainty determination

Hypocentres have been computed using a modified version of the minimum 1-D velocity model determined by O' Keefe (2008) (Table 1.2), in which we have removed the uppermost layer, that represented a mean surface velocity layer for stations on sediments and bedrock in O' Keefe's analysis. This layer is unnecessary in our case, since the majority of the SAMBA stations are situated on hard rock. Shear-wave velocities are determined using a V_P/V_S ratio of 1.68 as obtained by O' Keefe (2008). This simple 1-D velocity model was favoured over existing 3-D models of Eberhart-Phillips and Bannister (2002) and Brikké (2010) due to its simplicity and similarity to 1-D velocity structures determined previously in active seismic reflection/refraction studies (Smith et al. 1995; Kleffman et al. 1998; Scherwath et al. 2002) and from earthquake inversion (Eberhart-Phillips 1995; Leitner et al. 2001). The 3-D velocity models lack sufficient resolution at shallow depths for our purposes, having been derived on regional scales with 8 km node spacing in the uppermost crust. Moreover,

the complex near-surface-structure of these models make them difficult to apply in this study.

Earthquakes are initially located using SEISAN and are then relocated using the non-linear location program NonLinLoc 5.0 (Lomax et al. 2000), to better constrain the posterior uncertainties in the hypocentre parameters (Section 2.2.2). We expand our 1-D velocity model (Table 1.2) across a grid with 1 km node spacing in all dimensions. The Alpine Fault is orientated at a 55° angle with respect to the grid. Synthetic travel-times for P- and S-waves propagating from each grid node to each station are calculated within NonLinLoc. That algorithm can be applied to models with sharp first-order velocity contrasts, and accurately produces arrival-times of the first-arriving wave (such as diffracted and head waves) using Huygens' principle.

The posterior probability density function of the earthquake hypocentre is found using a grid-search method (Lomax et al. 2009). We adapted the hybrid “Oct-Tree” search method described by Lomax et al. (2000). This method maps the probability of the earthquake location on a coarse grid by multiplying the probability density function (PDF) in the centre of each cell by the cell volume. High-probability cells are subsequently subdivided into eight new cells, generating a cascade of sampled cells and calculated probabilities. The procedure is repeated until a predetermined number of processed nodes has been reached, or the smallest specified node size is reached (Lomax 2011). Oct-tree sampling is an importance-sampling method in the sense that it uses information gained from the previous sample to refine the search. This sampling method allows the identification of multiple modes of the probability density functions. The complete PDF can only be determined using a systematic grid-search, but a suitable representation of the PDF can be found by having samples in proportion to the probability. The resulting hypocentre PDF describes the earthquake hypocentre uncertainty as a volume of irregular form, which quantifies the agreement between predicted and observed arrival-times and may contain multiple solutions. It reflects uncertainties due to the network-event geometry, arrival-time reading errors and calculated travel-time errors. The final earthquake hypocentre location is selected as the point of maximum likelihood within the PDF solution.

Several parameters have to be predefined in NonLinLoc, which directly influence the uncertainty volume. These parameters are

- σ^{model} , the typical model error of the travel-time residual,
- L_{corr} , which is an estimate of the characteristic length scale of the velocity anomaly between two stations,

- $\Delta\sigma^{\text{model}}/\sigma^{\text{model}}$ which gives an estimate of the model error on the total uncertainty in proportion to the travel-time,
- $\sigma_{\text{min}}^{\text{model}}$ and $\sigma_{\text{max}}^{\text{model}}$, which set lower and upper limits on the travel-time error due to unmodelled velocity structure.

The first two parameters, σ^{model} and L_{corr} , are needed for the calculation of the model-error covariance matrix (Tarantola and Valette 1982b), which describes whether a correlation in the model errors exists between two stations. We tested values of 1, 4, 8 and 80 km for L_{corr} , with the latter two representing the average and the maximum station distance. The differences in the resulting uncertainty volumes were minimal and we set L_{corr} to 8 km, an anomaly length which we expect to be able to resolve with the station spacing.

In order to estimate the Gaussian uncertainty of the arrival-time readings and the calculated travel-times for the earthquakes recorded in this study, we have performed a comprehensive analysis of the travel-time residuals for the SAMBA data (Section 3.1.1 and Table 3.2). This analysis showed that the typical model uncertainty σ^{model} of the travel-time residual is small for the SAMBA station network in comparison to the reading uncertainty (Table 3.2). We obtain estimates of the reading uncertainty by two independent means. First, we determine the standard deviation of the travel-time residuals for all stations, and second, we analyse lag-times obtained from cross-correlation of similar events in a swarm. Error limits for the model uncertainty are here set to 0 and 0.2 s. The maximum value was obtained for station POGR which exhibits (pre-weighting) mean model uncertainties of 0.17 and 0.16 s for P- and S-phases. This station is the only station located in a sediment-filled valley, so the assumed velocity model is not fully suitable for this setting. Our estimate for the model error as a fraction of the whole travel-time is obtained for a sample set of 46 earthquakes, distributed across the study area and recorded at all stations. We calculate values of 0–9% with mean values of 3.3% for the P- and 1.6% for the S-phase travel-times. We assume hereafter that 5% of the travel-time is a reasonable representation of the velocity model uncertainty.

4.1.2. Magnitudes

Earthquake magnitudes are determined from waveform amplitudes corrected for geometric spreading and attenuation (equation 2.23). Amplitude-readings are obtained from half the peak-to-peak displacement on one of the horizontal components of a pseudo-Wood-Anderson seismograph, simulated in SEISAN (version 8.2.1, October 2008, and 8.3, May 2010) (Havskov and Ottemöller 2000). It is assumed that the gain of the Wood-Anderson

instrument is 2080 ± 60 , as reported by Uhrhammer and Collins (1990).

To determine magnitudes for the SAMBA data, we solve the general inverse problem

$$\mathbf{Y} = \mathbf{X}\mathbf{m} + \mathbf{e} \quad (4.1)$$

by using LU-matrix factorisation. The parameters involved in the equation are as follows:

- $\mathbf{Y}$ is a two-part vector containing the amplitude readings and the geometrical spreading term $A_j + \alpha \log_{10} \Delta_j$ of the j -th event, and the magnitudes M_j^c from the New Zealand earthquake catalogue as reported by GeoNet, which are used for calibration of the SAMBA magnitudes M_j .
- $\mathbf{X}$ is a matrix representing which component of each station has an amplitude reading and which events have calibration magnitudes M_j^c . $\mathbf{X}$ also contains the hypocentral distance Δ_{jk} , where j marks the event and k the station.
- $\mathbf{m}$ are the parameters of interest, namely the SAMBA magnitudes M_j , the attenuation term 0.4343γ and the station correction terms S_k .
- $\mathbf{e}$ is a vector of residuals for the SAMBA magnitudes ΔM_j and the GeoNet magnitudes ΔM_j^c used for calibration.

We have tested the dependency of the data on geometric spreading by separately inverting for the geometrical spreading coefficient α only (Section 3.2.1). We obtain $\alpha=1.6$ (Fig. 3.8). For geometric spreading due to expansion of the direct earthquake waves in space, we would expect $\alpha=1$. We therefore assume $\alpha=1$ and attribute the remaining amplitude decay to the effects of scattering and anelastic attenuation on the magnitude. Considering the hypocentral range and the dominant frequency of the wave, we find that the influence of attenuation can be up to half that of the geometrical spreading.

The anelastic attenuation parameter γ is known to be frequency dependent $\gamma=\gamma(f)$, as seen from the relationship between γ and the quality factor Q (equation 2.17). Eberhart-Phillips et al. (2008) studied 3-D attenuation in central South Island by analysing spectra of 5695 vertical-component velocity seismograms of earthquakes larger than M_L 2.5. For the observed frequency range of the local earthquakes of 2–40 Hz, Eberhart-Phillips et al. (2008) were able to fit the spectra equally well with both frequency-independent and frequency-dependent Q . Abercrombie (1998, and references therein) found a clear change in frequency dependence of $Q(f)$ at 10 Hz. Although Eberhart-Phillips et al. (2008) determined Q_P , we assume here that Q_S can be treated

similarly as frequency-independent, so we set $\nu = 0$ and $Q = Q_0$. We invert for $\gamma = \gamma_0 f$ according to equation 2.17.

4.1.3. Focal mechanisms

We determine focal mechanism solutions from P-wave first motion polarities for earthquakes recorded by the SAMBA array using the Bayesian approach of Walsh et al. (2009). This method treats observational uncertainties probabilistically in order to account for polarity errors arising from low signal-to-noise ratios and incorrectly wired seismometers, hypocentre location uncertainties, and imprecise knowledge of the seismic velocity structure. The prior probability of a falsely wired seismometer is here set to 20%, as suggested by Hardebeck and Shearer (2002) who analysed inconsistencies amongst emergent polarity picks of similar event clusters in the Northridge area, California. We do not discriminate between impulsive and emergent polarity picks, although Hardebeck and Shearer (2002) found that only 10% of the impulsive polarities were inconsistent. Uncertainty in the hypocentre location results in a variety of possible ray paths from the hypocentre to the receivers and, therefore, in a range of possible take-off angles and azimuths. This causes a cloud of corresponding positions on the focal sphere. The density and the distribution of these points depend predominantly on the uncertainty in depth, since this is usually larger than the horizontal uncertainty. As an example, a change in depth of 1 km for earthquakes shallower than 7 km can cause a change in the take-off angle of up to 15° for up-going rays Hardebeck and Shearer (2002), assuming a 1-D velocity model with a gradient of $dV/dz=0.25 \text{ s}^{-1}$ and $V_0=4.7 \text{ km/s}^{-1}$. Hardebeck and Shearer (2002) and Walsh et al. (2009) investigated the effect of the velocity model on the polarity distribution on the focal sphere. Hardebeck and Shearer (2002) found that the influence of the velocity model on the focal mechanism is larger than uncertainty in hypocentre location. Since take-off angles at shallow depth are most sensitive to the velocity model, the focal mechanisms of deeper earthquakes show smaller changes resulting from changes in the velocity model (McKenzie 1969; Hardebeck and Shearer 2002).

To characterise the uncertainty of the focal mechanism solution, we use the estimate of the standard deviation σ_Θ of the focal mechanism parameters by Arnold and Townend (2007) (equation 2.42). This approach assumes equal errors in strike, dip and rake.

4.1.4. Stress inversion

Assuming that individual focal mechanisms represent a uniform state of stress and are driven by the same stress tensor, regional stress estimates can be derived by inverting a set of independent focal mechanism observations. Stress inversion aims to find the principal stress axes that are consistent with all the focal mechanism solutions in a geographically confined area. We apply the Bayesian approach developed by Arnold and Townend (2007) which accounts for observational errors and fault plane ambiguities in the focal mechanism solutions to estimate tectonic stress parameters.

4.2. Results

4.2.1. Earthquake locations and their uncertainties

A total of 1791 events were recorded between November 2008 and December 2009 in the study area. All events recorded by more than five SAMBA stations (87.3%) are shown with their uncertainties in Fig. 4.1 (a), in comparison with the long-term seismicity in (b), and with magnitudes in (c). In order to show the depth variation of the hypocentre distribution along the strike of the Alpine Fault, the events are projected onto depth profiles in Fig. 4.2.

The error volume projected onto the horizontal plane is approximately circular for most of the earthquakes within the station network (Fig. 4.1), reflecting a uniform distribution of the recording stations around the events (azimuthal gaps for events in the inner triangle range between 57° and 192° ; see also Fig. 4.4).

The uncertainty in depths is controlled by the station distribution and varies throughout the network. Depth uncertainty increases markedly when there is no station within a distance of 1.5 focal depths of the earthquake. For this reason, vertical uncertainties increase towards the boundaries of the SAMBA network (as seen in depth transects 1, 2 and 7 of Fig. 4.2). Since the resolution of the hypocentre uncertainty varies throughout the study area, mean depth and mean uncertainties are displayed by grey bars in the background of the profiles in Fig. 4.2. Hypocentre depths are best constrained in the vicinity of the Alpine Fault where the station distribution is the densest (as can be seen by the increase in the error bars with distance from the Alpine Fault).

The distribution of the minimum hypocentral distances of all events to the nearest

Figure 4.1. (a) Epicentral locations of events recorded by more than 5 stations between November 2008 and December 2009. Earthquake hypocentres were determined using the non-linear location program NonLinLoc. The probability density function (red) of each earthquake represents the uncertainty of the hypocentre location, the blue dot marks the maximum likelihood hypocentre location. The light blue area shows which earthquake hypocentres are projected onto SIGHT transect T1 as shown in Fig. 4.3. (b) Seismicity recorded in the central Southern Alps since 1984 by GeoNet (grey), Reyners (1988) (blue), Leitner et al. (2001) (red), O'Keefe (2008) (green) and in this study (black).

Figure 4.1. (c) Epicentres as shown in (a) with the symbol size representing the magnitude. Coloured lines and numbers indicate the depth profile onto which events within ± 5 km are projected (Fig. 4.2). Letters are used to mark seismicity clusters described in the text. In the background, a simplified geological map based on the geological data from the 1:1000000 Geological Map of New Zealand (Officers of the New Zealand Geological Survey 1972 with limited updates by P.J. Forsyth and N. Mortimer (2004)) is shown. Textural zones are I Quaternary deposits, II Palaeozoic and Mesozoic Western Province basement (gneiss, granite, granodiorite), III Rakaia Terrane mylonite, curly schist and schist, IV Rakaia Terrane schist, V+VI Rakaia Terrane semi-schist, VII Rakaia Terrane greywacke sandstone and mudstone according to Cox and Sutherland (2007)

Figure 4.2 Hypocentral depth and depth uncertainty of earthquakes projected onto the corresponding profiles (number and colour) shown in Fig. 4.1 c), with the symbol size representing the magnitude of the event. Grey bars indicate mean depth and mean uncertainties (over seven points) for the events shown in the profile. Letters a–d indicate the position of clusters (Fig. 4.1 c) along the profile. Diamonds indicate the surface position of the Main Divide Fault Zone. The dashed line indicates the Alpine Fault for an assumed dip of 45° .

stations, as an indicator of the range of well-constrained depths, is bimodal with modes at 7 and 20 km. Its mean is $14 \text{ km} \pm 9.5(1\sigma) \text{ km}$. Consequently, deeper events are better constrained in the whole study area.

A comparison of the long-term background seismicity recorded by GeoNet during the last 20 years with the short-term seismicity recorded by SAMBA over 14 months (Fig. 4.1 b) demonstrates distinct spatial patterns of quiescence and persistent activity in the epicentral distribution. Earthquakes are spatially clustered and occur in patches along and to the east of the Main Divide Fault Zone (MDFZ) (Fig. 1.6), a major backthrust off the Alpine Fault (Cox and Findlay 1995). The earthquakes are not confined to mapped surface fault traces (as shown in Fig. 4.1 a) but are distributed throughout an approximately 25 km-wide region. However, there is some indication of reduced seismic activity in a 15 km-wide zone adjacent to the Alpine Fault that coincides with the average width of the Alpine schist (Little et al. 2005; Cox and Sutherland 2007) (textural zones III and IV in Fig. 4.1 c). This may reflect different amounts and characteristics of fractures in the schist relative to the greywacke further southeast.

Four distinct clusters stand out in the seismicity distribution (letters in Fig. 4.1 c). Cluster a) occurs in the middle of the SAMBA array, has well-constrained depths $<7 \text{ km}$, and comprises small events of $M_L < 2.5$. Cluster b) may represent continuing aftershocks of the M_W 6.1 Godley Valley earthquake. Cluster c) is confined to a northeast-trending high-seismicity zone which occurs within 30 km of the Alpine Fault. It borders a distinct aseismic gap that extends north-westwards to the surface trace of the Alpine Fault between the Wanganui and Whataroa rivers (Fig. 1.8). The fourth cluster, d), occurs at the northeastern boundary of the study area, has poorly-constrained depths and significantly larger magnitudes ($M_L < 3.5$). The depth distribution of clusters b, c and d can at best be partially resolved.

Fig. 4.2 shows the earthquake hypocentres projected onto profiles perpendicular to the Alpine Fault. Three of these transects coincide with geological cross sections by Little et al. (2005). The north-easternmost profile 1 (orange hypocentres) shows distinct clusters of events approximately 8 and 16–25 km southeast of the fault. The latter cluster coincides with the MDFZ (as indicated by a black diamond in all profiles). In the Whataroa river valley (transect 2, blue hypocentres), few earthquakes occur close to the Alpine Fault. Events cluster at about 30 km distance from the Alpine Fault (cluster c) over a wide depth range. These depths are quite uncertain (mean depth uncertainty $5 \pm 1.6(1\sigma) \text{ km}$) as there is no station in the vicinity, but the hypocentres align with a steeply-dipping fault mapped at the surface (Fig. 3a of Little et al. 2005). To the south, profile number 3 (purple hypocentres)

reveals a distinct earthquake cluster within 5 km of the surface trace of the Alpine Fault and a number of events between 15 and 40 km, increasing in depth with distance from the fault. Transect 4 (green hypocentres) shows several events 2–8 km southeast of the surface trace of the Alpine Fault at depths consistent with the estimated position of the fault. Abundant earthquakes with shallow hypocentral depths occur at 15 km distance to the fault trace (cluster a). At greater distances, the earthquake depths increase monotonically. Profiles 5 (red hypocentres, average uncertainty ± 2.6 km) and 6 (brown hypocentres, average uncertainty ± 3 km) indicate a bulge in the seismicity with a radius of curvature of ~ 10 km despite the depth uncertainty. Events near the Alpine Fault and 20 km southeast of the surface trace occur at larger depths than those events centred around the MDFZ (depths of 6 km). In this region, the distance between MDFZ and the Alpine Fault is narrowest (15 km versus 30 km elsewhere; Cox and Findlay 1995). Transects 5 and 6 exhibit few events 45–50 km from the Alpine Fault trace, but in general earthquakes occur within a narrow zone close to the Alpine Fault. Transect number 7, which straddles the southern boundary of the SAMBA network, reveals a cluster of events 4–8 km southeast of the surface trace at depths possibly associated with the Alpine Fault. The remaining events exhibit increasing depths with distance from the fault but the uncertainty in depth is quite large (4 ± 2.9 km).

Figure 4.3 shows the projection of earthquakes within 20 km on either side of SIGHT transect 1 (light blue area in Fig. 4.1 a) and six large earthquakes since 1946 from elsewhere (Fig. 1.6) onto the resistivity profile by Wannamaker et al. (2002). The depth-distribution of those earthquake hypocentres reveals that the seismogenic zone varies from northwest to southeast across the Southern Alps but nowhere it exceeds 17 km. The seismicity correlates with the high-resistivity regions and is clustered where the resistivity gradient is high (see also 4.12). Wannamaker et al. (2002) interpreted the low-resistivity regions as zones of interconnected fluids, which are trapped within the mylonitised, ductile shear zone. The six largest events (white circles, events are shown in Fig. 1.6) occur at greater depths on average than the lower-magnitude seismicity, probably at the base of the seismogenic zone (Sibson 1984) in this region.

A histogram of the depths of all events is shown in Fig. 4.4. This distribution has a mode at 6 km depth due to the large number of shallow events in the centre of the SAMBA array (profiles 4 and 5 in Fig. 4.2). The second maximum at the surface results from inconsistent phase arrival times or hypocentre locations outside the station network with poor depths resolution.

Figure 4.3. Hypocentre locations and depth uncertainty of local earthquakes (within 20 km of the transect, black) and large earthquakes (in the Southern Alps region, white, see Fig. 1.6 and Table 1.1 for details) projected onto SIGHT transect T1, for which a resistivity model has been derived by Wannamaker et al. (2002). Note that the average topography is 1.5 km in this area and that for distances ≥ 60 km, earthquakes fall outside the recording range of the SAMBA network. Depth uncertainties are calculated from the maximum length of the uncertainty volume determined using NonLinLoc. Reprinted by permission from Macmillan Publishers Ltd: Nature (Wannamaker et al. (2009)), ©2009

4.2.2. Magnitudes

We obtain a frequency-dependent attenuation factor $\gamma(f) = \gamma_0 f$ with $\gamma_0 = 1.89 \pm 0.02 \times 10^{-3}$ s/km, resulting in $\gamma = 1.89 \times 10^{-2}$ km $^{-1}$ at 10 Hz (representative of distances < 70 km) and $\gamma = 0.95 \times 10^{-2}$ km $^{-1}$ at 5 Hz (≥ 70 km) (Fig. 4.5). Our result is in agreement with estimates by Robinson (1987) ($\gamma = 0.67 \times 10^{-2}$ km $^{-1}$) and O'Keefe (2008) ($\gamma = 1.69 \times 10^{-2}$ km $^{-1}$), depending on the hypocentral distance of the earthquake to the station. The corresponding frequency-independent quality factor Q for direct S-waves is $Q_S = 531 \pm 53$, a reasonable value in comparison with the Q_P of 250–800 determined by Eberhart-Phillips et al. (2008) for the central Southern Alps. Eberhart-Phillips et al. (2008) found that a sharp boundary exists between weakly attenuating schist and highly attenuating regions of altered schist for the shallow Q_P -structure in the central Southern Alps (depths < 10 km). Q_P values of ~ 800 were attributed to the Haast schist, which is inferred to be depleted of fluids and highly compacted due to dehydration during metamorphism and intense deformation, respectively. In contrast, Torlesse greywacke has moderate to low Q_P of ~ 450 . Low crustal Q values typically result from fluid saturation and high temperature. Eberhart-Phillips et al. (2008) state that the Q_P depth profile in the central Southern Alps closely matches the resistivity model of Wannamaker et al. (2002). A correlation of the seismicity distribution with low Q_P was observed by Eberhart-Phillips et al.

Figure 4.4 Depth, root-mean squared (RMS) residual and maximum azimuthal gap distribution for all events shown in Fig. 4.1 c)

(2008) and interpreted to represent highly-fractured crust and increased fluid volumes in conjunction with fractures. The Q_P values for the Haast Schist and the damage zone in the central Southern Alps, assuming that $Q_P \sim Q_S$, can be used to constrain the attenuation parameter γ (according to equation 2.17), yielding $\gamma(Q_P=800)=0.5 \times 10^{-2} \text{ km}^{-1}$ and $\gamma(Q_P=200)=3.45 \times 10^{-2} \text{ km}^{-1}$ at a frequency of 5 Hz and an S-wave velocity of 3.2 km/s. These values provide upper and lower limits for the attenuation parameter. The attenuation coefficients we obtain for frequencies between 3 and 17 Hz lie within these bounds.

We find that it is necessary to fit the amplitude readings with a frequency-dependent attenuation coefficient. Otherwise, we obtain a high constant attenuation coefficient of $\gamma \sim 2 \times 10^{-2} \text{ km}^{-1}$ and uniformly positive magnitude residuals for hypocentral distances greater than 70 km (Fig. 4.5). The magnitudes determined using our frequency-dependent attenuation coefficient are not significantly different from those obtained by GeoNet ($1/N \sqrt{\sum (M_j^c - M_j)^2} \leq 0.1$). However, our preferred attenuation parameter produces consistent results for earthquakes of different sizes and frequency contents.

The magnitudes obtained for earthquakes recorded in the study area span $-0.3 \leq M_L \leq 4.2$ (Fig. 4.6). Earthquakes of low magnitude ($M_L \leq 2.5$) occur in great numbers in the centre of the station network, conversely events with magnitudes $M_L \geq 3$ are distributed to the northeast of the SAMBA array (Fig. 4.1 c).

The magnitudes determined in this study were calibrated using the magnitudes obtained by GeoNet (e.g. Ristau 2009). For 147 of the larger earthquakes ($1.6 \leq M_L \leq 4.2$) magnitude estimates are available (Fig. 4.6). However, magnitude calibration implies that if the reference values are consistently overestimated, so will be the SAMBA magnitudes. Comparisons of M_W estimates obtained from waveform modelling (Ristau 2008; Leitner et al. 2001) with M_L estimates for the central Southern Alps indicate that M_L values overestimate the actual energy release by at least 0.2 but more likely 0.5 magnitude units. Deichmann (2006) estimated that local magnitude errors due to uncorrected attenuation effects for events of $M_W < 1$ can amount to more than a whole magnitude unit. Due to the frequency-dependent attenuation factor, we consider the effect of uncorrected attenuation to be small. However, the size of the events relative to each other is more reliable than the absolute value, which might be uncertain by approximately half a magnitude unit. Once there are more M_W estimates from moment tensor inversion available for earthquakes recorded by SAMBA (currently only two events), we will repeat our magnitude inversion using these M_W for calibration.

Figure 4.5. Period of the maximum Wood-Anderson displacement amplitude, attenuation functions and magnitude residuals versus hypocentral distance. The frequency dependence of the maximum amplitude needs to be fitted using a frequency dependent attenuation parameter $\gamma(f)$ to produce normally distributed magnitude residuals over all hypocentral distances.

Figure 4.6. (a) Histograms of the magnitude distribution in the central Southern Alps region recorded by SAMBA (blue) and GeoNet (red) stations. The black line indicates the slope of the Gutenberg-Richter relation (b-value of 0.86) of the cumulative number of earthquakes versus magnitude. (b) Residuals of the magnitudes used for calibration in the inversion for all SAMBA magnitudes.

4.2.3. Focal mechanisms

Focal mechanism solutions for earthquakes recorded by the SAMBA network were determined for events with seven or more P-wave polarity picks. Two deep teleseismic events (M_W 6.6 of 4 October 2009, 10:58:03.6 UTC at 626 km depth and M_W 6.9 of 28 August 2009, 01:51:25.9 UTC at 634 km) with impulsive P-wave arrivals have been used to check the instrument polarities (Fig. B.1). This analysis revealed reversed polarities of the Z-component of the Mark Products L-4C3D sensors in comparison with the short-period borehole and the GeoNet broadband sensors. At one site, Reynold's Creek (REYN), all polarity picks have been removed from the dataset, since they are inconsistent with those of the other sensors.

Polarities of closely-spaced events have been checked in groups for their consistency. Reversed polarity picks were identified for the stations equipped with a borehole and a surface sensor, when the polarity was picked on the Z-component of the Mark Products L-4C3D surface sensor instead of the borehole sensor. These wrong polarity picks are easy to identify by visual inspection. Based on this checking procedure, the assumed probability of inconsistent polarity picks of 0.2 (Hardebeck and Shearer 2002) is likely to overestimate the number of inconsistent polarities in our dataset, since the polarities of the instruments are generally known and wrong polarity picks should be caused by misidentification only.

Due to the network size, the maximum number of available *P*-wave polarities is restricted to 13. Consequently, good coverage of polarity readings on the focal sphere is uncommon. However, this is an inherent problem of focal mechanism solutions. Those mechanisms obtained for the central Southern Alps region have standard deviations of the focal mechanism parameters of $\sigma_\Theta = 29 \pm 1.6^\circ (1\sigma)$. Events of this type were considered to be poorly constrained by Walsh et al. (2009) who compared their focal mechanism solutions with those of Reyners and McGinty (1999) for events with polarity picks from up to 36 stations. Their quality estimate was based on the angular difference between the focal mechanism solutions (equation 3.1). In Walsh et al.'s study the poorly-constrained focal mechanism solutions had minimum rotation angles of $21\text{--}84^\circ$, whereas for the well-constrained solutions the minimum rotation angle was typically $5\text{--}26^\circ$. Minimum rotation angles of $24\text{--}32^\circ$ were found by Hardebeck (2006) for three high-quality focal mechanism datasets with events a few hundreds of metres apart. Hardebeck (2006) concluded that these focal mechanism solutions were indistinguishable within error.

We obtained minimum rotation angles of $1\text{--}88^\circ$ (with a mean of $32 \pm 22^\circ$), for the

focal mechanism solutions of groups of events with similar polarity picks. We conclude that these values fairly describe the uncertainty of the focal mechanism solutions in our dataset. One well-defined nodal plane for the focal mechanisms was often observed, indicating that the strike and dip are well constrained but not the rake.

Figure 4.7. Focal mechanism solutions of 211 events derived in this study (coloured according to the uncertainty) and by Ristau (2008), Leitner et al. (2001) and Anderson et al. (1993) (dark grey, light grey and black, respectively). The symbol size represents the magnitude of the event, which is scaled up for the SAMBA data. Detailed information about each focal mechanism solution is listed in Table K.1.

Hardebeck and Shearer (2002) showed that focal mechanisms of earthquakes at shallow depths (<10 km) are most sensitive to changes in depth, especially for dip-slip events. This is because the take-off angles for up-going rays are sensitive to changes with the depth of the location. Ray take-off angles to the station are determined from the maximum likelihood hypocentre, which is in the best case centred in the middle of the uncertainty volume. Since we consider the depth uncertainty by calculating a range in take-off angles, we expect that these mechanisms are not sensitive to small changes in depth. The comparison of our focal mechanism solutions for the central Southern Alps with those of Leitner et al. (2001), Ristau (2008) and Anderson et al. (1993) (Fig. 4.7) shows overall similarity in the mechanisms but differences in the details of the focal mechanism parameters. These differences usually

result from the increased uncertainty of one focal mechanism parameter.

Considering the uncertainty in the focal mechanism solutions, it is notable that the majority of our solutions exhibit consistently strike-slip mechanisms. We find approximately equal numbers of reverse and normal faulting mechanisms which form half of all solutions obtained for the studied microearthquakes. The ratio of reverse to strike-slip mechanisms becomes equal for intermediate (black and grey solutions in Fig. 4.7) and large earthquakes (Table 1.1) in the central South Island. The expected ratio of reverse to strike slip mechanisms would be 1:2 because the fault-normal displacement amounts to approximately one third of the fault-parallel component in the Southern Alps region.

Focal mechanism solutions have been determined for events as small as $M_L=0.44$. Although the station coverage is best in the vicinity of the Alpine Fault, we have obtained only a few focal mechanism solutions for earthquakes on or close to the fault.

4.2.4. Stress inversion

The prevailing directions of the principal stresses and the maximum horizontal compression obtained from analysing clusters of focal mechanisms are shown in Fig. 4.8 (a). The number of events used for the inversion is indicated by the number next to the stress contour plots. Different clusters exhibit only minor differences in the direction of the maximum horizontal compressive stress S_{Hmax} (Table 4.1) from the average value of $S_{Hmax}=115\pm10^\circ$ at the 80% confidence level. Except for the north-easternmost cluster, where the directions of S_2 and S_3 overlap, the principal stress directions are well-defined and consistent over the study area. S_2 is near-vertical except for the south-westernmost region. Here, a rotation of the S_2-S_3 plane around the S_1 direction occurs, which increases the reverse component in this area. The stress ratio R , in contrast, varies throughout the study area. Although this parameter is the least constrained in the inversion (Table 4.1), it reflects the topography throughout the region with low R values where the topography is highest ($S_2 \gg S_3$) and R ranges between 0.5 ($S_2=0.5(S_1+S_3)$) and 1 ($S_2=S_3$) where the topography decreases. Lund and Townend (2007) showed that the maximum horizontal compressive stress S_{Hmax} is less sensitive to variations in R for strike-slip stress states than for normal or thrust faulting stress states.

Our results are in agreement with previous stress measurements from the South Island. These results are summarised in Table 4.2 and the most recent results are displayed along with our results in Fig. 4.8 (b). These consistent angles in the range 107° to 121° are as expected for the principal axis of contraction (111°) inside a uniformly deforming zone parallel to the

Figure 4.8. (a) Principal stress directions ($S_1 < S_2 < S_3$) and maximum horizontal compressive stress (S_{Hmax}) derived from focal mechanism inversion of groups of events as shown in Fig. 4.7. The number on top of the stereogram indicates the number of focal mechanism solutions inverted for the principal stresses. The encircled number in the dashed rectangles gives the cluster number for which results are listed in Table 4.1.

(b) Stress inversion results of the 211 focal mechanism solutions in comparison with the maximum horizontal stress directions derived in other studies for different regions of northern and central South Island. The inset shows a summary of the stress directions in comparison to the strike of the Alpine Fault (blue) and its respective thrust and strike-slip segments (light and dark grey).

Cluster	S ₁	S ₂	S ₃	R	S _{Hmax}
1	121.1°/8.8°	65.0°/65.9°	27.4°/22.2°	0.25 ^{+0.303} _{-0.250}	118±16.8°
2	115.0°/13.7°	115.7°/76.3°	25.0°/0.0°	0.15±0.09	115±10.2°
3	115.0°/20.4°	56.6°/67.8°	23.2°/8.3°	0.77 ^{+0.211} _{-0.321}	114±16.6°
4	123.7°/9.4°	53.8°/63.3°	29.1°/24.8°	0.69±0.239	122±13.1°
5	119.2°/5.4°	49.1°/74.6°	27.7°/14.4°	0.47±0.306	118±14.5°
6	123.7°/24.4°	67.2°/52.0°	22.1°/27.4°	0.89 ^{+0.097} _{-0.197}	122±12.0°

Table 4.1. Principal stress directions (trend/plunge) and maximum horizontal compressive stress of all events in one region as shown in Fig. 4.8 (a). $R=(S_1-S_2)/(S_1-S_3)$ specifies the stress ratio. The uncertainty in R is estimated from the upper and lower bounds given by the 10% and 90% percentiles of the posterior probability density function. The posterior distribution is skewed near values of 0 and 1 because the R-value is confined to lie within these limits.

S _{Hmax} or S ₁	Region	Method	Reference
300°/30°	northern Marlborough	IPP	McGinty et al. (2000)
118°/6°	southern Marlborough	IPP	McGinty et al. (2000)
299±17°	Marlborough	FMI	Balfour et al. (2005)
291°	Lake Tennyson	FMI	Balfour et al. (2005)
107±6°	west of Arthur's Pass	TT+GPS	Pearson et al. (1995)
116±9°	east of Arthur's Pass	TT+GPS	Pearson et al. (1995)
119±17°	north of Mt. Cook	FMI	Leitner et al. (2001)
121±36°	south of Mt. Cook	TT	Blick et al. (1989)
84°/16°	Jackson Bay	IPP	McGinty et al. (2005)

Table 4.2. Maximum compressive stress S₁ (trend/plunge) or maximum horizontal compressive stress S_{Hmax} (with uncertainties if reported) for South Island measurements obtained in previous studies. Abbreviations used to describe methods are: IPP inversion of P-wave polarities, FMI focal mechanism inversion, TT triangulation/trilateration

Alpine Fault subjected to loading at an azimuth of 077° by the relative plate motion (Reilly 1990).

All stress measurements indicate a strike-slip faulting regime prevailing over wide areas of the South Island. A homogeneous, regional stress field has been noted before by Leitner et al. (2001), whose principal stress directions for three broad regions in central South Island were indistinguishable from each other at the 80% confidence level. Nevertheless, we see some indication for a locally increased reverse component at the southern boundary of our study area. The high topography in the central Southern Alps has little effect on the overall stress field in general, but causes local segmentation of the Alpine Fault's surface trace in thrust and strike-slip segments (Norris and Cooper 1995). Based on the stress estimates, we ob-

serve that these segments have orientations close to optimal in terms of Andersonian faulting models (insert Fig. 4.8 b). Under the assumption that the horizontal stress is homogeneous, variations in the intermediate stress S_2 cause a rotation of the strike of a non-vertical strike-slip fault towards S_1 when there is a deficiency of load, and away from S_1 as a result of excess topography, according to the equation

$$\frac{S_2 - S_3}{S_1 - S_3} = 1 - R = \sin^2 \phi \quad (4.2)$$

given by Fialko et al. (2005) (where ϕ is the angle between the fault trace and the S_1 -axis). The topography therefore explains the segmentation into strike-slip faults coinciding with major river valleys crossing the Alpine Fault and thrust segments at higher angles to the Alpine Fault in regions of high elevation near the fault (see also Fig. 9 and 11 of Norris and Cooper 1995).

Our stress inversion results show that the Alpine Fault as a whole is orientated at a high angle (59°) to the direction of maximum horizontal compressive stress (Fig. 4.8 b). This implies that the normal stress acting on the fault plane is high. Regarding Andersonian faulting, this angle suggests that the Alpine Fault sustains a low ratio of shear stress to effective normal stress. Balfour et al. (2005) obtained a 60° angle between the average strike of faults in the Marlborough Fault System and the direction of the maximum horizontal compressive stress. In Southern Marlborough, where the strike of the faults changes to about 070° , this angle lessens to 44° . The San Andreas Fault is orientated at angles ranging between 60° and 85° to the maximum horizontal compressive stress (Townend and Zoback 2004). In Southern California, this angle between the San Andreas Fault and the maximum horizontal compressive stress direction is almost constant, despite pronounced changes in the strike of the San Andreas fault (Townend and Zoback 2001, 2004).

Low coefficients of friction have been suggested as the cause for the weakness of the San Andreas Fault (Townend and Zoback 2004). This has been confirmed by the low values ($\mu=0.13-0.21$) measured for the creeping section near Parkfield (Lockner et al. 2011). Balfour et al. (2005) concluded that the Marlborough faults are also frictionally weak due to either low coefficients of friction (coefficient of friction $\mu=0.35$, pore fluid factor $\lambda=0.4$), or increased pore fluid pressure ($\mu=0.58$, $\lambda=0.7$) or due to some combination of the two. This interpretation seems applicable to the Alpine Fault, too. No indication of steady creep on the Alpine and Wairau Fault has been observed since 1964 on man-made structures crossing the fault (Evison 1971; Sutherland et al. 2007, and references therein) or by GPS surveying in the Whataroa river (Beavan et al. 1999), but cannot be ruled out in the mountainous region

east of the Alpine Fault. Since high pore fluid pressures have been suggested as the cause for the low-velocity zone (Stern et al. 2001) in the vicinity of the Alpine Fault, we favour increased fluid pressures as the cause for the fault's reduced strength.

4.3. Discussion and conclusions

Hypocentre locations of more than 1700 earthquakes recorded by the SAMBA network have been determined between November 2008 and December 2009 in the central Southern Alps. The seismicity is diffuse and no particular fault or a small number of faults can be identified from the dense distribution of mapped faults in this region. However, the overall earthquake hypocentre distribution correlates with the boundaries between high- and low-attenuation zones as shown in the crustal attenuation model of Eberhart-Phillips et al. (2008), despite the rather large node spacing in the central Southern Alps. The Alpine Fault does not appear as a localised zone of deformation from the earthquake distribution. Instead, the inferred deformation is distributed between the Alpine Fault, the Main Divide Fault Zone and other major backthrusts to the east. Many active large faults in New Zealand appear aseismic on intermediate time-scales and are not delineated by the background seismicity (e.g. Anderson and Webb 1994; Robinson 2004). The Alpine Fault, however, is not totally aseismic because it exhibits several clusters of earthquakes 5–8 km southeast of its surface trace at depths consistent with a 45°-dipping fault. Earthquakes also occur at greater depths closer to the surface trace of the Alpine Fault. Therefore, the existence of a subvertical strand of the Alpine Fault, as expected from ~15 Myr of solely strike-slip motion on the fault (e.g. Cande and Stock 2005), cannot be ruled out. A high concentration of earthquakes on backthrusts at a range of distances from the Alpine Fault is observed, most obviously in the northernmost depth profiles (profiles 2 and 3 and marked by cluster b and c). Otherwise, the hypocentre locations are distributed throughout the whole seismogenic zone (<17 km), but mean depths generally increase to the southeast across the Southern Alps perpendicular to the Alpine Fault. Earthquakes are anomalously shallow (<6 km), of small magnitude ($M_L < 2.5$) and highly clustered within the SAMBA array beneath the area of highest topography. In the vicinity of Mt. Cook, a bulge in the seismicity occurs with its maximum centred beneath the Main Divide Fault Zone at depths of 6 km (transect 6 of Fig. 4.2). The same depth has been identified by Stern et al. (2007) as the upper boundary of a low P-wave velocity zone that extends downwards to depths of 30 km. A similar structure can be seen just south of Mt. Cook in the crustal seismic reflection profile of Long et al. (2003). Long et al. (2003) describe an “antifold” of 15 km half wavelength with its crest at 3 km depth and imaged to 10 ± 2 km depth. They interpreted that this structure

accommodated 1.8 km (=12%) horizontal crustal shortening in less than 1 Myr at current GPS strain rates. At the southwestern boundary of the SAMBA network, the seismically active zone is narrow and earthquake depths increase rapidly in the direction perpendicular to the strike of the Alpine Fault.

A depth of the seismogenic zone of 12 ± 2 km has been suggested from seismicity studies in the central Southern Alps (Leitner et al. 2001; Reyners 1988) and 5–12 km has been suggested as the locking depth based on GPS measurements (Beavan et al. 1999; Beavan et al. 2004; Wallace et al. 2007). We find that the seismogenic depth varies across the Southern Alps with depths of 10 ± 2 km beneath the surface trace of the Alpine Fault and 8 ± 2 km within 20 km distance of the fault. At distances larger than 20 km, it increases in depth to a thickness of 15 ± 2 km, similar to the shape of the high-resistivity zone. New earthquake hypocentre locations recorded by an additional short-term deployment of 12 stations in the area between the Wanganui and the Whataroa rivers during January and May 2010 indicate that our results for hypocentre depths along SIGHT transect T1 (Fig. 4.3) and the depth profiles 1–3 (Fig. 4.2) are robust and slightly overestimate the earthquake depths. This will be discussed in more detail and in comparison with double-difference relocations in future studies.

The lateral variation in the near-surface resistivity structure correlates well with the distribution of microseismicity. Our results show that no earthquakes are generated in the conductive zone $<100 \Omega\text{m}$ at depths greater than 20 km. The material is either too hot, too fluid-saturated or too weak (or a combination of all three) to produce earthquakes of recordable size. Earthquakes are generated in the high-resistivity zones adjacent to or above the low-resistivity regions. We find that clustering of the events is strongest where the resistivity contrast is highest. A similar distribution of microearthquakes has been observed for the San Andreas Fault where resistivity measurements revealed a fault zone conductor with resistivities $<10 \Omega\text{m}$ (Unsworth and Bedrosian 2004). The microseismicity occurs below or adjacent to these zones in high-resistivity zones. Similar distributions of earthquakes on the boundary between high- and low-resistivity regions have been observed for microearthquakes (Ichiki et al. 1999), swarms (Tank et al. 2003) and large earthquakes and their aftershocks (Eberhart-Phillips et al. 1990; Kasaya et al. 2002). Takeda et al. (1999) observed the occurrence of earthquakes at shallow depths in areas of low water saturation and high crack density surrounded by areas of high saturation. They attributed the generation of earthquakes to fluid flow into cracks in rocks with a low degree of saturation. Kato et al. (2010) observed intense seismic activity at shallow depths above a high-velocity body of intruded diorite, causing high heat flow and the release of fluids that invaded and fractured the overlying rock.

We suggest that extensive fracturing occurs in the central Southern Alps due to ascending fluids and that the Main Divide Fault Zone (Cox et al. 1997) and other backthrust faults act as active fluid conduits. A resistivity model by Wannamaker et al. (2009) for a transect crossing the Marlborough Fault System shows similar low resistivity regions $<100 \text{ } \Omega\text{m}$ at depths greater than 10 km beneath the Awatere and Clarence faults. As observed for the central Southern Alps, the crustal seismicity in this region is also confined to high-resistivity bodies. Wannamaker et al. (2009) attribute this seismicity to fault fracture meshes created by the upwards migration of fluids from the subducting slab of the Pacific plate.

Notably, all of the largest, more recent earthquakes (marked by a star in Table 1.1) in the central South Island occurred southeast of the Alpine Fault. These events occurred at the lower end of the earthquake depth-distribution, presumably near the base of the seismogenic zone. If we assume that the hypocentre locations determined in this study reflect the vertical extent of the seismogenic zone, then the seismogenic depth reaches a maximum at distances $>25 \text{ km}$ perpendicular to the surface trace of the Alpine Fault. At these distances, the seismogenic zone has almost twice the thickness of the seismogenic depth near the Alpine Fault. In other words, earthquakes generated in this area can potentially be larger than further west due to the vertical extent of the seismogenic zone.

The microearthquake locations in the region bounded by the Wanganui and the Whataroa river reveal in more detail the aseismic zone noted previously (e.g. Leitner et al. 2001). Within the vicinity of the Alpine Fault ($<10 \text{ km}$), the seismicity does not change significantly along strike. The area with reduced seismicity extends from 15 to 30 km of the fault and is bounded by the cluster c) (Fig. 4.1 c). The distribution of earthquakes is symmetric around this gap, but the number of events is not (Fig. 4.1 a and c). Earthquakes at the northeastern boundary of the low-seismicity region are larger in magnitude and more frequent than at the southwestern limit. The density of mapped faults in the aseismic region is reduced, too. The detection threshold of events at the southwestern limit of this area is better than to the northeast due to the station distribution. If similar sized events occurred at the southern boundary, they would have been recorded by the SAMBA stations. The low-seismicity area is characterised by a low-attenuation block in the attenuation model of Eberhart-Phillips et al. (2008). Changes in the geology with flat-lying sequences of schist and semi-schist occur in this region, as noted by Leitner et al. (2001). However, there is no obvious correlation between bedding dip and the distribution of earthquakes. The remnants of a major ice plateau are located in this area, but the plateau's exact location does not

coincide with the observed distribution of events. However, the seismicity gap could be associated with seismic quiescence following pronounced activity associated with localised bending and uplift resulting from rapid ice loss since the end of last glacial period 14000 yr ago (Suggate 1990). On the other hand, if that area forms a seismic gap, it could be a seismic gap of the second kind (Mogi 1979), which is characterised by quiescence in smaller magnitude activity before a large earthquake. If this is the case, the rupture area amounts to at least $\sim 300 \text{ km}^2$ corresponding to a magnitude 6.5 earthquake. Since similarly large earthquakes have occurred at some distance from the Alpine Fault, further detailed studies in this region are necessary.

The magnitudes of the recorded earthquakes in the study area span $-0.3 \leq M_L \leq 4.2$ but are smaller in the region of highest uplift, where the upper magnitude threshold is $M_L 2.5$. Based on our magnitude scale, we deduce that high-frequency waves ($f > 10 \text{ Hz}$) are attenuated more than twice as much as low-frequency waves ($f < 5 \text{ Hz}$). The frequency-independent attenuation coefficient used in routine magnitude determination New Zealand-wide thus underestimates the local magnitude if the distance between the earthquake source and the recording station is small ($< 70 \text{ km}$).

Focal mechanism solutions for microearthquakes recorded in this study indicate predominantly strike-slip faulting. For intermediate-size and large earthquakes as recorded over longer timespans, the ratio of reverse to strike-slip mechanisms is equal. This suggests that with time larger amounts of compressional strain are accumulated and released in larger magnitude events. Stress inversion of our focal mechanisms reveals a homogeneous strike-slip stress field in the central Southern Alps except for localised topographic modifications. The direction of maximum horizontal compressive stress is $S_{Hmax} = 115 \pm 10^\circ$. The resulting angle between the average strike of the Alpine Fault and S_{Hmax} is 59° as characteristic of a weak fault. However, small-scale thrust segments are oriented at angles of $65^\circ - 95^\circ$ and strike-slip segments are oriented at angles ranging between 25° and 50° to S_{Hmax} , implying that shallow segmentation of the Alpine Fault takes place on favourably-oriented structures.

4.4. Shear and normal stresses on the fault plane

We have determined the prevailing directions of the principal stresses and the maximum horizontal compression by analysing clusters of focal mechanisms using Arnold and Townend's (2007) algorithm and the stress transformations described by Lund and Townend (2007)

(Fig. 4.8). This method allows us to determine the directions of the three principal stresses and the stress ratio but not the absolute magnitude of each principal stress. To get a sense of the absolute stress values acting on the inclined plane of the Alpine Fault, we assume that S_2 is vertical and that its magnitude can be calculated from the overburden pressure $S_v = \rho gz$, where ρ is the assumed density of the rock (2700 kg/m³ according to Godfrey et al. 2000), g the gravitational acceleration and z the depth. The differential stress for strike-slip stress regimes is

$$\Delta S = 2\rho gz(\lambda - 1)(1 - F)/(1 + F) \quad (4.3)$$

with the pore fluid factor λ and $F = (\sqrt{\mu^2 + 1} + \mu)^2$ according to Zoback and Townend (2001). Figure 4.9 shows the absolute values with depth calculated for each principal stress assuming a stress ratio R of 0.5. Also displayed are the normal and shear stresses on the Alpine Fault determined using the method of De Paor (1990). For hydrostatic fluid pressure at depth ($\lambda=0.37$), the differential stress increases similarly to S_3 . The normal stress on the fault plane increases similarly to S_2 , whereas the shear stress is only one fifth that of the normal stress. For lithostatic fluid pressure, the differential stress is small (less than half of S_3 and approximately one-fifth of S_n). For a steeper dip (60°) of the Alpine Fault plane, the normal stress reduces slightly and the shear stress increases with depth. This is consistent with the apparent weakness of the Alpine Fault, implying that faulting occurs in response to low levels of shear stress (compared to normal stress and stress drop during the earthquake).

Since the stress ratio is the least constrained value in the stress inversion (uncertainties ± 0.2 ; Table 4.1), the influence of the stress ratio for a fixed depth of 6 km was examined (Fig. 4.10). The larger the stress ratio, the lower the shear component and the larger the normal component of stress on the Alpine Fault. This trend is independent of the fault's dip and the assumed pore fluid factor. In conclusion, the ratio of shear to effective normal stress is low on the Alpine Fault and decreases with increasing λ and R , implying near-lithostatic fluid pressures within the Alpine Fault zone, and similar absolute values of the intermediate stress S_2 (near-vertical) and the least compressive stress S_3 (near-horizontal) represent the weakest scenario.

4.5. Seismicity gap between the Whataroa and Wanganui rivers

A zone of sparse seismicity between the Whataroa and Wanganui rivers (Fig. 1.8) has been described by Eiby (1971), Evison (1971), Scholz et al. (1973), Caldwell and Frohlich

Figure 4.9. Principal stress directions and normal and shear stress on the fault plane with depth assuming a stress ratio R of 0.5, a vertical S_2 and horizontal S_1 and S_3 with orientations of 116° and 206° . The figures on the left are for lithostatic, those on the right for hydrostatic fluid pressure. Stresses in the top row were calculated for a 45° dipping Alpine Fault, whereas those for the bottom row result for a steeper fault dip of 60° .

Figure 4.10. Principal stress directions for a fixed depth of 6 km and varying stress ratio R. See Fig. 4.9 for details.

(1975), Eberhart-Phillips (1995) and Leitner et al. (2001). Our results show that the seismicity in the direct vicinity (within 10 km) of the fault is not significantly different from that southwest and northeast of this area and the gap extends from ~15 to 30 km southeast of the Alpine Fault (Fig. 4.11). The gap is bounded by a zone of high-seismicity rates and magnitudes to the northeast. At the southwestern boundary of the gap, earthquakes are on average one magnitude unit smaller, however there have been three events of $M_L \geq 4$ as reported by GeoNet since 1975 in this area (Fig. 4.11). In general larger events ($M_L \geq 4$) occur at the edges of seismicity clusters indicating stress concentrations in these areas. Figure 4.12 shows a more detailed comparison (than Fig. 4.3) of the depth distribution of events within 10 km of SIGHT transect T1 with the resistivity profile by Wannamaker et al. (2002). It shows that earthquakes follow the 390 Ωm resistivity contour, which appears to determine the depth of the seismogenic zone in this area.

Well-recorded earthquakes (≥ 7 stations) are shown in Fig. 4.13 in comparison with the attenuation structure obtained by Eberhart-Phillips et al. (2008). The gap correlates with high Q /low attenuation at shallow depths (≤ 5 km) whereas the earthquakes correspond to low Q /high-attenuation zones adjacent to the gap. This supports our interpretation that the gap represents a strong, unfractured block, surrounded by a highly-fractured damage zone. The deformation that occurs between the Alpine Fault and the Main Divide Fault Zone further northeast and southwest is diverted to larger distances southeast of the Alpine Fault.

The area between the Whataroa and Wanganui rivers is not the only low-seismicity region that stands out in short- and long-term seismicity rate studies. The Mackenzie Basin southeast and the Rangitata river valley east of the study area show equally low levels of seismicity (cf. Fig. 1.7). Since all three areas correspond to major valleys that have been glaciated (Suggate 1990), an alternative interpretation of the low seismicity to date is isostatic readjustment resulting from unloading due to deglaciation 14000 yr ago. As several studies have shown (see below), deglaciation can increase the slip rates of faults significantly for a few thousand years after which the fault returns to its previous slip rate. Therefore, the hypothesis is that the current quiescence is the result of pronounced seismic activity previously associated with isostatic readjustment.

According to Suggate (1990), the Southern Alps region has experienced rapid ice-loss since the last glacial maximum. The Whataroa river valley is currently in a transitional state between glacial and non-glacial form (Herman and Braun 2006, and references therein). The Perth glacier (that drains via the Perth river into the Whataroa river) currently has the second largest retreat rate of 44 m/a exceeded only by the Godley glacier (with 66 m/a;

Figure 4.11. Seismicity in the Whataroa-Wanganui gap along profiles parallel to the Alpine Fault. Events with better depth resolution recorded by DFDP10 stations between January and April 2010 are in colour. Events in cyan are earthquakes of $M_L \geq 4.0$ recorded by GeoNet since 1975. See Fig. 1.8 for locations.

Figure 4.12. Seismicity within 10 km of the SIGHT transect T1 (black) projected onto the resistivity profile of Wannamaker et al. (2009). White circles are large earthquakes in the central South Island. See Fig. 4.3 for details.

Figure 4.13. Seismicity in comparison with the attenuation structure at 5 (left) and 9 km depth (right) obtained by Eberhart-Phillips et al. (2008). Hypocentres within 2 km of the depth slice are shown in black.

Chinn 1996). The Whataroa–Perth valley systems also experienced substantial denudation (Herman et al. 2010, and references therein). Consequently, extensive erosion could be an contributing factor to the unloading. The glaciers in the Mackenzie Basin experienced a major retreat since the last glacial maximum (Suggate 1990). The geomorphology of the Rangitata Valley shows that no significant glacier re-advance could have occurred between 16.4 ka and 13.7 ka (Shulmeister et al. 2011). The ice volumes affected in the three areas are approximately 338, 121 and 248 km³ (corresponding areas 2375, 1750 and 1400 km²). Since isostatic adjustment is of importance for large and wide volumes affecting areas of regional scales, we only consider here the effect of this load removal on the principal stresses. In general, changes in the vertical stress due to the loss of overburden would be up to 10 MPa. This is only a small fraction of the stresses at depth as determined in Section 4.4. The effect would be largest near the surface, where the differential stress is low. In principle, if the rapid unloading causes the vertical stress (S_2 in the study area) to become comparable or less than the minimum horizontal stress (S_3) this would increase thrust faulting in the Southern Alps region. This may explain why the number of reverse faults mapped in outcrops east of the Main Divide (Cox et al. 2012) is much larger than the numbers of reverse focal mechanisms obtained for microearthquakes in this study (as shown in Fig. 4.7). On a regional scale, however, the effect of deglaciation appears to be significant. The volume of ice lost in the Southern Alps (~5550 km³ over an area of 54000 km² as modelled by N. Golledge, pers. comm., 2012) is overall comparable to the volume of water lost from Lake Bonneville between the Bonneville highstand (17.5 ka cal BP) and the Provo highstand (16.7 ka cal BP; see Karow and Hampel 2010) and the area affected (that experienced 69 m of uplift due to isostatic readjustment in the case of Lake Bonneville). Therefore, deglaciation may have played an important role on slip-rate variations of the Alpine Fault and should be considered in this area (Turpeinen et al. 2008)

Increases of slip-rates of major normal faults in the Wasatch Fault Zone have been documented and associated with the regression of Lake Bonneville and deglaciation in the Wasatch and Uinta Mountains in the Late Pleistocene/Early Holocene (e.g. Karow and Hampel 2010, and references therein). Hetzel and Hampel (2005) use finite difference models to show that glacial unloading decelerates slip because S_1 decreases, but isostatic rebound decreases S_3 and promotes normal faulting which can explain the observed slip-acceleration. These slip-rate accelerations can explain clustering of seven large earthquakes on the Wasatch front during the Holocene. Karow (2010) showed by finite element modelling that thrust faults show an increase of seismicity in response to unloading (that starts 0.5 ka after unloading and lasts ca. 3.5 ka after a loading period of 17 ka in his model), as seen after the deglaciation of Scandinavia (e.g. Stewart et al. 2000). Strike-slip

faults are affected similarly. Large earthquakes within the vicinity of the Dead Sea appear to coincide with intervals of lake regression (Karow 2010, and references therein).

5. The characteristics of earthquake swarms in the central Southern Alps

This chapter combines a summary of our current understanding of earthquake swarms and triggered seismicity with observations made in this study about these topics. The latter is written in the form of a publication. Katrina Jacobs has contributed to the obtained results on earthquake swarms in the Southern Alps. Amongst others, she has analysed the earthquake rates in different regions, from different networks, and for different time periods as described in Section 5.2.1, developed the CURATE algorithm to identify sequences as described in Section 5.3, determined the probabilities of triggered swarms (Section 5.5) and found the observed inter-event patterns that are shown in Section 5.6.

A statement by Hill et al. (1993) provides the context for this chapter: “Whatever the triggering process, the results [are] a cascading failure sequence (earthquake swarm) in crustal volumes already loaded to a critical stress state”.

5.1. General aspects of earthquake swarms and triggering

5.1.1. Earthquake swarms

The term “earthquake sequence” describes earthquakes related in space and time. “Earthquake swarms” are sequences of earthquakes which do not exhibit an event of greatest magnitude at, or near, the beginning of the sequence as characterises a “mainshock-aftershock sequence” (e.g. Mogi 1963) (Fig. 5.1). Swarms stand out as discernible disturbances in the background seismicity. In non-volcanic settings, swarms are often

Figure 5.1. Classification of the terms earthquake sequences, swarms and mainshock-aftershock sequences as used in this study based on original definitions by Mogi (1963). Note that a “distinct mainshock” is generally considered to be separated from its aftershocks by a magnitude separation value exceeding half a magnitude unit.

associated with rapid fluid redistribution in the vicinity of fault zones (Vidale and Shearer 2006) or with aseismic creep (e.g. Roland and McGuire 2009). This association has been established on the basis of their short temporal distribution (termed “bursts” by Vidale and Shearer 2006) and observations of hypocentre migration away from the point of initialisation.

One well-known example of a swarm is the Matsushiro earthquake swarm that occurred in Japan between 1964 and 1967 (Hagiwara and Iwata 1968; Nakamura 1969). The extensive and widely distributed seismicity (700,000 earthquakes in an elliptic area of about 34 km in length and 18 km in width, Hagiwara and Iwata 1968; Cappa 2010) was accompanied by surface cracking, 70 cm of ground uplift and the outflow of large volumes of groundwater (Mogi 1989; Yamashita 1999). The swarm was associated with pore-pressure changes induced by the M 7.5 Niigata earthquake that occurred nearby in 1964 (Mogi 1989).

In general, fluid migration follows zones of weaknesses (Grünthal et al. 1990; Miller et al. 2004) or aligns with permeability anomalies associated with lithological boundaries (Audin et al. 2002). Mogi (1963) postulated that tectonic swarms occur in regions that are highly fractured. At mid-ocean ridges, swarming is a characteristic style of seismic energy release (e.g. McGuire 2008; Roland and McGuire 2009). Sometimes, earthquake swarms precede large earthquakes, as in the case of the L’Aquila earthquake of 6 April 2009 (e.g. Papadopoulos et al. 2010) or the Izu-Hanto-Toho-Oki earthquake sequence of 1980 (Utsu et al. 1995, and references therein). Swarms in volcanic settings are attributed to the intrusion of magma, especially dikes. Benoit et al. (1996) and Vidale et al. (2006) observed

that volcanic swarms have longer durations than tectonic swarms and that the hypocentral depths are shallower in volcanic swarms. Vidale and Shearer (2006) further reported the predominant occurrence of swarms on vertical planes or linear streaks. This led them to suggest that swarms are predominantly produced on strike-slip faults.

Earthquake swarms differ from mainshock-aftershock sequences in many respects, and each is most likely caused by a different physical mechanism. Aftershocks are observed in regions of stress concentrations within and adjacent to the rupture zone of the mainshock. Coseismic static stress-changes, that diminish approximately in proportion to r^{-3} , cause loading of some faults in the vicinity of the mainshock rupture zone which respond with increased aftershock activity (King et al. 1994). Other factors, however, such as afterslip, pore-fluid pressure readjustments, viscoelastic readjustments, high dynamic stresses induced by directivity effects, and geological structures have been demonstrated to play a significant role in the aftershock distribution (e.g. Fialko 2004; Kilb et al. 2000). Dilatational fault jogs, for example, exhibit pronounced aftershock activity, most likely due to reduced mean stresses and pore-fluid pressures (Sibson 1987) causing fluid flow into these areas. Similarly, the comparisons of aftershock locations with pore-pressure changes in wells shows that aftershocks cluster in areas that have experienced a decrease in pore-pressure during the mainshock and fluid inflow afterwards (Jónsson et al. 2003). Kilb et al. (2000) showed a correlation between the seismicity rate change and the peak Coulomb stress change associated with dynamic waves in the aftershock zone of the Landers earthquake. Due to the rupture being unilateral, the high dynamic stresses that normally cannot be distinguished easily from the coseismic static stress changes could be identified in this case and used to explain the observed aftershock distribution.

Vidale and Shearer (2006) observed that earthquake swarms contain more earthquakes than mainshock-aftershock sequences for comparable magnitudes of the largest events (their Fig. 7). This fact is reflected in the high b -values commonly reported for swarms (e.g. Sykes 1970). The total duration of an individual mainshock-aftershock sequence as determined using a magnitude threshold is at least a factor of three smaller than that of an earthquake swarm (Vidale et al. 2006). However, the spatial distribution is more extensive for mainshock-aftershock sequences than for earthquake swarms. Vidale and Shearer (2006) observed that the proportion of mainshock-aftershock sequences with respect to the total number of sequences in Southern California is approximately 20%. Swarms made up 25% of the total whereas the remaining sequences were classified as “other”.

5.1.2. Earthquake triggering

Remotely triggered seismicity is commonly observed following earthquakes of $M_W \gtrsim 7$ at locations that may be hundreds of kilometres away from the hypocentre of the mainshock (e.g. Hill et al. 1993; Steacy et al. 2005). At these distances, the dynamic stresses caused by the passage of the surface waves ($>10^4$ Pa) are larger than static stress changes from the mainshock (<100 Pa) as pointed out by Brodsky and Prejean (2005) for the Long Valley caldera in California. In some cases, static stresses may even be smaller than tidal stress levels (e.g. Hill et al. 1993). Remote dynamic triggering has been observed in different tectonic settings worldwide (Velasco et al. 2008). However, seismically active regions in extensional tectonic regimes (e.g. Hill et al. 1993) appear more susceptible to dynamic triggering than others. In particular, young volcanic centres, geothermal areas and regions of active faulting frequently exhibit triggered earthquakes (Hill et al. 1993; Husker and Brodsky 2004). If the minimum compressive principal stress direction lies in the horizontal plane, as in strike-slip and normal faulting stress regimes, the differential stress required for fault rupture of dynamically triggered events is less than when the minimum principal stress is vertical (e.g. Hill et al. 1993; Harrington and Brodsky 2006). Consequently, rock under extension and high fluid pressures may undergo failure in response to small stress changes. Husker and Brodsky (2004) identified regions of high heat flow with large volumes of total dissolved solids as favourable for dynamic triggering.

Seismically active regions (with high background seismicity rates, corresponding to high tectonic loading rates) are also more susceptible to remotely triggered seismicity (e.g. Gomberg et al. 2004). This may be due to the larger number of faults close to failure in these areas. Although triggered events seem to generally occur in regions of high background seismicity, recent observations have shown that the largest number of triggered events in Greece occurred in a region of relatively low levels of seismicity (e.g. the Arta cluster of Brodsky et al. 2000). Remotely triggered swarms have also been observed in previously aseismic areas of Long Valley caldera (Prejean et al. 2004). In other words, the contrary view has been suggested as well: if the stressing rate is low, the perturbing stress changes are large relative to ambient stressing rates (Steacy et al. 2005). Gomberg et al. (2004) suggested that triggered seismicity in the western USA occurs in the transitional zone between old, cold, strong crust and younger, warmer and weaker crust.

In summary, the tectonic settings that repeatedly exhibit remotely triggered seismicity are commonly volcanic and geothermal settings, probably because the conditions are most favourable to dynamic triggering, but remote triggering is not restricted to these settings.

Moreover,

- both Rayleigh and Love waves from earthquakes cause dynamic stress changes that can remotely trigger seismicity (Velasco et al. 2008, see next section);
- induced seismicity involves brittle slip on critically stressed faults favourably oriented in the prevailing stress field (e.g. Hill et al. 1993; Gomberg et al. 2004);
- triggering represents clock-advance of failure on the fault (e.g. Anderson et al. 1994; Steacy et al. 2005); and
- areas undergoing swarm activity frequently respond more often to dynamic stress changes (e.g. Hill et al. 1993).

5.1.3. Previous observations of earthquake triggering

Dynamic triggering has been observed in many places, notably in the western USA, Alaska, Syria, Greece, Iceland and China. Observations of remotely triggered seismicity in California are particularly well described in the literature and summarised below. Locations named in the text are shown in Fig. 5.2.

Long Valley caldera in eastern California appears particularly susceptible to earthquake triggering. Elevated levels of seismicity occurred after the 1992 M_W 7.3 Landers (Hill et al. 1993; Brodsky 2006), the 1999 M_W 7.1 Hector Mine (Gomberg et al. 2001) and the 2002 M_W 7.9 Denali earthquakes (Prejean et al. 2004; Brodsky and Prejean 2005). This region is monitored with a dense seismometer network and the catalogue completeness is low (cut-off magnitude of 1.2 according to Brodsky and Prejean 2005). Hill et al. (1993) reported 340 earthquakes within seven days of the Landers earthquake. These events coincided with a transient contractional strain pulse of about 45 days' duration that was recorded by one strainmeter and inferred to result from increased fluid pressure (Hill et al. 1993). Approximately 200 events of $M \leq 3$ occurred in Long Valley Caldera after the Denali earthquake (Prejean et al. 2004).

At the Geysers geothermal field in northern California, triggered seismicity has been observed for several earthquakes of $M \geq 6.6$ (Stark and Davis 1996), including the 1989 M_W 6.9 Loma Prieta and 1992 M_W 7.1 Petrolia earthquakes (Hill et al. 1993). The largest amount of triggered seismicity occurred in response to the 1992 M_W 7.3 Landers earthquake (Stark and Davis 1996). The triggered seismicity consisted of a swarm of 60 events of $M < 1$ (Hill et al. 1993), with locations that mimicked the background seismicity (Stark and Davis

Figure 5.2. Map of California showing key locations named in the text and areas in which remotely triggered seismicity was observed (rectangles). Faults in California are adapted from Jennings (1975).

1996).

The Coso geothermal field exhibited triggered seismicity after the Hector Mine, Landers and Denali earthquakes (Prejean et al. 2004). Following the Hector Mine earthquake, 26 small, unlocatable events were recorded (Prejean et al. 2004). After the Landers earthquake, 44 events of $M \leq 4.4$ were located (Prejean et al. 2004). A total of 80 earthquakes of $M \leq 2.3$ were triggered in response to the Denali earthquake (Prejean et al. 2004).

A small cluster of earthquakes (16 events) occurred 15 km northwest of Yellowstone caldera after the Landers earthquake. Abundant seismicity (~ 600) of $M < 3.2$ was observed inside and outside the Yellowstone caldera following the Denali earthquake (Husen et al. 2004). Redistribution of hydrothermal fluids inside the caldera and alteration of the fault properties in the northwestern part of Yellowstone outside the caldera have been suggested as causes for the triggered seismicity (Husen et al. 2004).

Increases in seismicity were observed in southwestern Utah in the Hurricane fault zone after the Landers earthquake (60 earthquakes; Hill et al. 1993), and along the 500 km-long section of the Intermountain Seismic Belt (including the 380 km-long Wasatch Fault Zone) in central Utah (Pankow et al. 2004) after the Denali earthquake. In the latter case, the majority of earthquakes (65 events of $M \leq 2.6$ from a total of ~ 110 of $M \leq 3.2$) occurred within the first day.

Two swarms of $M < 0.9$ occurred at Mt. Rainier in the central Washington Cascades following the Denali earthquake (Prejean et al. 2004). Several sites of triggered seismicity were identified by Husker and Brodsky (2004) in Idaho and western Montana after the Denali earthquake. After the Landers earthquake, 51 triggered earthquakes of $M_L \leq 1.7$ occurred near Cascade, Idaho (Hill et al. 1993). Gomberg et al. (2001) reported triggered seismicity at Indio, the Salton Trough and Cerro Prieto, Mexico, after the Hector Mine earthquake. Anderson et al. (1994) reported 227 events in Nevada within 83 days following the Landers earthquake. Other places in the western United States, where triggered seismicity occurred have been documented by Hill et al. (1993).

At Katmai volcano, Alaska, seven locatable and four unlocatable events occurred after the Denali earthquake. Triggered earthquakes have been observed following three previous large earthquakes in the greater region (Moran et al. 2004, and references therein). After the M_W 9.2 Sumatra–Andaman islands earthquake in 2004, a swarm of 14 microearthquakes ($M \leq 1.9$) was observed at Mount Wrangell, Alaska (West et al. 2005). An

earthquake swarm ($M < 3.7$) of 21 events in southwestern Syria, along the Dead Sea Fault system, was reported by Mohamad et al. (2000) after the M_S 7.3 Gulf of Aqaba earthquake of 22 November 1995. This swarm occurred in a region where the background seismicity rate is 0.5–1 events per day. Brodsky et al. (2000) observed triggered seismicity in several places in Greece following the M_W 7.4 Izmit earthquake in 1999. Lei et al. (2011) detected significant rate changes after the 2004 M_W 9.3 Sumatra earthquake in southwestern China. The triggered events were spatially clustered at step-overs, fault junctions, fault ends and bends. Several locations in northern China responded with remotely triggered seismicity to the 2008 M_W 7.9 Wenchuan earthquake (Peng et al. 2010). These events occurred predominantly in aftershock zones of recent large events. Lei et al. (2011) and Peng et al. (2010) concluded that remote triggering occurs where the ground is damaged and where large numbers of faults and fractures are present.

Not all regions susceptible to dynamic triggering exhibit triggered seismicity. In Japan, only one example of triggering occurred after the great (M 8) Nankaido earthquake in 1946 (Hill et al. 1993). Harrington and Brodsky (2006) searched for triggered earthquakes after several large events in Japan and found very little evidence. They came to the conclusion that stresses cannot accumulate over time because Japan is so frequently shaken by large-amplitude, low-frequency events. After the M_W 9.0 Tohoku earthquake, a significant increase in the seismicity was observed throughout Japan (Toda et al. 2011). This increase was inferred to be caused partly due to large static stress changes (≥ 0.1 MPa or 1 bar) over large parts of Japan. In some regions in northern Japan, the stress field changed due to a rotation of the principal stress directions (Yoshida et al. 2012). Large events (M 5–6) occurred within 3 rupture lengths of the mainshock (Miyazawa 2011). At distances ≥ 200 km the static stresses were smaller than dynamic stresses caused by the surface waves and at ≥ 500 km the dynamic stresses were one order larger than the static stresses. Miyazawa (2011) showed that the seismicity triggered by dynamic stresses followed a triggering front consistent with surface wave propagation. According to Miyazawa (2011), all delayed triggered earthquakes were inferred to be triggered due to static stress changes.

In many places, triggered seismicity begins with or shortly after (within minutes to hours) the arrival of the Love or Rayleigh waves (West et al. 2005; Prejean et al. 2004). This has been shown by Velasco et al. (2008) for triggered seismicity around the globe with an increase of 37% above the background rate observed after the Love surface waves and an overall 60% increase after the Rayleigh waves. The observations of triggered seismicity by West et al. (2005) at Mt. Wrangell showed that the triggered earthquakes align in phase with the maximum vertical 30 s Rayleigh wave that exceeded an amplitude threshold of 0.25 cm

ground displacement. This phase corresponded to maximum horizontal extensional stresses of <25 kPa at those shallow depths where the triggered events occurred. West et al. (2005) reported this triggering to be consistent with simple shear failure on normal faults.

In many cases, the small triggered events occur during or immediately after the passage of the surface waves. These events can be obscured by the waveforms of aftershocks. However, some evidence for delayed triggering also exists. Gomberg et al. (2001) observed a delay of 1.35 days at Cerro Prieto after the Hector Mine earthquake. Mohamad et al. (2000) reported a swarm 167 min after the Aqaba earthquake at ~ 500 km distance, followed by a two-month period of quiescence. Similarly, at Yellowstone caldera, where seismicity begins with the passage of the surface waves, one isolated swarm of more than 20 events occurred after a delay of 8 days following the Denali earthquake in the West Thumb geyser basin (Husen et al. 2004). One burst of seismicity delayed by 13.4 days after the Denali earthquake occurred amongst clusters in the Intermountain Seismic Belt in central Utah (Pankow et al. 2004). One swarm of 112 events of $M \leq 3$ delayed by 23.5 hr relative to the onset of the Denali earthquake surface waves occurred at the Long Valley caldera (Prejean et al. 2004). The Little Skull Mountain earthquake, the largest event possibly triggered by the Landers earthquake, took place 22 hr after the mainshock (Gomberg and Bodin 1994; Bodin and Gomberg 1994). Gomberg and Bodin (1994) proposed aseismic slip on faults in the vicinity of Little Skull Mountain triggered by the Landers earthquake as the cause for this delay. Syracuse et al. (2010) reported a swarm of seismicity two days after the 2007 M_W 8.1 Solomon Islands earthquake at Kilauea volcano. They inferred fluid migration to be the cause for this swarm. Lei et al. (2011) observed event clusters resembling mainshock-aftershock sequences immediately (20 min) and several hours (7.5 hr) after the Sumatra earthquake. Other remotely triggered events (clusters 3–6) occurred with greater delays (≥ 14 hr) and exhibited swarm-like characteristics (as a progressive increase in magnitude). Table 5.1 summarises these observations of delayed triggering.

In general, dynamically triggered seismicity lasts for a few hours to several days (e.g. Husker and Brodsky 2004; Brodsky 2006), but in several cases seismicity has persisted for weeks (Hill et al. 1993; Husen et al. 2004; Pankow et al. 2004). Brodsky (2006) showed for the Long Valley caldera that the triggered events following the Landers earthquake resemble aftershock sequences from M_L 4.1 earthquakes. The late, triggered events can be explained as aftershocks of those events triggered by the surface waves. Accordingly, more than 80% of the triggered earthquakes could be explained via a process of local aftershock generation (Brodsky 2006). However, this mechanism does not explain isolated sequences delayed by several days at Cerro Prieto, Yellowstone or Utah.

Interval times between sequences of triggered events range from a few months at the Geysers geothermal field (Stark and Davis 1996) to three years at Long Valley caldera (Brodsky and Prejean 2005). However, this time may depend on the background seismicity rate and the seismic energy released in the previous swarm. Remotely triggered events occur in the uppermost crust (e.g. 20% of all events at Long Valley occur at depths <3 km; Brodsky and Prejean 2005), <5 km at Yellowstone (Husen et al. 2004), around 2 km or less at Mt. Wrangell (West et al. 2005) and at depths shallower than 4 km for the majority of events at the Geysers (Stark and Davis 1996).

Mainshocks that excited most of the reported remotely triggered seismicity (the Landers, the Hector Mine and the Denali earthquakes) ruptured unilaterally (Hill et al. 1993; Kilb et al. 2000; Gomberg et al. 2001, 2004) and regions that exhibited remotely triggered seismicity were in line with the maximum radiated seismic energy. Rupture directivity causes a strong focusing effect of seismic waves and increases the dynamic stress in the direction of rupture propagation (Gomberg et al. 2003). The decrease in maximum magnitude and the total seismic moment of the remotely triggered events with distance from the Landers and Tohoku earthquakes support this (Hill et al. 1993; Miyazawa 2011).

The cumulative number of remotely triggered events exhibits similar decay characteristics as an aftershock sequence (Husen et al. 2004; Pankow et al. 2004; Brodsky 2006). Therefore, a modified (Utsu) Omori-law can be used to fit the data

$$dN/dt = n(t) = k(t + c)^{-p}, \quad (5.1)$$

where $n(t)$ is the number of earthquakes per unit time at time t ; k represents a measure of the productivity of the mainshock and depends on the lower magnitude threshold of the given earthquake catalogue; c describes a temporal offset that compensates for incomplete data in the earliest part of the aftershock sequence, and p describes the exponential decay of the aftershock rate. For all events of $M > 1.5$ at Yellowstone, Husen et al. (2004) obtained $p = 1.02 \pm 0.07$. This result was later on confirmed by Brodsky (2006) who calculated $p = 0.98 \pm 0.07$. Pankow et al. (2004) estimated a value of $p = 0.65 \pm 0.04$ for the sequence of events in Utah. Brodsky (2006) obtained $p = 1.03 \pm 0.03$ for the triggered seismicity at the Geysers and $p = 1.16 \pm 0.17$ for the Long Valley caldera for events of $M > 2$. Stark and Davis (1996) suggested that this decay reflects the redistribution of stress in a similar manner to that which occurs in a mainshock-aftershock sequence.

Brodsky and Prejean (2005) investigated whether dynamic triggering at the Long Val-

ley caldera depends on the amplitude of the surface waves, the cumulative dynamic stress, or the amplitude of the seismic wave at a certain frequency. They found that, above a shaking threshold of 0.2 cm/s, dynamic triggering is independent of the amount of shaking (amplitude), its duration, and the energy-density of the waves. However, the frequency of the wave appears to be crucial for triggering (Brodsky and Prejean 2005). Long-period waves (>30 s) are most effective at triggering local earthquakes at Long Valley caldera. Different periods ranging between 10 and 40 s have been suggested by Anderson et al. (1994) and Prejean et al. (2004) to be responsible for triggering. Since the amplitude decay with depth depends on the period of the seismic wave, the frequency-dependence observed for triggering may vary in different locations depending on the depth at which earthquakes are triggered (e.g. Brodsky and Prejean 2005) .

5.1.4. Suggested mechanisms of earthquake triggering

It is currently assumed that remotely triggered seismicity reflects slip on favourably oriented faults, which fail in response to either temporal fluctuations of the driving shear stress τ , the effective normal stress ($S_n - p$) or short-term variations of the local frictional resistance μ or combinations of these. The temporal reduction of normal stress on the fault plane as the result of particle acceleration in the direction normal to the shear plane by S- or Love-waves polarised in the plane of maximum tectonic shear stress has been suggested by Hill et al. (1993): this amounts to unclamping of the fault caused by oscillations normal to the fault surface. Another mechanism involves crack growth in response to accelerated stress corrosion on the crack tip imposed by the dynamic stresses of the triggering surface waves (Brodsky et al. 2000). Stress corrosion describes the response of a crack to a sudden change in ambient conditions, such as an increase in loading associated with the presence of fluids or high temperatures. The crack tip weakens due to chemical corrosion caused by fluid–rock interactions (Das and Scholz 1981). All other proposed mechanisms (described in more detail below) involve pore fluid pressure changes that lower the effective normal stress on the fault. Hill et al. (1993) suggested that local pore pressure fluctuations occur at periods of 5–20 s in response to compression associated with the Rayleigh wave. Brodsky et al. (2003) proposed a mechanism involving the unclogging of blocked fractures, which describes the rupturing of seals in temporarily blocked fractures and consequent fluid redistribution. This mechanism may explain delays in the occurrence of remotely triggered earthquakes relative to the passing of the surface waves, because fluid flow depends on the velocity with which these fluids can propagate and hence on the permeability. Fluid diffusion typically occurs at velocities of 0.1–1 km/day, depending on the hydraulic diffusivity of the fractured rock (typical range 0.01–100 m²/s) (Steacy et al. 2005, and references therein). Much faster

swarm migration rates of 0.1–1 km/hr observed in the Salton Trough have been explained as triggering by aseismic slip. This has also been suggested by Bodin and Gombert (1994) for faults of the Eastern Californian Shear Zone after the Landers earthquake to explain the delayed occurrence of the Little Skull Mountain earthquake. A similar mechanism has been suggested for the Mexicali Valley, where aseismic slip commenced after the Hector Mine earthquake and microseismicity occurred after a delay of more than 30 hours (Glowacka et al. 2002).

Several mechanisms have been suggested to explain fluid pressure increases in magmatic systems. These processes are associated with bubbles in magma bodies, which increase the compressibility of the magma chamber significantly. Linde et al. (1994) proposed a mechanism termed advective overpressure to explain triggered seismicity at the Long Valley caldera. According to this model, gas bubbles shaken loose due to dynamic stresses imposed by the waves from distant earthquakes rise slowly in the magma and lead to pressurisation. This mechanism would explain a delay in the triggered seismicity due to the rise-time of the bubbles. Rectified diffusion describes the pumping of volatiles into preexisting bubbles that experience a change in shape and volume in response to an oscillating stress field imposed by waves from the triggering event (Sturtevant et al. 1996). Because the surface of the bubble is larger during expansion, more of the dissolved, slightly supersaturated vapour in the fluid/rock/vapour system diffuses into the bubble than out of the bubble, so the bubbles grow and cause a pore pressure increase. Relaxation of any sustained stress differences in a partially crystallised magma chamber, or sinking of crystals shaken loose during the passage of the seismic waves within the magma chamber have also been suggested by Hill et al. (2002) as possible responses to dynamic stresses. Manga and Brodsky (2006) proposed bubble nucleation in a crystallising magma close to critical supersaturation in which the dissolved gas nucleates bubbles in response to small pressure changes associated with the triggering waves.

It has been suggested that several mechanisms may operate at the same time, since most of the proposed mechanisms cause an instantaneous response to the dynamic stresses imposed by the surface waves and only those that involve fluid flow account for a delay between the triggering waves and the remotely triggered events. The table compiled by Prejean et al. (2004) and reproduced in Figure 5.3 summarises the environments in which these mechanisms are most likely to occur and their ability to account for delays. A different mechanism not involving fluid processes but instead involving damage of the contact surfaces of faults caused by the dynamic stresses of the triggering surface waves was described by Parsons (2005). His approach is based on rate and state friction laws.

Possible Mechanisms for Triggering

	Events During Wave Train			Swarms Delayed by Hours to Days		
	A	G	M	A	G	M
Nonlinear friction (e.g., Gomberg <i>et al.</i> , 1998)	X	X	X			
Stress corrosion (e.g., Gomberg, 2001)	X	X	X			
Unclogging fractures (Brodsky <i>et al.</i> , 2003)	?	X	X	?	X	X
Advective overpressure (Linde <i>et al.</i> , 1994)						X
Rectified diffusion* (Brodsky <i>et al.</i> , 1998)		X	X		X	X
Sinking crystal plumes (Hill <i>et al.</i> , 2002)						X
Relaxing magma body (Hill <i>et al.</i> , 2002)						X

Table assumes that triggered earthquakes are not aftershocks of local events. If triggered earthquakes are aftershocks, nonlinear friction and stress corrosion can lead to delayed swarms. A, anywhere; G, geothermal systems; M, magmatically active area. X indicates that triggering can happen in this situation. ? indicates this mechanisms could happen anywhere, but is more effective in geothermal and magmatic systems.

*Note that Ichihara *et al.* (2003) have pointed out theoretical problems with rectified diffusion.

Figure 5.3. Table summarising possible triggering mechanisms as published by Prejean *et al.* (2004)

Rate and state friction laws describe the dependency of fault friction on the rate at which the fault slips and the frictional state of the fault (Dieterich 1994). The state variable is influenced by the slip history (e.g. fault surface roughness, contact area, time-dependent strengthening of frictional contacts, gouge zone thickness, total slip, pore pressure, and normal stress history; see Dieterich 1994). Models based on rate- and state-dependent friction have been successful in describing various fault slip phenomena. Gombert et al. (1998) (“frictional instability model”) and Gombert (2001) (“accelerating failure model”) showed that a transient stress perturbation imposed late in the earthquake cycle causes greater clock-advance than an equal perturbation early in the cycle.

In contrast to the simple Coulomb failure model (in which clock-advanced failure due to dynamic stresses only occurs if the failure threshold is exceeded and in which the system returns to its former state when the transient perturbation is removed), the rate- and state-dependent model is influenced by the transient stress perturbation even after this perturbation has gone. The quantity that characterises failure eventually accelerates towards failure. Parsons (2005) formalised this idea by proposing that damage to fault contacts induced by dynamic stress changes alters the critical slip distance D_c . A reduction in the mean D_c decreases the time to failure (Parsons 2005). Accordingly, dynamic stress changes affect the state of the fault, which causes clock-advance or enhances a delayed response to triggering (Parsons 2005). Parsons’ hypothesis is supported by observations by Li et al. (2003), who showed that damage on the Landers fault increased due to the dynamic waves from the Hector Mine earthquake 25 km away. This shaking-induced damage was noticed as a temporary reversal of the healing process that the fault had undergone within the first seven years following the Landers earthquake. The healing was manifested by the wave velocity increase of fault zone guided waves that were repeatedly generated by explosions.

5.2. Earthquake triggering in the Southern Alps and its implications

As noted in the previous section, remote dynamic triggering of small and intermediate events by large earthquakes has been observed worldwide (Velasco et al. 2008), but particularly in extensional or transtensional tectonic settings (Gombert et al. 2001; Brodsky and Prejean 2005; West et al. 2005). The triggered seismicity starts with, or shortly after, the passage of the surface waves and typically continues for a few hours to several days (e.g. Husker and Brodsky 2004). This process has been termed “waveform triggering” (Brodsky 2006). In several places, however, such as Yellowstone and central Utah, seismicity

persists for weeks after the passage of the surface waves of the triggering earthquake (Hill et al. 1993; Husen et al. 2004; Pankow et al. 2004). This may be related to the stressing rate and the background seismicity as predicted from rate and state friction theory (Dieterich 1994). Sequences that begin several hours (Hill et al. 1993; Gomberg and Bodin 1994; Bodin and Gomberg 1994; Gomberg et al. 2001) to several days (Husen et al. 2004; Pankow et al. 2004) after the surface waves of the triggering event have been referred to as “delayed triggering” (see Table 5.1 for a summary of these observations). These have been observed both in combination with and without initial “waveform-triggered” events. Two mechanisms have been proposed to explain the delay in non-magmatic systems (Table 5.3): unclogging of blocked fractures (Brodsky et al. 2003) and aseismic slip (Bodin and Gomberg 1994, e.g.). Brodsky (2006) showed for the Long Valley caldera that the triggered events following the Landers earthquake resembled aftershock sequences of M_L 4.1 earthquakes. The majority of the triggered events could be explained as aftershocks of those events triggered by the surface waves. However, this explanation is insufficient for isolated, “delayed-triggering” sequences of events.

Many authors have proposed that dynamically triggered events occur in areas where swarms occur regularly on critically stressed faults that experience clock-advance in response to the dynamic stresses imposed by the surface waves of the triggering earthquake (e.g. Gomberg 2001). If this were the case, we would expect similarities in the spatial and temporal characteristics between the triggered and background swarms. In the central Southern Alps region, we regularly observe swarms in an area of abundant seismicity. As described in Section 5.4 below, each swarm consists of a cascade of events of small magnitudes ($M_L \leq 2.8$) with similar waveforms and focal mechanisms. Remotely triggered seismicity occurred in the same area after the 15 July 2009, M_W 7.8 Dusky Sound and the 3 September 2010, M_W 7.1 Darfield earthquake. We observe both “waveform triggering” of earthquakes and several instances of “delayed triggering” swarms. We compare those remotely triggered swarms to the background swarms in terms of waveform similarities, depths and magnitude distribution, duration, inter-event times and focal mechanisms to address previously described hypotheses about triggered seismicity and to contribute to ongoing discussions about the mechanisms of delayed triggered events and the causes of triggering in general.

5.2.1. Seismicity rates in the study area

The method applied to identify sequences is based on changes in the average earthquake rate (Section 5.3). To demonstrate that the seismicity rates throughout the study area are high, we compare the seismicity rates determined using our station array with those

Delay [hr]	Distance [km]	Triggering event [M _W]	M _{max} M _L (M _S)	Number of events	Location	Reference
32.4	260	7.1, Hector Mine	≤4.1		Cerro Prieto	Gomberg et al. (2001)
≥22	280	7.3, Landers	(5.4)		Nevada	Gomberg and Bodin (1994)
2.8	~500	7.3, Gulf of Aqaba	3.7	21	Dead Sea	Mohamad et al. (2000)
7.5	2700	9.3, Sumatra	≤5.1	~100	Yunnan, China	Lei et al. (2011)
≥14	2700	9.3, Sumatra	≤3.9	~300	Yunnan, China	Lei et al. (2011)
≥96	2700	9.3, Sumatra	≤3.5	~120	Yunnan, China	Lei et al. (2011)
321.6	≥3000	7.9, Denali	≤3.2		central Utah	Pankow et al. (2004)
≥192	3100	7.9, Denali	≤3.2	>20	Yellowstone	Husen et al. (2004)
2.5	3100	7.9, Denali	≤0.9	8	Mt. Rainier	Prejean et al. (2004)
23.5	3450	7.9, Denali	≤3.0	112	Long Valley	Prejean et al. (2004)
≥48	6000	8.1, Solomon Islands			Kilauea	Syracuse et al. (2010)

Table 5.1. Summary of observations of delayed triggered events sorted by distance from the large triggering event. The maximum magnitude $M_{\max}$, the number and the location of the remotely triggered events are listed with the reference to the author reporting them.

Figure 5.4. Study area in the central Southern Alps, showing the station network and the seismicity recorded between November 2009 and April 2010. The rectangles mark sub-regions of high and low seismicity. Focal mechanisms are shown for background swarms (blue), triggered swarms (orange) and mainshocks (green) of mainshock-aftershock sequences as well as the 1984, M_w 6.1 Godley Valley earthquake (grey). All focal mechanism solutions shown are lower hemisphere projections with the coloured areas representing compressional first arrivals (for details see Table K.1). The inset shows the setting of the study area in the central South Island and focal mechanisms of major earthquakes: the 2009 M_w 7.8 Dusky Sound earthquake approximately 350 km southwest and the 2010 M_w 7.1 Darfield and 2011 M_w 6.3 Christchurch earthquakes $\sim$ 180 km east of the study area.

obtained by Reyners (1988) for the period 1975 to 1983 using the similarly dense Lake Pukaki station network (Haines et al. 1979; Reyners 1988). That network was designed to observe any reservoir-induced seismicity associated with the infilling of Lake Pukaki for hydroelectric purposes. During impounding of the lake, the seismicity rate at shallow depth within 15 km of the lake shore increased while the background seismicity decreased. The induced seismicity was concentrated near the lake and occurred along the Ostler Fault Zone, a ~ 50 km-long, north–south-striking, west-dipping, reverse fault (Ghisetti et al. 2007). A M_L 4.6 earthquake that occurred near the lake during the infilling was inferred by Reyners (1988) to have been triggered by pore pressure changes associated with the rise in groundwater level. After the final water level had been reached, the seismicity spread outwards away from the lake, possibly reflecting the diffusion of a pore pressure front.

In order to compare the seismicity rates, we only consider events above the magnitude of catalogue completeness of the Pukaki array (M_L 1.8, as reported by Reyners 1988). The rates are calculated for areas within each station network only. Since a seismicity rate decrease was observed in the background seismicity during the infilling of Lake Pukaki and a slight increase was observed thereafter, we use the rate of 6.09 events per 30 days prior to 1978, when the system was completely undisturbed (Reyners 1988). This results in a seismicity rate of 1.52 events per 30 days per 1000 km². In comparison, we recorded $N(M \geq 1.8) = 180$ events within an area of 3300 km² (covered by the SAMBA array) between November 2008 and April 2010, equalling to 3.05 events per 30 days per 1000 km², a rate twice that observed by Reyners (1988).

The rate of seismicity and the magnitude of catalogue completeness both vary markedly throughout the study area. A cut-off magnitude of $M_c = 1.0$ best describes the completeness of detection in the whole study area. The westernmost area (magenta rectangle in Fig. 5.4) exhibits seismicity rates of 7.3 events per 30 days per 1000 km² above the cut-off magnitude. In the central region (red rectangle), the seismicity rate is highest with 35.0 events per 30 days per 1000 km². The seismicity is best recorded in this area and the earthquake catalogue is complete to $M_c = 0.55$ (with a seismicity rate of 70.8 events per 30 days per 1000 km² above this cut-off magnitude). This is the area where the majority of swarms are recorded. The easternmost area (blue rectangle) exhibits a uniform earthquake distribution in space and time with rates of 33.4 events per 30 days per 1000 km². The magnitude of completeness in this region is $M_c = 0.85$. b -value approximately 1.

5.3. Research methodology

We apply the CURATE-algorithm developed by Jacobs et al. (submitted) to identify sequences in the SAMBA catalogue. We use the same definition for an earthquake sequence as Jacobs et al. (submitted); any group of earthquakes with a rate above the background rate that are close in space. The method uses a CUSUM technique (Page 1954) to identify deviations from the background rate. The CUSUM is the “reduced” cumulative sum of the rate change above the cut-off magnitude from the mean rate. Fig. 5.5 (a) shows periods of increased rate as periods of positive slope on the CURATE plot. The earthquakes occurring during periods of increased rates are then limited spatially from the mean location of events during that time. Separate sequences that occur within a designated number of days inside the defined spatial limit are amalgamated to allow for a temporary lull in the occurrence of earthquakes. Jacobs et al. (submitted) provide full details of the method. For our catalogue a number of spatial and temporal values were tested of which a radius of 2.5 km and a 3 day period best separated individual sequences.

We distinguish between swarms and mainshock-aftershock sequences by considering the magnitude difference between the largest and the second-largest events and the time of occurrence of the largest event. The value of this magnitude separation is often taken to be 1.2 from Båth’s law (Båth 1965), but that value is an average and is typically determined for larger-magnitude sequences than we consider here. We use a lower value of 0.7 to distinguish between the two types of sequences. This threshold lies at the lower end of the broad range (0–3) of observed values for the magnitude differences between the largest and second-largest events in mainshock-aftershock sequences worldwide (Felzer et al. 2002). Since Sherburn (1992) have observed low values of 0.3 for swarms and 0.5–1 for other types between swarm and mainshock-aftershock sequences in the Central Volcanic Region of New Zealand, we consider this value appropriate for the non-volcanic Southern Alps region. If the largest event occurs at the beginning of the sequence ($\leq 20\%$ of the duration), we consider it a mainshock-aftershock sequence. If the magnitude separation is less than 0.7 and/or the largest event occurs later in the sequence we call it a swarm.

During the search for sequences, we apply two different magnitude of completeness thresholds: $M_c=1$, which is applicable to the whole study area (Fig. 5.5 b), and $M_c=0.5$, which is the cut-off magnitude within the centre of the SAMBA array (red triangle in Fig. 5.4) where the majority of the sequences occur.

During the aftershock sequences of the earthquakes which caused triggering, noise

Figure 5.5. (a) Cumulative sum (blue) and CURATE (black) of earthquakes recorded in the centre of the SAMBA network between November 2008 and April 2010. Vertical grey lines mark times of the sequences listed in Table 5.2. (b) Magnitudes of earthquakes recorded in the whole study area versus time. Sequences of events that were identified in (a) are coloured with background swarms in blue, triggered swarms in orange and mainshocks in green. Note that some of the mainshock-aftershock sequences are not in the centre of the SAMBA array and are therefore not shown in (a). The horizontal line marks the cut-off magnitude of $M_c=1$.

levels are particularly high, so that our standard event identification routine based on short-term/long-term averages (Boese et al. 2012) performs poorly. To obtain the same catalogue completeness during such times and to check for remotely triggered events during the arrival of the surface waves from the mainshock, we identify triggered earthquakes as high-frequency events in the spectrogram of station WHYM (Fig. 5.6) in addition to the standard identification method (which shows a significant increase in seismicity during this time). WHYM is one of the two stations closest to the swarm locations and has been recording continuously throughout the whole period SAMBA has been in place. All events that could be clearly identified as earthquakes were picked, even if they had been recorded by only one station. This procedure proved effective and in fact lowered the cut-off magnitude by approximately 0.5 during these periods.

In addition to the primary identification with the CURATE algorithm, earthquakes in the sequences were tested for their waveform similarity using cross-correlation. For this, we cross-correlate the waveforms of all 1993 earthquakes that occurred between November 2008 and April 2010. Cross-correlation windows of 0.25 s and 0.35 s for the P- and S-phase, respectively, were used starting 0.1 s before the actual phase pick. For such short time-windows, we set an upper limit of 0.2 s to the lag time which is large enough to identify mispicked phases, but sufficiently small to prevent correlation with converted phases. We bandpass-filtered the waveforms at 2–10 Hz, 3–15 Hz and 7–30 Hz in order to find the best frequency-band for the microearthquakes whose magnitudes were predominantly in the range $-1 < M_L < 2.5$.

Nakahara (2004) analysed the dependency of P-wave cross-correlation coefficients on frequency for aftershocks of the 1995 Hyogo-Ken Nanbu (Kobe) earthquake (M_W 6.9) and observed high cross-correlation coefficients for frequencies lower than 8 Hz and small coefficients for frequencies above 8 Hz. By filtering the theoretical source-time-functions of two events with different magnitudes, he showed that the dependency of the cross-correlation coefficients on frequency results mainly from the similarity of the two signals in the frequency band 2–4 Hz and the differences seen for 16–32 Hz. Therefore, Nakahara (2004) determined 8 Hz as the predominant frequency for an event with $M=3$ which is the characteristic frequency limit for high versus low cross-correlation coefficients. Since the majority of earthquakes observed in this study are significantly smaller, we choose the 3–15 Hz filter band for further analysis, deliberately retaining the high-frequency content. At the same time, we lowered the minimum correlation coefficient to a threshold of 0.5 at at least five stations (cf. Schaff and Waldhauser 2005). With these settings, we obtained a cross-correlation catalogue consisting of 8136 event pairs with mean cross-correlation

Figure 5.6. (a) Spectrogram, raw and high-pass filtered seismograms of the vertical component of station WHYM showing the $M_w 7.1$ Darfield earthquake of 3 September 2010. Large aftershocks near the mainshock as listed in the GeoNet catalogue are shown by annotated arrows or red dashed lines whereas remotely triggered local earthquakes are shown by arrows indicating high-frequency energy or black dashed lines. Note that the GeoNet catalogue is incomplete after the main earthquake due to clipping, power outages, etc., so that small local events are missed shortly after the mainshock.

Figure 5.6. (b) Spectrogram, raw and high-pass filtered seismograms of the M_L 6.3 Christchurch earthquake of 22 February 2010. Marked events by arrows and red dashed lines are aftershock as in (a).

coefficients of 0.662 and mean absolute lag times of 0.053 s.

The identification of swarms based on high cross-correlation coefficients alone is not suitable for our dataset, since the swarms do not stand out in terms of high cross-correlation coefficients despite their waveform similarity (Fig. 5.7 and 5.8). In general, few events have cross-correlation coefficients above 0.8 and for the majority of all events the values range between 0.5 and 0.75. This may simply reflect the multitude of small active faults in the region producing earthquakes or it could be the result of correlated Gaussian noise (Du et al. 2004) which lowers the cross-correlation coefficient. We inspected all events that are linked by cross-correlation coefficients >0.5 at 5 stations to those events in the sequences identified with the CURATE algorithm and chose the events which clearly belong to one family by visual inspection. In other words, a sequence is identified based on seismicity rate-changes but the individual events in a sequence are selected due to their waveform similarity. The final sequence catalogue is listed in Table 5.2.

We analyse inter-event time patterns of sequences for those earthquakes larger than the completeness magnitude. The average inter-event time (instantaneous rate) is calculated at the time of each earthquake in the sequence. We then plot this running mean inter-event time versus the normalized duration which allows the comparison of temporal patterns between sequences of different durations. This may help us to distinguish between mainshock-aftershock and swarm type sequences.

5.4. Observed earthquake sequences

Larger earthquake sequences were observed in the study area in September 1997 (the “Mt Cook swarm” described by Leitner et al. 2001) and in October 2006 (the “Fox swarm” of O’ Keefe 2008). The “Mt Cook swarm” included two M_L 5 earthquakes followed by at least eight aftershocks. The “Fox swarm” consisted of at least 38 earthquakes of which three exceeded M_L 4. Two-thirds of the “Fox swarm” events occurred within one day. Earthquake relocations indicated a point cluster at 8.6 ± 2.6 km depth (O’ Keefe 2008). Four smaller earthquake swarms in late 2006 were identified near Whymper Hut, in the upper Clyde River, near Whataroa and near Waitaha (O’ Keefe 2008).

Here, we report on 14 sequences comprising ≥ 10 events of similar waveforms (Fig. 5.7 and 5.8). This excludes five sequences from our analysis with fewer similar events because there may be more of these smaller sequences that have not been identified due to detection

Figure 5.8. The triggered earthquake swarm on 18 June 2009 with the same specifications as shown in Fig. 5.7. Note that multiplets occur so close in time that events are overlapping (events 7–8, 10, 24–26). Note that the duration without the last event is 7.9 days. The last event occurred 94.9 days after the swarm and was identified in the same location due to its waveform similarity with the remaining events.

issues. These smaller sequences occur throughout the study area. Of the 14 sequences, all except four (three mainshock-aftershock sequences and one long swarm) occur in the centre of the SAMBA array (Table 5.2 and Fig. 5.9), an area of 10 km×12 km between stations WHYM and LABE. Together, the sequences account for 15.3% of the total number of recorded earthquakes above the cut-off magnitude (M_c 1.0). The magnitudes of the sequence events range between $0 \leq M_L \leq 3.7$ and cross-correlation coefficients range between 0.32 and 0.8 with a mean of 0.58.

The durations of the recorded “short” swarms range between less than one day to up to 12 days, with inter-event times spanning a few seconds to several hours. We also recorded several “long” swarms that continue for ≥ 30 days. The majority of these swarms occur in the area of highest seismicity rates, where the best azimuthal coverage is obtained with SAMBA and the depths of the earthquakes are well-constrained (error $2.8 \pm 1.8(1\sigma)$ km). Earthquake swarm events occur at mean depths of 5 km (Section 5.5). Focal mechanism solutions of the largest event in each swarm (Fig. 5.4) exhibit predominantly strike-slip mechanisms with a few oblique-reverse and reverse faulting mechanisms. The focal mechanism solutions indicate steeply dipping faults. If we identify the fault plane as the nodal plane whose slip vector forms the smaller angle to the maximum compressive stress direction S_1 , 13 of 17 measurements lie within $\pm 20^\circ$ of the angles, consistent with Andersonian faulting in the prevailing stress-field (Boese et al. 2012). A comparison of these planes with mapped faults is shown in Fig. 5.10. For four solutions, we prefer the other nodal plane as the fault plane due to similarities with the strike of local faults and consistency with spatially close swarms. For these focal mechanisms, however, both nodal planes are near-optimally oriented in an Andersonian sense (as shown by the numbers below the stereonets in Fig. 5.10).

Three of the recorded mainshock-aftershock sequences occur east and northeast of the area of persistent swarm activity (Fig. 5.4). The waveforms of events in these sequences have smaller cross-correlation coefficients, so the focal mechanism of the largest event may not be representative of all events in these sequences. The largest earthquake observed for the mainshock-aftershock sequences (M_L 2.4–4.2) are on average larger than those of the swarms (M_L 1.1–2.8).

5.5. Remotely triggered seismicity

Remotely triggered seismicity has been observed after two major earthquakes in New Zealand at distances >180 km from the SAMBA array. The Dusky Sound earthquake

Start Date	Lat. [°]	Lon. [°]	CURATE algorithm			Cross-correlation		Max. mag.	Type	Triggered
			$M_c \geq 0.5$	≥ 1.0	misid.	total	before	after	$M_c \geq 0.5$	
12.03.2009	-43.494	170.321	9	4	0	16	2	0	14	2.4 MA
24.05.2009	-43.481	170.362	18	7	0	19	0	0	19	2.6 short
15.07.2009	-43.493	170.303	17	11	1	35	0	3	22	2.8 short
16.07.2009	-43.472	170.400				17	0	5	9	1.4 short
18.07.2009	-43.522	170.330	30	11	15	47	2	2	21	1.8 short
26.07.2009	-43.487	170.283	8	-	4	10	2	2	8	2.0 short
29.07.2009	-43.528	170.125	5	4	0	30	0	1	17	2.2 long
14.08.2009	-43.254	170.778	10	9	0	15	1	3	15	4.2 MA
12.09.2009	-43.442	170.331	6	5	0	17	2	6	13	2.0 long
22.09.2009	-43.525	170.275	8	-	0	28	0	1	27	1.6 long
09.11.2009	-43.512	170.700	10	10	0	22	2	2	20	3.7 MA
16.12.2009	-43.496	170.326	16	4	0	24	0	0	20	1.9 short
29.01.2010	-43.493	170.323	9	8	1	11	2	0	11	1.1 short
30.03.2010	-43.497	170.382	7	7	0	13	2	0	13	2.8 short
28.08.2010	-43.508	170.152	12	6	0	11	0	0	10	2.6 MA
28.08.2010	-43.571	170.501	9	-	2	21	0	0	19	2.0 short
03.09.2010	-43.492	170.369	21	8	9	17	0	0	14	1.5 short
04.09.2010	-43.462	170.260	6	-	2	11	0	0	9	1.3 short
										x

Table 5.2. List of earthquake sequences in the central Southern Alps region identified using the CURATE algorithm and the cross-correlation method. For the CURATE method, number of identified events above the cut-off magnitudes of $M_c=1.0$ and $M_c=0.5$ are listed, and the number of misidentified events for the lower cut-off magnitude. Large numbers of misidentified events result from temporally and spatially close earthquakes in the triggered seismicity. For the cross-correlation method the total number of events with similar waveforms and those before and after the main swarm are given. A minimum of 10 similar events identified with the cross-correlation method is required to be listed in the sequence catalogue. For each sequence the mean location, the maximum magnitude (from cross-correlation method) and the type are shown with “MA” referring to mainshock-aftershock sequences, “long” (≥ 30 days) and “short” (≤ 12 days) referring to the duration of swarms. For “long” swarms only the largest swarm is listed for the CURATE method. Later events have been identified as smaller sequences. Swarms that were remotely triggered are marked. Note that the triggered swarm on 16 July 2009 has not been identified with the CURATE algorithm due to its low-amplitude events but was identified as a cluster of events when checking the cross-correlations of the preceding and succeeding triggered swarms.

Figure 5.9. (a) Seismicity versus time in the centre of the SAMBA array (red rectangle in Fig. 5.4) where the majority of the swarms occur. (b-c) Hourly distributions of events after the $M_W 7.8$ Dusky Sound earthquake and the $M_W 7.1$ Darfield earthquake that caused triggered microseismicity in the central Southern Alps

Figure 5.10. Preferred fault planes for swarm events in the central Southern Alps in map-view (left). By determining the angle between the slip vector and the direction of the maximum compressive stress S_1 (direction and plunge shown) for both nodal planes, we identify the fault plane (red) as the plane which is more consistent with Andersonian faulting. For those events not consistent with neighbouring events or fault orientations in the region, we adopt the other nodal plane (blue). The number indicates the angle between the slip vector and S_1 which shows that for these events both nodal planes are near optimally oriented in an Andersonian sense. Faults are adapted from Cox and Barrell (2007). Panels on the right show the dip (top) and the angle to S_1 of all nodal planes (blue) and the preferred fault plane (red). The grey boxes illustrate angles between the slip vector and S_1 that range within 20° of the direction of S_1 .

(Table 5.3) occurred approximately 350 km south of the study area and ruptured the subduction zone interface between the subducting Australian plate and the overlying Pacific plate (Beavan et al. 2010b). It caused about 2.6 cm of permanent westward coseismic displacement at Haast (Mahesh et al. 2011) approximately 85 km southeast of the centre of the SAMBA array. This large earthquake stands out because of its waves' large low-frequency (0.01–0.1 Hz) and relatively small high-frequency contents (>5 Hz; Fry et al. 2010). The Dusky Sound earthquake caused a maximum acceleration of 0.003 g on the vertical and horizontal components of station FRAN (a borehole sensor at a depth of 100 m in the vicinity of Franz Josef). The peak ground acceleration (PGA) was measured for 1 s period waves (Table 5.4). Similar values were observed on strong motion accelerometers in the area (Table 5.5). Since absolute PGA values are influenced by the local geology at the site and other factors (signal period, source dimensions of the earthquake, radiation characteristics, etc.), we also report the peak ground velocity (PGV; Table 5.4), which depends more on low-frequency components of ground motion than PGA (Boatwright et al. 2001, and references therein) and is therefore a better measure than PGA of the intensity of ground shaking (Tso et al. 1992). The peak dynamic stress σ_d can be calculated from the peak ground velocity u_d using the relationship

$$\sigma_d \sim \mu_S u_d / v_S, \quad (5.2)$$

where $\mu_S = 3 \times 10^{10}$ Pa is the shear modulus and v_S the phase velocity (Jaeger and Cook 1979).

Within the first 24 hours after the Dusky Sound earthquake, 146 microearthquakes occurred in the central Southern Alps region, of which 93.2% could be located (recorded by ≥ 3 stations; Fig. 5.9). This is the highest number of events per day recorded since the installation of the SAMBA array. The triggered seismicity commenced shortly after the passage of the surface-wave trains and continued for approximately 120 hours. The earthquakes were of small amplitude and could be identified best as high-frequency energy on spectrograms of station WHYM. The triggered seismicity comprised three “delayed-triggered” swarms (orange mechanisms in Fig. 5.4) in the centre of the SAMBA array. The first swarm with events of slightly larger magnitude ($M_L \leq 2.8$) occurred eight hours after the passage of the surface waves of the Dusky Sound earthquake. This swarm continued for several days, with a second burst occurring 4 days later in the same location (Fig. 5.9). The second swarm started 20 hr after the Dusky Sound earthquake, 8 km northeast of the first swarm. A third swarm with the largest number of events started 71.6 hr after the passage of the surface waves and lasted slightly over 4.5 hr (Fig. 5.11). The centre of the third earthquake swarm lies 3.7 km SE and 7.7 km SW of the first and second swarms, respectively. On the whole, the microseismic activity lasted five days, and was followed by three days of quiescence

(Fig. 5.9 b).

Elevated seismicity rates in the study area were also observed after the $M_W 7.1$ Darfield earthquake, which occurred on a previously unknown fault in the Canterbury Plains, 175 km due east of the centre of the SAMBA array (Gledhill et al. 2011). This earthquake caused similar strong shaking at station FRAN as the Dusky Sound earthquake had with PGA and PGV values of 0.006 g and 2 cm/s on the horizontal components for waves with periods of 0.6 s and 3 s, respectively. The triggered events were not as well recorded as after the Dusky Sound earthquake, because three out of ten SAMBA stations were not operating at the time following a particularly harsh and snowy southern hemisphere winter. For this reason, most of the triggered seismicity was recorded on only one or two stations and only 53.4% of the events could be located. Figure 5.6 (a) shows triggered events (marked by arrows) as high-frequency energy hidden in the (lower-frequency) signal of the aftershocks. Two remotely triggered swarms of more than 10 similar events could be identified in the area where swarms occurred previously.

A long sequence of events in the vicinity of the 1984 Godley Valley earthquake that began before the Darfield earthquake and continued for six days after, was also studied. This sequence had only three events in the first three days after the Darfield event. This may be due to the detection being affected by the elevated noise levels during the aftershock sequence. In general, triggered and background seismicity during this time appear more distributed and are not as well-recorded as after the Dusky Sound earthquake.

The large, energetic $M_L 6.3$ ($M_c 6.7$) Christchurch earthquake on 21 February 2011 (Holden 2011; Beavan et al. 2011, and references therein) did not trigger significant seismicity in the central Southern Alps. There is no hint of high-frequency events after the Christchurch event or within two subsequent days, as shown in Fig. 5.6 (b). This event caused comparable PGA but lower PGV values than the previous two South Island earthquakes. No remotely triggered seismicity was observed after other large distant earthquakes in 2009 and 2010 as listed in Table 5.3. All of these events caused significantly less shaking in the central Southern Alps (Table 5.4) than the large South Island earthquakes.

We test if the geometry and delay of the triggered swarms after the Dusky Sound earthquake can be explained by the expansion of a triggering front of high pore-fluid pressure. The expansion radius r can be determined from $r = \sqrt{4\pi Dt}$ where D is the hydraulic diffusivity and t is time (Shapiro et al. 1997). Reyners (1988) inferred a hydraulic diffusivity of

Triggered seismicity	Earthquake name	Date	Time (UTC)	M _W	Lat [°]	Lon [°]	Depth [km]	Dist [km]
x	Dusky Sound	15 July 2009	09:22	7.8	-45.75	166.58	12	350
x	Darfield	3 September 2010	16:35	7.1	-43.55	172.18	10	180
–	Christchurch	21 February 2011	23:51	6.3	-43.6	172.71	5	180
–	Vanuatu*	7 October 2009	22:03	7.6	-13.05	166.18	35	3410
–	Fiji	9 November 2009	10:44	7.2	-17.21	178.41	585.1	3020
–	Solomon Islands	3 January 2010	22:36	7.2	-8.91	157.31	31	4050
–	Haiti	12 January 2010	21:53	7.0	18.45	-72.44	10	13570
–	Ryuku Islands (Japan)	26 February 2010	20:31	7.0	25.90	128.42	22	8830
–	Chile	27 February 2010	06:34	8.8	-35.85	-72.72	35	9125
–	Baja (Mexico)	4 April 2010	22:40	7.2	32.13	115.303	10	10110
–	Sumatra	6 April 2010	22:15	7.8	2.36	97.13	31	8860
–	Sumatra	9 May 2010	05:59	7.2	3.77	96.06	45	9055
–	Tohoku (Japan)	11 March 2011	05:46	9.0	38.32	142.37	24	9530

Table 5.3. Large New Zealand and distant earthquakes recorded by the SAMBA network. Dist refers to the distance from the earthquake epicentre to the centre of the SAMBA array. Note that three earthquakes ($M_W > 7.3$) occurred near Vanuatu on 7 October 2009, with only the first event listed here.

Trigg. Seis.	Earthquake	Site Code	Peak Ground Acceleration [g]			Period [s]	Peak Ground Velocity [m/s]			Period [s]	Dyn. stress [MPa]
			V	H1	H2		V	H1	H2		
x	Dusky	FOZ	0.004	0.004	0.008	~1	0.009	0.008	0.008	~9	0.08
	Sound	FRAN	0.003	0.003	0.001	~1	0.007	0.008	0.010	~14	0.09
x	Darfield	FRAN	0.005	0.006	0.006	~0.6	0.012	0.010	0.020	~3	0.17
–	Christchurch	FRAN	0.001	0.001	0.008	~1	0.003	0.002	0.001	~3	0.03
–	Fiji	FOZ	3×10^{-5}			~0.2	0.0001	0.0003	0.0002	~10	0.003
		FRAN	1×10^{-5}			~0.6	0.0002	0.0004	0.0001	~26	0.003
–	Chile	FOZ	2×10^{-6}	1×10^{-6}	1×10^{-6}	~3.6	0.0001	0.0001	0.0001	~18	0.001
–	Tohoku (Japan)	FOZ	7×10^{-6}	1×10^{-5}	8×10^{-6}	~17	0.0002	0.0004	0.0002	~22	0.003
		FRAN	5×10^{-6}	1×10^{-5}	3×10^{-6}	~17	0.0002	0.0004	0.0001	~26	0.003

Table 5.4. Observed PGA (relative to 9.81 m/s^2) and PGV at sites of the SAMBA (FRAN) and GeoNet networks (FOZ). V, H1 and H2 mark the vertical and the two horizontal channels of these sensors. The peak dynamic stress was calculated according to equation 5.2 for the maximum PGV value observed on the three channels.

Figure 5.11. Background seismicity (black), swarm events (blue) and triggered swarm events (orange) in the centre of the SAMBA array in map view and on a cross-section along the horizontal (X-) axis (left). Circles and crosses indicate the mean locations of observed triggered swarms after the Dusky Sound and Darfield earthquake, respectively. Dashed circles reflect distances that fluids could propagate by diffusion. The red line marks the border of a common fluid source from which fluid diffusion would be consistent with the triggered swarms. Faults shown are from the QMAP Aoraki 1:250000 by Cox and Barrell (2007). The figures on the right show the increase of the differential stress and the pore-fluid pressure with depth (top) and the depth distributions (bottom) of all events without either triggered or non-triggered swarms (black), the swarms (blue) and the triggered seismicity only (orange). The average and the standard deviation of each are shown, too.

Distance to Source [km]	Peak Ground Acceleration (g)			Site Code	Name of Recording Site	Subsoil Category
	V	H1	H2			
302	0.005	0.005	0.005	FOZ	Fox Glacier	B
305	0.004	0.008	0.010	MCNS	Mount Cook Annex*	C–D
312	0.002	0.005	0.004	TKAS	Tekapo*	A–B
319	0.006	0.010	0.011	FGPS	Fox Glacier DOC	D
335	0.007	0.011	0.009	FJDS	Franz Josef DOC	D
375	0.006	0.011	0.011	HAFS	Hari Hari Fire Station	D
391	0.003	0.007	0.006	WVAS	Waitaha Valley	D

Table 5.5. Observed PGA at sites on the eastern (*) and western side of the central Southern Alps after the Dusky Sound earthquake from Fry et al. (2010)

1.5 m²/s from the lag time of groundwater rise associated with the impounding of Lake Pukaki. Using this diffusivity estimate, which may not be strictly appropriate because it was derived for shallow depths on the eastern, dry (leeward) side of the Southern Alps, we determine that a pore-fluid pressure front could migrate 0.7, 1.2 and 2.2 km due to diffusion (for the given delay-times of 8, 20 and 71.6 hr). These values are too small to explain the spatial separation of the swarms. Since we have three equations (3 diffusion radii of the pressure front), we can invert for a common fluid source assuming a hydraulic conductivity and a starting position between the three swarms. We find a circle that touches the three diffusion circles around each triggered swarm after the Dusky Sound earthquake (dashed in Fig. 5.11) which marks the boundary of a fluid source from which a pore pressure front radiates outwards. This boundary is indicated in red in Fig. 5.11 and it is also consistent with the two swarms after the Darfield earthquake for diffusivities of 4.5 m²/s.

5.6. Discussion

5.6.1. Observed earthquake sequences

Our observations show that earthquake swarms are a common feature of the seismicity in the central Southern Alps region where background seismicity rates are high compared to elsewhere in the central South Island. They are best recorded in a small area of dense seismicity in the centre of the SAMBA array. However, small sequences of 6–8 events have been identified further east and northeast of that particular area. The swarm events are of low magnitude ($M_L \leq 2.8$), exhibit similar waveforms and have short durations, usually ranging between a few hours and a few days. However, at least three sequences comprising two phases of activity linked by a few events have been recorded. This may either indicate

short recurrence times (a few weeks) between individual repeating swarms or that the process causing the swarms can involve several phases of activity. The depth distribution of the background swarms overall reflects the distribution of all events in this area (Fig. 5.11). The absolute location uncertainty is too large to resolve any systematic changes that might indicate a shallowing of the hypocentre depths with time. We have analysed the relative travel-times at station WHYM for the swarms comprising the largest number of events, but we cannot see any systematic changes. This may still be a resolution problem due to the 2 km difference between the deep and shallow event depths.

Our analysis of the inter-event times of events in the swarms and mainshock-aftershock sequences shows that these processes are distinctly different at the beginning of the sequence. The average inter-event time between earthquakes in a swarm decreases steadily until a turning point is reached from which point on the inter-event time increases steadily (Fig. 5.12 (a–c)). This pattern differs from the temporal distribution of events in a mainshock-aftershock sequence in which case a systematic increase in inter-event times is observed (Fig. 5.12 (f–h)). The latter is consistent with the Omori law of aftershock decay. Several swarms show that the largest events occur in the transition from decreasing to increasing inter-event times. This may indicate that the maximum stresses are released before the start of a decay process, marked by the increase in inter-event times.

These patterns suggest that the underlying processes of the swarms and mainshock-aftershock sequences are distinct and different, although the decay portions (increasing inter-event times) may be similar. These patterns do not depend on the cut-off magnitude because theoretically the same number of small magnitude events are taken out at the beginning, middle and end of the sequence if the cut-off magnitude is increased and the overall pattern remains unaffected. We also note that not all swarms strictly follow this pattern or exhibit it as clearly as the examples shown. This may be due to the small numbers of events above the cut-off magnitude or the fact that multiple mechanisms exist that can cause an earthquake swarm. A continuum of types between swarms and mainshock-aftershock sequences (Sherburn 1992; Vidale and Shearer 2006) rather than discrete types (Sykes 1970) has been observed.

We obtain at least one focal mechanism solution for each swarm, most often from the largest event. If multiple focal mechanism solutions exist as shown in Fig. 5.13, these are identical within their uncertainty limits (cf. Boese et al. 2012) and are similar to those of other events in the vicinity. All solutions except one shown in Table 5.6 have one steeply-dipping nodal plane ($\geq 50^\circ$). This is consistent with the findings of Vidale and Shearer

(2006), who reported that swarms predominantly occur on vertical planes and form linear streaks. They suggested that swarms are preferentially produced on strike-slip faults, and that observed vertical hypocentre alignments could reflect density-driven fluid flow.

Figure 5.13. Focal mechanisms of all events within 0.05° of the location of the swarm starting on the 22 September 2009. Note the slight difference for events in July which do not belong to the swarm and those during the September/October sequence. The number below the date indicates the number of P-wave first motions. The size reflects the magnitude of the earthquake.

It is commonly assumed that swarms occur on fault planes that are favourably oriented in the prevailing stress field according to Andersonian faulting theory. We have tested this idea by determining the angle between the slip vectors of both nodal planes and the direction of maximum compressive stress S_1 determined previously (Boese et al. 2012). Of the two resulting angles, at least one complies with angles of $\pm 30^\circ$ or $\pm 150^\circ$ from S_1 as expected from Andersonian faulting theory, confirming that the low-magnitude earthquakes occur in response to the regional stress field on near-optimally oriented faults.

Date	Time (UTC)	Strike [°]	Dip [°]	Rake [°]	Strike [°]	Dip [°]	Rake [°]	Mag. M_L	Angles to S_1 [°]
16.03.2009	17:32:25.7	12.5	72.2	11.4	278.9	79.1	161.8	2.4	101.29
24.05.2009	5:17:59.6	16	55.7	7.9	281.5	83.5	145.4	2.35	100.59
15.07.2009	19:25:11.3	331.8	80.8	-2	62.1	88.1	-170.8	1.96	139.81
16.07.2009	14:49:55.2	325.8	74.4	-21.6	61.8	69.3	-163.3	1.44	145.81
18.07.2009	9:45:01.8	63.6	64.1	-168.1	328.4	79.3	-26.4	1.65	124.59
29.07.2009	11:41:43.4	240.2	68.6	179.1	330.6	89.2	21.4	2.28	56.775
14.08.2009	16:24:05.8	152.2	38.5	-28.7	265.4	72.6	-124.9	4.44	69.812
12.9.2009	5:38:12.5	144.5	85	1.2	54.4	88.8	175	2.13	34.1
26.09.2009	13:02:41.1	219.2	69	131.3	331.4	45.5	30.2	0.94	50.642
09.11.2009	13:59:05.0	60.8	76.4	164.1	154.7	74.5	14.1	3.91	118.9
17.12.2009	4:00:15.6	329.2	74	-10.5	62.1	79.9	-163.7	1.86	143.14
30.01.2010	2:42:14.0	249.6	66.4	85.8	79.9	24	99.4	1.22	57.762
03.04.2010	7:22:49.6	15.4	18.5	108.6	175.9	72.5	83.9	3.02	132.67
28.08.2010	15:19:12.2	348.6	44.6	-66.8	137.6	49.8	-111.2	2.6	87.055
31.08.2010	8:49:20.6	12.4	17.8	149.1	132.1	81	74.6	1.72	102.38
04.09.2010	20:11:25.4	203.1	67.9	18.5	105.9	72.9	156.8	1.22	75.696
04.09.2010	23:51:26.2	214	47.4	-69.3	4.8	46.5	-111	1.47	150.89
									63.122

Table 5.6. Focal mechanism parameters of events shown in Fig. 5.4 and the calculated angles between the slip vector and the maximum compressive stress direction S_1 as shown in Fig. 5.10.

5.6.2. Remotely triggered seismicity

The remotely triggered swarms do not differ significantly from the background swarms in terms of their magnitude and depth distributions, focal mechanisms or inter-event time patterns except that there are more swarms within a shorter timespan. A systematic depth difference exists between triggered and non-triggered background swarms in comparison to the overall background seismicity. The triggered and non-triggered swarms have larger depths (a mean with a standard deviation of the mean of 4.9 ± 0.10 km and 5.0 ± 0.70 km, respectively; Fig. 5.11) compared to the average depth of 3.4 ± 0.13 km of all events. This difference in depth is statistically significant and may be related to the minimum differential stress required for failure. Finite element modelling results for the central Alpine Fault obtained by Ellis et al. (2006) show that the differential stress is highest at 6–7.5 km depth depending on the initial temperature model used. Assuming a hydrostatic pore-fluid factor, a Byerlee coefficient of internal friction of 0.6, a stress ratio $R=0.5$ and an intermediate principal stress due to overburden, $S_2 = \rho gz$, the differential stress $(S_1 - S_3)/2$ increases by 16.7 MPa per kilometre depth (Section 4.4 and Fig. 4.9). Consequently, a difference in depth of a few kilometres can reflect significant stress changes in terms of triggering thresholds (the stress difference for 3.4 km versus 4.9 km is shown in Fig. 5.11). According to Upton et al. (1995), meteoric waters penetrate to depths of 5–6 km in the central Southern Alps. If a fracture system extends to these depths as the background seismicity suggests, a pore-pressure gradient of 15 MPa per kilometre depth exists (assuming hydrostatic fluid pressure). Unclogging of blocked fractures (Brodsky et al. 2003) or an increase of damage on the fault's contact surfaces (Parsons 2005) in response to the dynamic stresses imposed by the surface waves of the triggering event have been suggested as plausible triggering mechanisms. If a fracture connection is established, rapid fluid redistribution would occur while the pore-pressure difference is large and decay away as the pressure difference decreases. This would fit our inter-event time observations for the background and triggered swarms. According to this interpretation, the time interval between swarms represents the time required for the pressure to build up once the seal is reestablished.

The first triggered swarm after the Dusky Sound earthquake was the most energetic in terms of moment release ($M_L < 2.8$) whereas the second and third triggered swarms did not exceed M_L 2. We did not calculate Coulomb stress changes to investigate if one swarm triggered the next swarm because static stress changes are insignificant at the distances between the remotely triggered swarms. Static stress changes are expected to affect only the area within several source dimensions, which is on the order of a few hundred metres for small earthquakes (e.g. Sibson 1989). The first and second triggered swarms are spatially

separated by ~ 8 km, the third swarm occurs at approximately the same distance from the second and 3.7 km from the first. The geometry and delay between the swarms and the Dusky Sound earthquake are consistent with the expansion of a pore-fluid pressure front for diffusivities of $4.5 \text{ m}^2/\text{s}$ following the arrival of the surface waves. The same fluid source is also consistent with swarms after the Darfield earthquake as shown in Fig. 5.11.

Parotidis et al. (2003) showed that the spatio-temporal patterns of swarms in Vogtland/NW-Bohemia in 2000 were consistent with fluid diffusion from a fluid source of magmatic origin. They concluded that the swarm pattern depends on the hydraulic diffusivity and on the criticality (the critical pore-pressure value leading to failure) in the area. High fluid-pressure has been identified in the Southern Alps region as the cause of reduced seismic wave-speeds seen in wide-angle reflection and teleseismic P-wave arrival data at 6–8 km depth (Stern et al. 2001, 2002; Stern et al. 2007) on SIGHT transect 2 approximately 30 km to the south of the swarm area. A zone of low resistivities indicating interconnected fluids at depth coincides with the low-velocity zone (Wannamaker et al. 2002).

The triggered swarms and the majority of all other swarms occur in one area within the SAMBA network which seems most prone to a seismic response to small stress changes. As no remotely triggered seismicity was observed following large distant earthquakes and after the $M_W=6.3$ Christchurch earthquake (21 February 2011), this suggests that there may be a stress threshold, amongst other factors, that has to be exceeded to produce a fluid response. The distant events listed in Table 5.4 caused PGV of 0.01 cm/s corresponding to peak dynamic stresses of $\sim 9 \times 10^{-4}$ MPa. The Christchurch earthquake caused comparable ground accelerations to the Dusky Sound earthquake but slightly lower PGV values (0.3 cm/s corresponding to peak dynamic stresses of 0.03 MPa). However, the maximum PGV of the Christchurch earthquake was recorded on the vertical component whereas the other large earthquakes caused extreme values on one of the horizontal components. The ground motion of both triggering earthquakes, the Dusky Sound and the Darfield earthquake, locally exceeded PGV and PGA values of 0.7 cm/s and 0.001 g, respectively. These values correspond to peak dynamic stresses of ≥ 0.09 MPa. Observed peak dynamic stresses at Mammoth Mountain (Long Valley caldera) after the Denali, Hector Mine and Landers earthquakes were 0.06, 0.25 and 0.3 MPa, respectively (Prejean et al. 2004). Similar values of 0.01, 0.1 and 0.5 MPa were observed at the Coso geothermal field (Prejean et al. 2004). In Yellowstone those stresses reached 0.2 MPa (Husen et al. 2004), 0.12–0.35 MPa in Utah (Pankow et al. 2004), 0.09 MPa at Mt. Rainier in central Washington (Prejean et al. 2004), 0.07 MPa at the Geysers (all after the Denali earthquake; Prejean et al. 2004), and 0.18 MPa in Greece (after the Izmit earthquake; Brodsky et al. 2000). Peng et al. (2010) estimated

peak dynamic stress of 0.085–0.17 MPa at locations in north China after the Wenchuan earthquake.

The upper limit of static stress changes in the central Southern Alps at a distance of 350 km from the Dusky Sound earthquake (for a seismic moment of 2.72×10^{20} N m according to Fry et al. 2010) is $\Delta\sigma = 504.8$ Pa = 0.5 kPa. The same value results for the Darfield earthquake at a distance of 180 km (moment release of 3.5×10^{19} N m from USGS, $\Delta\sigma = 477.6$ Pa). Since these values are smaller than the dynamic stresses imposed by all other large distant earthquakes, we consider the static stresses to be insignificant. Regarding the frequency content of the triggering waves, the seismicity response to the Dusky Sound earthquake in the central Southern Alps was more intense (more earthquakes) than for the Darfield event, despite the larger PGA and PGV caused by the Darfield earthquake. The Dusky Sound earthquake generated more low-frequency energy than the Darfield event, which is clearly reflected in the periods of maximum ground velocities (Table 5.4). The maximum PGV values for both the Darfield and the Christchurch earthquakes were measured for waves with similar frequencies. Brodsky and Prejean (2005) showed that above a shaking threshold of 0.2 cm/s dynamic triggering is independent of the amount of shaking (amplitude), its duration and the energy-density of the waves but it is dependent on the frequency of the wave. Waves that caused remote dynamic triggering exhibited periods ranging between 10 and 40 s (Anderson et al. 1994; Prejean et al. 2004; Brodsky and Prejean 2005). In this respect, the high stress-drop Darfield earthquake (Quigley et al. 2012) is unusual because of the dominant periods of 3 s. The GeoNet stations in the vicinity of the SAMBA array are all clipped for longer periods and our sensor at FRAN is a 4.5 Hz short-period seismometer so the long-period waves may not be well-recorded.

Our results suggest a stress threshold for triggering of $0.03 \text{ MPa} < \sigma_{\text{thres}} < 0.07 \text{ MPa}$ at the 100 m-deep borehole station FRAN. Gomberg et al. (2001) showed that the peak dynamic stress value depends on the local amplification at the site (since it is derived from the maximum amplitude of the velocity seismogram). Sites which recorded the Landers and Hector Mine earthquakes and exhibited remotely triggered seismicity after one of these events always showed larger PGV values for the triggering mainshock compared to the non-triggering one. However, the comparison of these PGV values between different stations showed a wide range of absolute values indicating site-dependence of the absolute value that has to be exceeded at any locality.

Based on the number of recorded sequences above the cut-off magnitude and the average time between these sequences, the probability of recording one sequence (or more) in

any given five day time period is 12.7%. The probability of observing two swarms or more with at least six events above the cut-off magnitude $M_c=1$ (Table 5.2) within five days after the Dusky Sound earthquake, but unrelated to it, is 0.8%.

We can fit the change in rate after the Dusky Sound earthquake by a modified Omori-law as shown in Fig. 5.14. Assuming $p=1$ in equation 5.1 (Utsu et al. 1995) and $c=10$, we obtain $k=K/\ln 10=10.8$ for a cut-off magnitude of 0.55 in the swarm area.

Figure 5.14. Cumulative sum of events versus Julian day since 1 January 2009 in the centre of the SAMBA network (red rectangle in Fig. 5.4). Dashed lines before and after the Dusky Sound event (on day 195) show the average rate and the average rate plus the Omori-law decay right after the earthquake. The inset shows how values for K and S in $N(t) = S + K \log(1 + t/10) + Rt$ have been obtained from a grid search. R has been obtained from the seismicity rate before the Dusky Sound earthquake.

5.6.3. Long-term record of swarms and triggering of swarms

While many swarms have been observed in the centre of the SAMBA network, only a few small swarms were recorded in the rest of the study area. We analysed the seismicity in the GeoNet catalogue between 1993 and 2008 for different subregions throughout the study area (e.g. the areas of the Mt. Cook and Fox swarm). This reveals several clusters in the area where we have identified many swarms and another sequence in the Fox area in late 2001

(in addition to the Fox swarm; O' Keefe 2008). Closer examination of our swarm catalogue with current search parameters reveals five clustered events in the Fox area (magenta) in January 2010.

In a different approach, we analyse the PGA recorded at Haast (approximately 85 km southwest of the array) for large earthquakes that caused similar amounts of shaking as those that caused the observed triggered seismicity. Candidate earthquakes for possible triggering are listed in Table 5.7. Note that a station upgrade of the GeoNet stations was undertaken in 2003 so apparent changes in the catalogue around this time are an artefact of the network upgrade. There may be a causal relationship between both the 1993 Secretary Island earthquake and 2003 Fiordland earthquake and sequences comprising large events three months later (Fig. 5.15). That is, these sequences may have experienced clock-advance due to the large earthquakes. There is no indication of remotely triggered seismicity commencing immediately after the large earthquake, but the triggered swarms recorded in 2009 and 2010 are not listed in the GeoNet catalogue because of their small magnitudes. A catalogue with a lower magnitude of completeness is required to resolve this issue.

Date	M _w	Earthquake
10 August 1993	6.8	Secretary Island
18 June 1994	6.7	Arthur's Pass
21 August 2003	7.2	Fiordland
21 August 2003	6.2	Fiordland aftershock
02 November 2003	6.4	Fiordland aftershock
13 December 2004	8.1	Macquarie Island
22 November 2004	7.1	Puysegur Trench
15 October 2007	6.7	George Sound
15 October 2007	6.2	George Sound aftershock
19 December 2007	6.8	Gisborne
15 July 2009	7.8	Dusky Sound
04 September 2010	7.1	Darfield
21 February 2011	6.3	Christchurch earthquake
13 June 2011	6.0	Christchurch aftershock
23 December 2011	6.0	Christchurch aftershock

Table 5.7. Major earthquakes that caused comparable amounts of shaking at Haast as the 2009 Dusky Sound and 2010 Darfield earthquakes (≥ 1 mm/s), as shown in Fig. 5.15.

Figure 5.15. Search for triggered sequences in the central Southern Alps region after large earthquakes. (a) Peak ground velocities (PGV) from major earthquakes recorded at Haast ca. 85 km southeast of the centre of the SAMBA array (Table 5.7). The dashed line indicates the threshold proposed by Brodsky and Prejean (2005) above which triggering is independent of the waves' amplitude, duration and energy-density but depends on the frequency of the waves. (b-d) Seismicity versus time for the subregions as shown in Fig. 5.4. Black circles indicate sequences detected in this study, annotated arrows show the Mt. Cook (MC) and Fox swarms (FX). The daily seismicity rate is 0.86 (b), 1.66 (c) and 0.45 (d) events per 30 days per 1000 km² for the GeoNet catalogue (M_c 2.6) between 1993–2007. (e) Seismicity versus time for the whole study area.

5.7. Conclusions

The detection of small earthquake swarms strongly depends on the local station network and its detection threshold (e.g. Frankel et al. 1980). With our dense station network in the central Southern Alps, we recorded 16 swarms of low magnitude ($-0.5 < M_L \leq 2.9$) between November 2008 and September 2010. Three swarms have been remotely triggered by the M_W 7.8 Dusky Sound and at least two by the M_W 7.1 Darfield earthquakes. The triggered swarms are similar to the background swarms in terms of waveform similarity, maximum magnitudes, depth distribution, duration and inter-event time patterns, but they comprise more events (which may be a result of the different detection methods of events in the aftershock sequence). The triggered and non-triggered background swarms have larger depths than the background seismicity, which may correspond to the depth of ambient peak differential stresses, according to modelling results. The majority of the swarms occur in the centre of the network and have well-constrained hypocentres. The focal mechanism solutions of the largest event in each swarm indicate either strike-slip, reverse faulting or a combination of both. All mechanisms have one steeply-dipping nodal plane ($\geq 50^\circ$) in common. The focal mechanisms of both the triggered and the non-triggered background swarms are in agreement with near-Andersonian faulting on favourably oriented faults in the stress field. The crust is highly fractured in this region with a wide variety of fault orientations, so potentially there is an abundance of faults near failure.

The observed remotely triggered seismicity started with the passage of the surface waves of the Dusky Sound and Darfield earthquakes and continued for several days. The triggered swarms occurred delayed by 8, 20 and 71.6 hours after the Dusky Sound and 6.4 and 7.3 hours after the Darfield earthquake. Since we observed dynamic triggering twice within two years in the same area, we conclude that earthquake triggering is a common phenomenon in the Southern Alps. However, in the long-term record there is no indication for triggered swarms (probably because the GeoNet catalogue cut-off magnitude of M_c 2.6 according to Eberhart-Phillips 1995; Petersen et al. 2011, is too high).

Based on results presented in this chapter, a triggering threshold of 0.09 MPa has to be exceeded for remotely triggered seismicity to occur in the centre of the SAMBA array. This threshold is comparable to triggering thresholds observed worldwide and does not indicate that larger values are required in compressional settings than in transtensional and extensional regimes.

Inter-event times versus durations of these swarm earthquakes above a cut-off magni-

tude of M_L 1.0 show a characteristic pattern that is markedly different from that of a typical mainshock-aftershock sequence. Both types of sequences are distinct in their initial phase: a swarm exhibits decreasing inter-event times followed by an increase. The events in a mainshock-aftershock sequence only show this increase or decay process. This suggests that different mechanisms exist and that a distinct repeatable process causes the swarms. The initially accelerating failure is consistent with fluid-flow within or into weak localities with an abundance of cracks and faults. Due to the delay of the triggered swarms and existing evidence of high fluid-pressure (Stern et al. 2007) and interconnected fluids at depth (Wannamaker et al. 2002), we favour a fluid-driven mechanism over the aseismic slip model and the order and occurrence of the triggered swarms is consistent with a common fluid source from which a pore pressure front diffuses away.

In general, our observations are consistent with the hypotheses that have been made to explain remotely triggered seismicity, namely:

- areas with frequent occurrence of swarms more often respond to dynamic stress changes than other areas (e.g. Hill et al. 1993);
- triggering involves brittle slip on critically stressed faults favourably oriented in the prevailing stress field, and;
- triggering represents clock-advance of failure on a fault or a number of faults (e.g. Anderson and Webb 1994; Steacy et al. 2005).

The latter is based on the highest seismicity level in the area and the fact that the triggered seismicity is followed by a quiescence period of approximately the same length as the elevated seismicity period. This suggests that many faults near failure were clock-advanced in their seismic cycle. However, in order to establish if the periodicity between swarms was significantly disturbed, we need a longer history of swarm activity in this region which allows for a thorough analysis of the intervals between sequences.

6. Deep earthquakes, tremor and other preliminary results

This chapter summarises unpublished results related to subsidiary topics explored during this research. Aaron Wech has been involved in the search, identification and analysis of tremor as described in Section 6.2. He has determined the tremor locations, uncertainties and magnitudes and written the publication in *Geophysical Research Letters* that summarises this section (Wech et al. 2012).

6.1. Deep earthquakes

Several earthquakes with depths exceeding 30 km have been recorded by the SAMBA and GeoNet networks (Table 6.1 and Fig. 6.1). These events have magnitudes $M_L < 3.5$ and occur mostly west and southwest of the SAMBA station network. The deep earthquakes occur outside the SAMBA network but the majority of the events are still sufficiently close to the network that at least two stations are within distances smaller than the focal depth of the events. All events have at least three S-Picks. The uncertainties of the hypocentres in horizontal and vertical direction as shown in Table 6.1 are small, reflecting that the depths are accurate despite the locations outside the station network. The azimuthal coverage of the stations relative to the earthquakes, however, is not ideal with gaps $\geq 130^\circ$.

For three of these events, more than seven polarity picks from the SAMBA stations are available, enabling focal mechanism solutions to be calculated. Coverage on the focal sphere is sparse, so we have augmented the data of the largest events ($M_L \sim 3$) that were recorded by GeoNet with polarities available from these stations. Three focal mechanisms have been obtained, one using only the available polarity data from the SAMBA and the other two from the combined SAMBA/GeoNet stations. The P-onsets of the deep events are usually impulsive as shown in the examples of Fig. 6.2. The two events listed in the GeoNet catalogue were recorded by stations across the whole South Island with some clear P_n arrivals. The ray take-off angle of P_n phases is uncertain (due to the complicating

Figure 6.1. Subcrustal earthquakes (depth >35 km) reported by Reyners (1987), Kohler and Eberhart-Phillips (2003) and in this study (Table 6.1) recorded by various networks in comparison.

effects of the crustal structure), so ideally they should be assigned larger uncertainties in the focal mechanism analysis. This is currently not possible with the algorithm used here (Walsh et al. 2009) and would require substantial modifications of the code. Therefore, we compare the focal mechanism solutions obtained using these P_n polarities with those excluding P_n (Fig. 6.3). The additional P_n polarity picks are in agreement with the reverse mechanisms obtained excluding P_n . However, due to the large epicentral distances of these stations, they correspond to the rim of the focal sphere and significantly influence the focal mechanism parameters. In conclusion, the focal mechanisms of the three deep events consistently exhibit reverse mechanisms, but the strike, dip and rake have large uncertainties and are poorly constrained. A more extensive station network in the Southern Alps region is needed to better resolve the focal mechanism parameters of such deep, small to moderate earthquakes. Nevertheless, our results are in agreement with a composite focal mechanism obtained by Reyners (1987) for events at depths of ≥ 25 km in this region.

Figure 6.2. Travel-time versus distance plot for a deep earthquake recorded on the vertical component of the stations on 17.10.2009 as listed in Table 6.1.

Figure 6.3. P-wave polarities and focal mechanism solution of the deep events (top 12.12.2008, centre 25.10.2009, bottom 17.10.2009) including (left) and excluding P_n onsets (right, all polarity picks of the GeoNet stations marked). Note that the last event was only recorded by the SAMBA network. Compressional onsets are shown in blue, dilatational ones in red.

Date	Time	Lat. [°]	Lon. [°]	Dep. [km]	RMS [s]	Stn.	Phs.	Gap [°]	Uncertainty [km] [km]	
8.12.08	5:46:20.4	-44.64	168.04	98	0.39	18	25	155	1.1	1.2
12.12.08*	14:11:45.7	-44.05	169.16	64	0.40	21	28	133	0.8	1.5
8.01.09	10:04:25.3	-44.03	169.14	58	0.29	16	21	139	0.8	1.5
26.01.09	0:29:26.1	-42.94	171.10	39	0.28	7	9	142	1.2	2.0
27.01.09	12:29:32.1	-43.69	169.80	55	0.16	6	10	177	2.8	5.6
10.03.09	13:26:50.6	-43.84	169.23	73	0.25	9	17	181	1.3	2.3
23.03.09	17:22:26.3	-44.00	169.02	70	0.26	12	15	165	1.2	3.6
4.04.09	8:41:56.1	-43.94	169.43	74	0.37	19	25	141	1.2	1.6
11.07.09	9:30:26.7	-43.53	169.94	61	0.33	11	18	160	1.8	3.8
22.08.09	10:57:51.3	-43.33	169.88	59	0.23	15	23	201	1.2	1.6
17.10.09*	7:53:20.3	-43.41	169.81	65	0.22	14	26	195	1.3	1.4
25.10.09*	15:02:40.8	-43.60	169.68	61	0.31	28	43	167	0.8	1.0
25.12.09	9:57:34.0	-43.68	169.95	54	0.04	8	12	290	2.1	8.1
17.01.10	22:39:2.0	-43.36	169.79	61	0.17	6	11	224	1.8	2.8

Table 6.1. Subcrustal earthquakes recorded by the SAMBA and GeoNet network. Stars mark those events for which a focal mechanism could be obtained as shown in Fig.6.1 and 6.3. The number of stations and phases are listed, as are the location uncertainties in horizontal and vertical directions.

Two of the three deep events occur at depths in the lithospheric mantle where the Australian and Pacific plates intersect (Stern et al. 2002; Reyners 1987). This may represent a region where compression due to plate bending is largest (cf. Forsyth 1982).

6.2. Tremor

The emergent, low-frequency (2–8 Hz), low-amplitude and long-duration signal (several minutes to hours) observed on the seismograms of stations along the subduction zones of Cascadia, Nankai, Alaska, Mexico and Costa Rica and the strike-slip San Andreas fault has been termed tremor (Beroza and Ide 2011, and references therein). Tremor is inferred to represent slow shear-slip failure (rupture speeds slower than those of ordinary earthquakes) in the transition zone between stick slip and stable sliding of a fault (Shelly et al. 2009) or on the plate interface (Shelly et al. 2006; Wech and Creager 2007). This has been shown by the locations and first motions of low-frequency earthquakes occurring on or near the plate interface (Shelly et al. 2006) and indicate plate-boundary slip (Ide et al. 2007). Low-frequency earthquakes have been identified in parts of the tremor signal (Shelly et al. 2006, 2007) and it has been suggested that tremor results from a superposition of these events (Shelly et al.

2006; Beroza and Ide 2011, and references therein). Tremor has also been attributed to fluid-related processes at depth (Obara 2002) and high fluid pressure has been inferred near tremor sources in Japan, Cascadia and Mexico (Shelly et al. 2006; Audet et al. 2009; Song et al. 2009). It is not clear, however, if fluid movement is responsible for tremor and aseismic slip or if aseismic slip causes fluid redistribution which generates tremor (Shelly et al. 2006).

A clear correlation of tremor in subduction zones of Cascadia and Japan with slow-slip events has been demonstrated by Rogers and Dragert (2003) and Obara et al. (2004) where both tremor and slow-slip exhibit remarkable periodicity. In strike-slip settings, however, no deformation signal has yet been observed in association with tremor (Johnston et al. 2006). Tremor observed on the San Andreas fault extends along a 150 km-long segment from the end of the locked segment near Cholame, through the transitional zone southeast of Parkfield, as far as Bitterwater near the creeping section northwest of Parkfield (Shelly et al. 2009; Peng et al. 2009). The tremor signal is shorter (4–20 min), less frequent (≤ 5 events detected in any 24-hour period) and has smaller peak amplitudes ($M_W < 0.5$ for equivalent earthquakes) than in subduction zones (Nadeau and Dolenc 2005). The upper and lower depth-ranges of families of low-frequency events are 18 and 28 km (Shelly and Hardebeck 2010) with depth errors of individual events on the order of 3–5 km (Shelly et al. 2009). Tremor migrates along the fault at velocities of 15–40 km/hr (Shelly et al. 2009), and the largest amplitudes of the tremor signal (30–55 nm/s on the velocity seismogram) occur at the end of the locked zone near Cholame (Shelly and Hardebeck 2010).

Several episodes of triggered tremor have been observed along the San Andreas Fault following distant (Peng et al. 2009) and some regional events (Guilhem et al. 2010). Tremor responds to much lower stress changes than triggered earthquakes, especially if the incidence direction of the surface waves of the triggering earthquake is in a direction similar to the strike of the fault (Peng et al. 2009; Hill 2008). Tidal stress changes seem to be sufficient to trigger tremor since a correlation of tremor with the 12.4 and 25 hr tidal periods has been observed in several places (e.g. Rubinstein et al. 2008).

The data recorded by the SAMBA stations have been systematically scrutinized for tremor in the period March 2009 to October 2011 (when the SAMBA station network was in its complete configuration and recording continuously). Envelopes of the filtered waveforms were calculated, which were then decimated, smoothed and reduced by a mean amplitude. Tremor could be identified due to its long-duration signal appearing on the majority of stations as shown in Figure 6.4. The tremor is located using cross-correlations of the tremor signal and a grid-search method over the delay between the maxima of the cross-correlation

Figure 6.4. (a) see (b) for details.

Figure 6.4. (b) Tremor signal on waveform envelopes on (a) 14 July 2009 and (b) 15 November 2009. Arrows indicate start and end time.

functions assuming S-wave propagation velocities to the stations (Wech and Creager 2008). In previous studies, this method has been shown to reliably identifying a common source of the signal but is insufficient for determining accurate depth estimates (depth uncertainties in the order of 15 km; Wech and Creager 2008; Shelly and Hardebeck 2010).

Most of the tremor originates southeast of the SAMBA station network close to or beneath station COVA, which started recording in January 2010 (southernmost red station in Fig. 6.5). The tremor occurs irregularly with a few days to several months between individual tremor sequences and signal durations of 5–30 min. Magnitudes have been determined by calibrating the maximum amplitude of the signal with those determined for earthquakes. Magnitudes of the tremor signal are equivalent to $M_L \sim 1$ earthquakes. Despite the large depth uncertainties, the tremor can be concluded to occur at and below the lower end of the seismogenic zone (Fig. 6.6). This depth range overlaps with a high-attenuation zone (the lowest Q values in the whole of South Island) extending downwards from 22 km depth (Eberhart-Phillips et al. 2008). Eberhart-Phillips et al. (2008) associated the 10 km-thick, low-Q region with the downdip extension of the Alpine Fault. They interpret the high attenuation as an indication of metamorphic fluids within a high-strain zone. This region has slightly lower V_P velocities than surrounding regions (Eberhart-Phillips and Bannister 2002) but in general the structure resolved in the attenuation model is not resolved in the 3-D velocity model (Fig. 6.5). However, the 2-D velocity profile along SIGHT transect T2 (Fig. 6.6) reveals a low-velocity zone extending downwards to 25–30 km depth (Scherwath et al. 2003). The tremor locations occur in the vicinity of Mt. Cook, the highest elevation in the Southern Alps, at depths in the mid to lower-crust and upper mantle. There appears to be a lineation with depth that dips in the opposite direction to the Alpine Fault (Fig. 6.6). This suggests that tremor extends downwards beyond the Alpine Fault and may represent deep slip where the bending of the Pacific and Australian plates is largest. Kohler and Eberhart-Phillips (2003) propose that deep earthquakes in the Southern Alps region are associated with high strain gradients. The deep earthquakes recorded in this study locate at 50–60 km depths, below the tremor (Fig. 6.6). Therefore, increasing strain-rate values with depth may explain the semi-brittle and brittle behaviour observed. If the tremor locations outline the plate interface, then its dip is much steeper than previously assumed. In general, this is in agreement with two-fault models of GPS measurements (e.g. Beavan et al. 1999) that require an antithetic, slipping fault at depth.

The tremor durations of <30 min obtained in this study are similar to those recorded at the San Andreas Fault (Nadeau and Dolenc 2005). Tremor durations show a large variability but tremor in the Cascadia and Japan subduction zones can last weeks to months (e.g.

Figure 6.5. Tremor locations (black) for tremor events between March 2009 and October 2011 in comparison with the Q -structure (Eberhart-Phillips et al. 2008) and velocity structure (Eberhart-Phillips et al. 2010, small) at 30 km depth. The dashed line marks the position of the cross-section illustrated below. The white area outlines a region of high reflectivity observed in the migrated CDP'98 reflection profile (grey line in both panels; Stern et al. 2007).

Figure 6.6. Tremor locations (white) and seismicity (black) projected within ± 15 km onto the velocity profile along SIGHT transect T2 of Scherwath et al. (2003). The blue lines show reflectors as explained in Fig. 6.5. Note that the resolution of the velocity model is variable in the upper mantle and confined to the central part of the profile.

Beroza and Ide 2011). In addition to the comparable durations, several similarities between the tremor recorded in the Southern Alps with the tremor near Cholame exist. Firstly, the tremor locations form a cluster that does not extend northwards along the fault (Fig. 6.5 and Fig. 3a of Becken et al. 2011). Secondly, the depths of tremor locations are comparable (Nadeau and Guilhem 2009; Shelly and Hardebeck 2010; Becken et al. 2011). And thirdly, bright seismic reflections and a low-velocity zone occur near or within the tremor region (at depth of 15–22 km near the San Andreas Fault according to Trehu and Wheeler 1987; at depths of 5–35 km along SIGHT transect T2; T. Stern pers. comm. 2012). The tremor near Cholame occurs in a high-resistivity zone adjacent to a low-resistivity zone extending along strike which is inferred to be the fluid source for the tremor (Becken et al. 2011). Along the strike of the San Andreas fault, a change from high (1000 Ω m) to low resistivity (10–100 Ω m) takes place at a depth of ≥ 10 km and the tremor activity diminishes in the low-resistivity region. In general, low resistivities extending to the surface within fault zones have been associated with creeping faults (Unsworth and Bedrosian 2004; Bedrosian et al. 2004). However, a distinctive, narrow, low-resistivity fault zone along the locked segment of the San Andreas Fault in the Carrizo Plain has also been identified (Unsworth et al. 1999).

On the San Andreas Fault, tremor occurs in a region of transition between locked and creeping sections at the surface. This region has been associated with resistivity changes (Becken et al. 2011). Due to similarities of the Alpine Fault tremor with that of the San Andreas Fault, and the Alpine Fault's attenuation and velocity structure, this may suggest

strength variations along the strike of the Alpine Fault. The dip of the Alpine Fault changes from steeply-dipping to vertical between Fox Glacier and Haast. Whether this change is abrupt or gradual is not known. Little et al. (2005) suggest a localised steepening of the dip of the Alpine Fault by 15–20° between Franz Josef and Fox Glacier just north of the tremor locations. Spatial variations in fault zone properties at shallow depths have been documented recently along the Alpine Fault in this region (Boulton et al. 2012).

6.3. Tomography

A simultaneous inversion for hypocentres and velocity structure can resolve systematic uncertainties in the velocity model as shown in Fig. 3.2 a) and b). The tomography method of Zhang (2003) is applied, which is based on the double difference approach of Waldhauser and Ellsworth (2000) and makes use of three types of data in the inversion: absolute arrival times, differential catalogue travel-time residuals, and differential cross-correlation travel-time residuals. Due to the incorporation of cross-correlation data, double-difference tomography has better fine-scale resolution in areas of high earthquake density than conventional tomography (Zhang and Thurber 2006). The forward problem, the calculation of travel-times between earthquakes and stations, is computed using the pseudo-bending 3-D ray-tracing algorithm of Um and Thurber (1987) (which is accurate for ray lengths up to at least 60 km; Haslinger and Kissling 2001). This algorithm adjusts an initially straight ray-path between source and receiver in a piecewise manner by perturbing the position of the midpoint to determine the final ray path. Distance weighting controls the maximum separation between pairs of events. Weighting is applied according to the *a priori* data uncertainty, the different weights assigned to P- and S-phases of catalogue and cross-correlation data in the particular iteration, and the relative weighting factor between catalogue and cross-correlation data, which determines the overall contribution of the individual data types in the iteration.

The velocities obtained are smoothed over neighbouring grid nodes in order to retain only as much structure as can be resolved from the data. The same smoothing weights in the horizontal and vertical directions are used. Damping is required to ensure convergence of the system since not all model parameters can be determined. The optimal damping and smoothing parameters are determined by calculating trade-off curves of the data variance versus the model variance (Eberhart-Phillips 1986) as shown in Fig. 6.7. For this, we run one iteration of joint hypocentre relocation and velocity inversion with equal weights for cross-correlation and catalogue data with a series of damping (10, 25, 50, 75, 85, 95, 105,

115, 150, 250, 500, and 1,000) and smoothing values (3, 5, 10, 15, 20, 30, 50, and 100). The optimal smoothing value which minimises the model variance is 10. Damping values vary between 45 and 160 depending on the iteration.

The model resolution is given by the ray density in each block and the number of rays traversing the block in different directions. In *tomoDD*, instead of calculating the full resolution matrix, the derivative-weighted sum (DWS) is calculated as an indicator of well- and poorly-sampled regions. The DWS represents a measure of the number of rays that traverse near a grid point. It is affected by the weight of the pick in the input data as well as the relative weighting of the data types (cross-correlation versus catalogue data) in the particular iteration. The ray direction is not taken into account. We consider a DWS threshold of 15 as the lower resolution limit according to the value used by Zhang and Thurber (2006).

The earthquake data selected for the joint inversion consist of 1012 events in the study area that were recorded by more than six stations and have azimuthal gaps of $\leq 200^\circ$. A total of 9266 P and 8591 S catalogue picks are available and 78505 and 69417 cross-correlation differential travel-times for P- and S-waves, respectively.

We examine the effects of several different starting velocity models; the 1-D velocity model of O' Keefe (2008), our modified version of that model (without the low-velocity surface layer; Table 1.2), a gradient model based on the same velocities and the 3-D model of Eberhart-Phillips et al. (2010), which was smoothed and interpolated onto the same grid used for the 1-D models. We test V_P/V_S -ratios of 1.64, 1.68 and 1.73 for the starting velocity models. A V_P/V_S -ratio of 1.73 corresponds to the velocities in a Poisson solid. O' Keefe (2008) obtained a V_P/V_S -ratio of 1.68 in the inversion of a minimum 1-D velocity model (Kissling 1988). We obtain stable values of 1.64 ± 0.03 determined from Wadati diagrams of the input hypocentre data as shown in Table 6.2 and Fig. 6.8. This suggests that the V_P/V_S -ratio is even lower than obtained by O' Keefe (2008) and is consistent with laboratory results for Southern Alps schist samples studied by Godfrey et al. (2000) (for steeply-dipping bedding, compare their Fig. 13b), average velocities of SmS and PmP-phases in the Australian Plate (Fig. 4.9 of Pulford 2002) and V_P/V_S -ratios obtained tomographically by Eberhart-Phillips and Bannister (2002) ($V_P/V_S < 1.65$ within 20 km of the fault). The grid used in our inversion is rotated by 55° so that the X-axis is oriented parallel to the Alpine Fault and the Y-axis perpendicular to it as shown in Fig. 6.9. The finest grid node spacing of 5 km extends from just offshore to 60 km southeast of the Alpine Fault and from GeoNet station FOZ (south-westernmost station) to 10 km northeast of the north-easternmost station WVZ. This area of densest grid spacing is enclosed by a

Figure 6.7. Tradeoff curves of (a) the data variance (weighted catalogue RMS) versus the model variance (variance of the model parameters), (b) the conditioning number (which expresses the stability of the solution and indicates ill-conditioned system of equations) versus the damping and (c) the data variance versus the solution variance (variance of the hypo-centre changes and model parameters) for different damping and smoothing values. The symbol and the colour used for damping and smoothing apply to all three panels.

Number stations	RMS [s]	Mean V_P/V_S	Std V_P/V_S	Number events
3	0.5	1.65	0.05	1067
6	0.1	1.64	0.04	280
6	0.2	1.65	0.04	655
6	0.3	1.65	0.04	709
7	0.1	1.64	0.03	196
8	0.1	1.64	0.04	116
9	0.1	1.65	0.03	65
10	0.1	1.64	0.03	28

Table 6.2. Mean and standard deviation of V_P/V_S -ratios determined for differing number of events, stations and RMS values from the input hypocentre locations

10 km-wide band with 10 km node spacing. All nodes outside this region have ≥ 15 km spacing. The origin of the grid is centred at Mt. Cook village for consistency with the 3-D velocity model of Eberhart-Phillips and Bannister (2002). The spacing in depth is 2 km between -3 and 18 km depth, 5 km below 18 km depth and 10 km below 40 km depth. This spacing is slightly smaller than the expected sizes of the velocity anomalies that can be resolved.

We start with the relocation of all earthquakes to test if we can reproduce our hypoDD results from Section 3.4.4. This is the case, except that the overall number of relocated events is larger (1012 versus 936 events) because tomoDD can account for topography. In the second step, we fix all hypocentre locations and invert for velocity structure only. We then continue with the joint inversion of hypocentre locations and velocity structure for the different starting velocity models. Comparison of the results (Fig. 6.12) shows that the same features are resolved despite the different “background” velocities, which mainly affect the amplitude of the resolved velocity structure. All inversions have in common that the hypocentres shift downwards by on average 1.75 km and slightly eastwards (approximately 320 m; Table 6.3).

To test the resolution capabilities, we perform a checkerboard test in which the input velocity model is perturbed by 5% within blocks of three, four and six nodes separated by one node, which retains its previous velocity value. This corresponds to 15×15 , 20×20 and 30×30 km² blocks of alternating high and low velocity in the area of finest grid spacing. The same 5% perturbation is applied to the V_P/V_S -ratios. For this “checkerboard” velocity model, synthetic absolute, cross-correlation and differential travel-times are calculated. We use the standard deviation of the P- and S-phase arrival-times of the input data of 0.11 s

Latitude [°]		Longitude [°]		Depth [km]		Velocity model
mean	std	mean	std	mean	std	
0.0055	0.0081	-0.0007	0.0108	1.66	2.16	Eberhart-Phillips et al. (2010)
0.0003	0.0066	0.0048	0.0101	1.79	2.09	O' Keefe (2008)
0.0003	0.0066	0.0050	0.0101	1.83	2.15	gradient
0.0005	0.0064	0.0045	0.0100	1.75	2.01	$V_P/V_S=1.73$
0.0005	0.0064	0.0045	0.0100	1.75	2.01	$V_P/V_S=1.68$
0.0005	0.0063	0.0047	0.0100	1.72	1.95	$V_P/V_S=1.64$

Table 6.3. Shifts of hypocentres during the joint inversion using different starting velocity models

Figure 6.8. Histogram of V_P/V_S -ratios of the input data from all events

and 0.16 s, respectively, to determine the distribution of Gaussian noise which is added randomly to the synthetic absolute travel-time data. These synthetic travel-times are then used as input for the inversion of the velocity model with the same settings as in the velocity inversion of the real data. The similarity between the recovered and the input checkerboard model reveals how well-resolved the velocity structure actually is (e.g. Lvque et al. 1993). Figure 6.13 shows the comparison of the input and recovered velocity models.

Lvque et al. (1993) showed that the resolution capabilities resolved using a checkerboard model can be misleading under unfavourable but not unrealistic circumstances (if the solution matrix has eigenvalues equal to zero). They showed that if small-size structures can be retrieved this has no implications for the resolvability of larger structures and that the larger the structure the harder it is to resolve. To test the resolution, spheres and blocks of different size, orientation and amplitude should be used instead. Ideally several

Figure 6.9. Grid orientation and initial hypocentre locations used for the inversion.

Figure 6.10. Velocity profiles (V_P left, V_S right) parallel to the Alpine Fault at X=-20=const. for different input velocity models: preferred 1-D velocity model (top), gradient model (centre) and Eberhart-Phillips et al.'s 3-D model (bottom).

Figure 6.11. (a) Comparison of tomographic results for the preferred velocity model (left) and gradient model (right) as starting models for depths of 3 (top), 5 (centre) and 7 km (bottom) depth. The corresponding resolution maps are shown in (b)

Figure 6.11. (b) V_p -velocity changes using the model of Eberhart-Phillips et al. (2010) as a starting model (left) and resolution (right) for depth sections at 3 (top), 5 (centre) and 7 km (bottom) depth.

Figure 6.12. (a) Obtained V_S -velocity changes in comparison using V_p/V_s -ratios of 1.73 (left), 1.68 (middle) and 1.64 (right) in the starting model. The upper row shows depth section at 5 and the lower at 7 km depth.

Figure 6.12. (b) V_S -velocity changes using the model of Eberhart-Phillips et al. (2010) as a starting model (left) and resolution (right) for depths sections at 3 (top), 5 (centre) and 7 km (bottom) depth.

velocity inversions with different grid sizes and orientations should be performed (e.g. Koulakov et al. 2009) to test the resolvability and the stability of the obtained velocity structures and the extent of horizontal smearing. Also stepwise refining of the grid spacing in regions of high resolution is commonly applied (e.g. Eberhart-Phillips 1986). The same inversion with finer grid spacing of 2 km horizontally and 1 km vertically was tested. This resulted in a small region in the centre of the SAMBA array which could be resolved. Since the shape and amplitude of the blocks in the checkerboard model are poorly resolved, we omit further resolution tests. Our velocity model serves rather as an indicator of the inaccuracies of the 1-D velocity model used for the hypocentre location. The retrieved velocity structure suggests higher V_S -velocities at 7 km depth than used in the starting velocity model. However, the shape of the velocity anomaly is unreliable as shown in the checkerboard test. Too few rays propagate parallel to the Alpine Fault to resolve the shape of the anomalies in that direction. The DWS and the resolved structure is confined to a triangular area controlled by the distribution of the SAMBA stations. A repetition of the velocity inversion when more data and a denser station distribution are available would be preferable.

6.3.1. Expected errors from velocity variations

Although we can only resolve a small area in the centre of the SAMBA array with the current dataset, we can use the obtained velocity to give better estimates of the uncertainty caused by the velocity structure. If the velocities are elevated by 10% at 5–7 km depth (as indicated from the velocity structure inversion), this will shift the hypocentres downwards by ~ 100 m per kilometre propagated through the higher-velocity medium. This effect is not as large as the effect expected due to anisotropy. Petrologic studies of the Haast schist have shown that the difference between fast and slow P-wave velocities due to anisotropy can amount to 0.77 km/s (Godfrey et al. 2000), corresponding to 13% P-wave anisotropy. A similar result (17%) was obtained by Okaya et al. (1995). The P-wave velocity of 5.67 km/s applied to the uppermost layer in our study matches the velocities of the slow P-wave (propagating perpendicular to the foliation) obtained by Godfrey et al. (2000). Consequently, in the most unfavourable case of P-waves propagating subvertically (parallel to the foliation) over a 8 km path through the anisotropic schist, we would expect a hypocentre location error of 1 km, solely due to the anisotropic rock. However, this error lies within the error volume determined from probabilistic non-linear location which usually comprises errors in the order of 2 km in each direction (cf. Section 4.2.1).

Figure 6.13. Input and retrieved synthetic velocity models obtained for different block sizes and depths

6.4. Moment and strain release by microearthquakes

Estimates of the strain-rate represented by microearthquakes in comparison with the strain-rates determined geodetically can be used to determine what fraction of the plate motion has been relieved and what has been accumulating (Hyndman and Weichert 1983; Jackson and McKenzie 1988). Geodetic observations over the last 15 years have shown that shear-strain rates are highest within 30 km of the Alpine Fault (Beavan and Haines 2001; Beavan et al. 2007, Fig. 6.14 and 6.15). The maximum horizontal contraction direction shows a systematic clockwise rotation with distance from the Alpine Fault. This direction differs by 10° from the maximum horizontal stress direction determined from microearthquakes in this study. The magnitude distribution of earthquakes versus the strain-

Figure 6.14. Shear-strain rates obtained from 1996-2010 GPS data with the grey bars giving the orientation of maximum contraction while the bar length and the background colour indicate the absolute value. The five bars with error estimates are maximum horizontal compressive stress directions obtained from focal mechanism inversion as shown in Fig. 4.8. The length of the latter bar has no meaning as the magnitude of the stress cannot be determined from focal mechanism inversion. All other data shown are updated results of Beavan and Haines (2001) and Beavan et al. (2007)

rate obtained geodetically are shown in Fig. 6.16 for two transects (Fig. 6.17). Estimates of the long-term average slip rates of earthquakes are calculated from the cumulative moment of the earthquakes. Since the distribution of microearthquakes recorded in this study resembles

Figure 6.15. Velocity and strain rate versus distance from the Alpine Fault as published by Beavan and Haines (2001) (their Fig. 7). The profiles are averages of measurements in the central Southern Alps region where along-strike variations are minimal. The top panel (a) shows shear-strain rates and velocities parallel to the Alpine Fault whereas (b) shows contractional strain-rate and velocities normal to the Alpine Fault.

that of the at least 30 years recorded by GeoNet (Fig. 4.1 b), we assume that the 1.5 years of data available are representative of the decadal seismicity. Secondly, we assume that the local magnitude M_L is equal to the moment magnitude M_W , although Abercrombie et al. (2001) and Leitner et al. (2001) obtained $M_W = M_L - 0.25$ and $M_W = M_L - 0.7$, respectively, which indicates that our assumption is likely to overestimate the moment magnitude by up to 2.57×10^{15} N m. This conversion is required for the calculation of the seismic moment M_0 , which Hanks and Kanamori (1979) determined to be related to M_W by

$$\log M_0 = 1.5M_W + 9.0 \quad (6.1)$$

with M_0 in [N m]. This relation was obtained for intermediate to large earthquakes but is consistent with microearthquakes scales in California (Bakun 1984). An alternative scale derived from microearthquakes in southern Germany (Hainzl et al. 2002) results in larger

M_0 for $M_L > 1.5$.

Figure 6.16. Earthquake magnitude (black) and computed, present-day shear-strain rate (blue) for transects shown in Fig. 6.17 along the Whataroa river in (upper panel, profile 2) and just south of Franz Josef in (lower panel, profile 5). The shear-strain rates at the grid nodes shown in Fig. 6.15 were projected onto the direction of the profiles shown in Fig. 6.17 within distances of ± 10 km.

The average strain rate $\dot{\epsilon}_{ij}$ in a volume V can be calculated by summing the moment tensor components M_{ij} of all earthquakes:

$$\dot{\epsilon}_{ij} = \frac{1}{2\mu\tau V} \sum_{n=1}^N M_{ij}^n \quad (6.2)$$

(cf. Jackson and McKenzie 1988, and references therein), where μ is the shear modulus and τ the time. The volume is determined as the area marked in Fig. 6.17 ($102 \text{ km} \times 38 \text{ km}$) multiplied by the average seismogenic depth (12.2 km) in this area. The time that our data spans is 1.48 yr. The components of the moment tensor are given by the scalar seismic moment and the components of the unit normal vector to the fault plane and the unit slip vector in the fault plane (Stein and Wyssession 2003). Since focal mechanism solutions could not be obtained for all events but magnitudes could, we first calculate the cumulative moment for the area $M_0 = 5.359 \times 10^{15} \text{ N m}$, assuming that the slip vector of all events lies in the direction of the maximum resolved shear-stress in the prevailing stress field. This is equivalent to a single M_L 4.48 earthquake with typical slip values of 4 cm on a rupture plane of 1 km^2 (Sibson 1989). The total value depends only slightly on the largest annual events (M_L 4.19 without the two largest events). In relation to the total area, this accounts for only a small fraction of the plate motion. Using equation 6.2, the total strain release is found to

Figure 6.17. Cumulative moment release for square cells of 0.04° length in the study area. The two profiles shown by grey lines correspond to profiles 2 and 5 in Fig. 4.2.

be $\varepsilon=1.9 \times 10^{-9}$ and the strain release rate $\dot{\varepsilon}=1.3 \times 10^{-9}=0.001$ microstrain/yr. In comparison to the strain rate derived geodetically of ~ 0.4 microstrain/yr (Beavan and Haines 2001; Beavan et al. 2007, Fig. 6.14), an insignificant amount of strain is released microseismically. Similar results have been obtained for the energy released by microearthquakes on creeping faults (Gans et al. 2003). The total strain released by microearthquakes is estimated to be $<0.01\%$ of the total interseismic slip on the Hayward Fault, which repeatedly fails in $M_L \geq 6$ events (e.g. Malservisi et al. 2005). This suggests that the strain resulting from the relative plate movement between the Australian and Pacific plate in the upper crust is almost entirely accumulated in the interseismic cycle and results in large earthquakes on the Alpine Fault and secondary structures as shown by Cox et al. (2012) in the central Southern Alps region. Cox et al. (2012) estimated the magnitudes from the length of the fault traces adjacent to the Alpine Fault to be $M \leq 7.5$.

6.5. Fault-guided waves

Fault-guided waves are trapped waves that propagate in the damage zone of a fault, which is characterised by reduced velocities relative to the wall rocks and consequently forms a natural waveguide (Li et al. 1997). A guided wave develops due to constructive interference between multiple P- and S_V (Rayleigh-type) or S_H -reflections (Love-type) at the boundaries

of the fault zone (Li et al. 1990). The higher the velocity contrast the better the trapping efficiency of the waveguide (e.g. Li et al. 1998). Fault-guided waves can be observed on stations that are located in the fault damage zone or in close proximity (less than 5 times the width of the damage zone, based on modelling; Li and Vidale 1996). They are best observed on dense linear arrays crossing the fault or perpendicular to the fault. In this case, those earthquakes that occur within or in close proximity to the fault zone and exhibit fault-guided waves can be discriminated from off-fault events. As the fault-guided waves travel slower than the S-wave, the moveout with distance or depth characterises the velocity reduction in the damage zone: the broader and slower the waveguide, the longer the duration and the lower the frequencies (Li and Vidale 1996). Characteristic of fault-guided waves are the low-frequency content (3–6 Hz), the large amplitude (that exceeds the S-wave) and their dispersive nature (e.g. Li et al. 1997).

Fault-guided waves provide a means of mapping the fault at depth (Li et al. 1998, e.g. for the Nojima fault by), determining the fault plane of blind faults, and testing if fault planes are continuous (Li and Vidale 1996). The width of the fault damage zone, with typical values ranging between a few tens to hundreds of metres (e.g. Li et al. 1994), and the velocity reduction (generally 20–30%) can be obtained by comparison of observations with synthetic data.

Due to the dispersive nature and attenuation of fault-guided waves, not all earthquakes within the fault can be used for analysis. Fault-guided waves are best recorded at distances of 10–25 km if they are generated by earthquakes of small magnitude ($M_L < 3.5$; Y.-G. Li pers. comm., 2011). At this range, the incidence angles are steep, the phase arrivals are not too emergent and the earthquake sources still resemble a point source.

We have examined waveforms recorded by the SAMBA and DFDP10 arrays. Only three stations (POCR, DCR, GCK) qualify as able for recording fault-guided waves due to their proximity to the fault. However, these stations have intermediate to poor recording qualities and often exhibit cultural noise. The latter two DFDP10 stations were only recording during January–April 2010, which restricted the search for fault-guided waves markedly. Station COSA appears to be located in the vicinity of the Alpine Fault due to cataclasite outcrops next to the road a few hundred metres north of the station. However, this cataclasite is part of the abandoned, out-of-sequence Waikukupa thrust east of the active Hare Mare thrust (Norris and Cooper 1997).

Since the position of the Alpine Fault at depth is not known and the seismicity is dis-

tributed, we have examined many events. A few candidate events were found but no obvious fault-guided waves were identified. In a second approach, we examined events with long, dispersive coda waves recorded by station WHYM to test whether a smaller fault in the vicinity of that station could have produced fault-guided waves. Since there is only this one station, we examined whether a moveout for events from different depths/distances can be observed. However, it is not clear which fault (within 1 km distance) could have generated these waves. We perform a search for events with long wave trains following the S-waves that exhibit either long periods, large amplitudes or both. A total of 328 earthquakes within a $0.1^\circ \times 0.1^\circ$ square in the vicinity of station WHYM were checked for fault-guided waves. The distribution of the earthquakes is scattered throughout the area and does not align with any mapped faults (Fig. 6.18). As all of the events lie at distances greater than 1.5 km from the station, only a dipping fault could be responsible for the generation of fault guided waves. Li et al. (1998) showed for earthquakes generated on and around the Nojima fault that long wavetrains are not always associated with fault trapped waves (Fig. 4b of Li et al. 1998); this association could only be showed using a multitude of recording stations across the fault with similar low-frequency waves in the S-coda. Particularly shallow earthquakes often exhibit elongated wavetrains due to reverberations in the near-surface layers (e.g. Thouvenot et al. 2009). We suggest that the majority of the events with long, dispersive codas at WHYM originate due to the shallow depths of the events (Fig. 6.18). The frequencies of the S-coda decrease with time for most of the events, suggesting that the phases are attenuated. Fault-guided waves are difficult to identify without better knowledge of which events occur within and outside the fault plane and which stations are located within the fault's surface trace. This can be investigated if more stations are available within or across the fault zone, but not with the current configuration of the SAMBA stations. Fault-zone guided waves have been identified on records of a short-period borehole sensor located in the fault zone at the DFDP-1 drillhole. This station was installed in February 2011 by J.D. Eccles, University of Auckland. Those data will be combined with SAMBA recordings for further research on fault-guided waves that exceeds the scope of this thesis.

Figure 6.18. Shallow earthquakes (≤ 4 km) recorded on the vertical component of station WHYM between November 2008 and March 2009 (top) and all shallow events (red) investigated in the marked area plotted in map view (bottom).

Figure 6.19 Shallow earthquakes (≤ 4 km) recorded on the vertical component of station WHYM between March 2009 and April 2010. The text above each trace gives the date, time and depth of the event shown.

7. Summary and conclusions

The SAMBA network in combination with the DFDP10 and GeoNet networks formed the densest station array with the longest station deployment ever installed in the area of highest topography of the Southern Alps. The main focus of this study has been the microseismicity, which is revealed in much more detail than in any previous study, and the remotely triggered seismicity, which has not been observed before or reported on in New Zealand. Our results apply particularly to processes occurring in the upper crust but phenomena observed in the lower crust and upper mantle are briefly reported on, too.

7.1. Main contributions

Our study enables a number of key conclusions to be drawn regarding the seismicity distribution and seismogenesis as outlined below. These have been or are planned to be the subject of publications.

- The thickness of the seismogenic zone varies with distance from the Alpine Fault ($10 \pm 2(1\sigma)$ km beneath the surface trace, 8 ± 2 km within 20 km distance and 15 ± 2 km at larger distances from the fault), and follows the 390 Ω m resistivity contour (Fig. 4.12). Earthquake hypocentres are concentrated in areas of high-resistivity, whereas rocks with resistivities <100 Ω m are too hot, too fluid-saturated, or too weak to produce detectable earthquakes. Consequently, the combined observations allow us to determine width and extent of zones prone to failure in earthquakes and directly reflect rheological processes at depth.
- Remotely triggered seismicity is the result of dynamic stresses of ~ 0.9 MPa imposed by the surface waves of large earthquakes that cause fluid diffusion from a shallow (overpressured) fluid reservoir into the surrounding fractured rock. This provides a mechanism for ubiquitous observations of elevated seismicity following large earthquakes that persists for as long as weeks after the passage of the surface waves.

- The Alpine Fault is the second major transform plate boundary found to generate tectonic tremor of short durations below the seismogenic zone at depths of 15–50 km, suggesting that tremor is the manifestation of slow, deep slip in various tectonic settings.

These and other key results are elaborated on and put into a broader context below.

7.2. Main results

7.2.1. Earthquake distribution

The earthquake density, depth and magnitude distribution is highly variable at distances up to 40 km southeast of the Alpine Fault. The zone that exhibits microseismicity adjacent to the Alpine Fault is narrow (15–20 km) at the southern end of the study area, wide (35–45 km) in the centre and narrow (20–25 km) in the north. This overall width reflects the distance between the Alpine Fault and the Main Divide Fault Zone (Fig. 1.4). The seismicity in plan view is distributed around a major low-seismicity zone extending between the Whataroa and Wanganui rivers at 15–30 km distance from the fault (Fig. 4.1). The low-seismicity region is not obviously related to lithology, topography and bedding in this area but correlates with high resistivities and low attenuation, suggesting a strong and unfractured block that is enclosed by a damage zone of high earthquake activity (high attenuation) (Fig. 4.13). A spatial pattern of quiescence and persistent earthquake activity is distinct in the long-term and short-term seismicity. In general, the seismicity distribution does not correspond to mapped faults at the surface (Fig. 4.1). The seismicity is confined to shallow depths (90% above 12.2 km and centred at 5.3 km), and correlates well with high resistivities and high attenuation in the area.

Little low-magnitude ($-0.14 \leq M_L \leq 2.8$, with the lower value being the detection threshold) seismicity occurs within 10 km of the Alpine Fault, most likely reflecting different numbers and characteristics of fractures in the highly-compacted and metamorphosed schist relative to the greywacke further southeast. Four large seismicity clusters have been identified which may represent active backthrusts that act as fluid conduits (Fig. 4.1 (c) and 4.4). The energy released by earthquakes in the northeastern cluster is ~ 100 times larger than in the southwestern cluster (Fig. 6.17) due to the generally larger events (87 events of $1.8 \leq M_L \leq 4.3$ versus 28 events of $1.8 \leq M_L \leq 3.5$). Larger earthquakes occur at the edges of the seismicity clusters around the gap and in the vicinity of the 1984 M_W 6.1 Godley Valley

earthquake (compare also $M_L \geq 4$ events recorded by GeoNet in the last 35 years; Fig. 4.11).

The velocity structure is poorly-resolved in tomography with the current dataset but in general the modified 1-D velocity model of O' Keefe (2008) is adequate for the location of shallow earthquakes within the SAMBA station network.

Some deep earthquakes just west and southwest of the SAMBA array at depth ≥ 50 km were recorded. These events occur below the semi-brittle zone at ~ 30 km in which the majority of the tremor locates. The brittle faulting at these upper mantle depths may be explained by high strain-rates caused by plate bending in this zone.

7.2.2. Magnitude distribution

The magnitudes of earthquakes in the central Southern Alps region reported by GeoNet for source–receiver distances of ≤ 70 km are underestimated due to the small attenuation coefficient of $\gamma = 6.7 \times 10^{-3} \text{ km}^{-1}$ employed. Frequency-dependent attenuation has to be considered in order to obtain normally distributed magnitude residuals. We derive a frequency-dependent attenuation coefficient $\gamma = \gamma_0 f$ with $\gamma_0 = (1.89 \pm 0.02) \times 10^{-3} \text{ s/km}$. The magnitude uncertainties obtained span 0.2–0.5 magnitude units in part due to the uncertainty of the GeoNet magnitudes used for calibration.

7.2.3. Focal mechanisms and stress field

Focal mechanism solutions have been obtained for 379 earthquakes ($M_L > 0.44$) and exhibit predominantly strike-slip mechanisms. Typical uncertainties in strike, dip and rake are $\sim 30^\circ$. The estimation of the stress field from focal mechanisms indicates a homogeneous stress field prevails throughout the study area, with the maximum horizontal compressive stress direction S_{Hmax} oriented at $115 \pm 10^\circ$. The Alpine Fault is oriented at a 60° angle to S_{Hmax} and is thus misorientated in the sense of Andersonian theory of faulting (Anderson 1951), suggesting either low coefficients of friction or high fluid pressures are required for the fault to fail.

7.2.4. Swarms and triggered seismicity

Earthquake swarms occur frequently in the central Southern Alps region and are well recorded by the SAMBA array. All 14 swarms described here comprise at least 10 similar, low-magnitude events ($M_L \leq 2.8$). They have either short durations (≤ 12 days) or consist of several bursts of activity (≥ 30 days). The swarms occur on near-vertical faults that are favourably oriented in the stress field according to Anderson's theory of faulting. A characteristic inter-event time pattern was identified with accelerating inter-event times in the beginning of the swarm followed by deceleration, characteristic of a decay process. Remotely triggered earthquake swarms occur if the peak ground velocity exceeds a threshold of 0.7 cm/s at station FRAN, corresponding to peak dynamic stresses of ~ 0.9 MPa. The triggered seismicity is followed by a quiescence period of approximately the same length as the time period of the elevated seismicity. This suggests that the faults experience clock-advance. The locations and delays of the triggered swarms are consistent with a common fluid source from which fluids diffuse in response to the dynamic stresses imposed by the surface waves. In general, the seismicity in the centre of the SAMBA array has a lot of similarities with that in a geothermal area, due to the thermal structure and fluid contents, but the swarms appear to be of one type (as seen from their inter-event time pattern) and are not as complicated as in geothermal areas.

7.2.5. Tremor

Tremor observed in the vicinity of the Alpine Fault below the seismogenic zone suggests deep slip on the downdip extension of the Alpine Fault and in the region where the bending of the Pacific and Australian Plates is largest. Similarities with tremor detected on the San Andreas Fault near Cholame in terms of duration, depth, spatial extent and evidence for high fluid-pressure nearby may indicate seismic property variations in the lower crust and upper mantle along the strike of the Alpine Fault.

7.3. Implications

Our results demonstrate the value of dense, intermediate-term station networks that enable microseismicity patterns to be studied in detail. Long-term measurements with sparse station coverage can only reveal the main features of the seismicity (e.g. Robinson 1978), but good depth resolution of those networks is limited to deep earthquakes (with depths on

the order of the station spacing). The distribution of earthquakes in the central Southern Alps region shows little correlation with the faults exposed at the surface, suggesting that these mapped faults represent older, exhumed structures resulting from previous deformation. Microseismicity interpreted in conjunction with resistivity, attenuation and velocity models is a good indicator of rheological processes at depth. It has been shown for many locations worldwide (Table M.1, modified from Table 1 and 2 of Gürer and Bayrak 2007), that earthquakes occur in resistive structures or at the boundaries of resistive to conductive zones. High conductivity is associated with high saline fluid content (Wannamaker et al. 2002; Bedrosian et al. 2004; Wannamaker et al. 2009) and conductive regions have been inferred as fluid sources for earthquake and tremor zones (Ichiki et al. 1999; Goto et al. 2005; Gürer and Bayrak 2007; Becken et al. 2011). The maximum depth of the seismogenic layer often correlates with the top of the conductive lower crust (Gürer and Bayrak 2007). The depth of the brittle–ductile transition zone depends on the geothermal gradient, the geometry and mode of faulting, the fluid pressure, the crustal composition, the water content and the strain rate in the quasi-plastic regime (Fig. 7.1; Sibson 1984). Within 30–40 km southeast of the Alpine Fault the seismogenic thickness exhibits short-wavelength “roughness” according to Sibson (1984) with pronounced depth variations of up to 5 km. We briefly discuss below the importance of these factors in the central Southern Alps region.

The temperature distribution plays an important role in the direct vicinity of the Alpine Fault, but temperature models (Shi et al. 1996; Toy et al. 2010) exhibit long-wavelength variations, so they may not fully explain short-scale variations in the thickness of the seismogenic layer between 15 and 30 km southeast of the Alpine Fault. Sibson (1984) stated that changes in the geothermal gradient by $10^{\circ}/\text{km}$ depth along the strike of the fault (which would cause equivalent depth changes of 4–6 km of the seismogenic thickness, depending on rock type) are not unreasonable for tectonically active regimes. Therefore, we cannot exclude the possibility that the temperature may be more inhomogeneous throughout study area than revealed by the temperature models. The second most significant factor affecting the depth of the brittle–ductile transition zone according to Sibson (1984) is crustal composition. There are no indicators for lithological changes along the strike of the Alpine Fault in the study area. However, in combination with the water content and chemical alteration of the rocks, the composition may vary significantly in depth and laterally. Variations in fluid pressure certainly exist throughout the area based on the resistivity and attenuation structures. High shear-strain rates in the quasi-plastic regime within a 35 km zone adjacent to the Alpine Fault (Fig. 6.15; Beavan and Haines 2001) have created a broad zone of intense shearing and low permeabilities, as shown by the high-conductivity zones, in which trapped fluids are transported towards the Alpine Fault.

The fluid distribution appears to be tectonically controlled as suggested by the resistivity structure (Fig. 4.3). High conductivities occur beneath the strong and unfractured block between the Whataroa and Wanganui rivers and the Alpine Fault, which may form a seal. The conductivity decreases at 30 km distance from the fault, where the seismicity is high (Boese et al. 2012) and steeply-dipping faults have been mapped at the surface (Little et al. 2005). In this region, fluids can escape through backthrusts acting as fluid conduits. Wide-angle reflection data and teleseismic delays of P-wave arrivals provide evidence for high fluid pressure and reduced P-wave velocities extending from 6–35 km depth adjacent to the Alpine Fault on SIGHT transect T2. Tomographic 3-D inversion (Brikké 2010) and the 2-D velocity profile along SIGHT transect T1 (Van Avendonk et al. 2004) suggest that the low-velocity zone persists along the strike of the Alpine Fault but that the uppermost part of this zone shifts away from the fault, consistent with the seismicity.

The area of high swarm activity may be located above a fluid pocket. This area has a high fracture density and the remotely triggered swarms are consistent with diffusion from a fluid source that starts with the passage of the surface waves of the triggering events. This suggests the region is particularly susceptible to small stress changes. The background seismicity in this area consists of an abundance of small events and several earthquake swarms.

The focal mechanism solutions obtained in this study indicate that microearthquakes do not contribute to the uplift of the Southern Alps. The compressional component of the plate motion must either be accommodated aseismically (as it is the case at depth where a 1–2 km thick, high-strain mylonite zone accommodates the plate motion according to Norris and Cooper 2003) or released during the rupture of large earthquakes. Three observations support the latter: large reverse faults east of the Main Divide Fault Zone have been mapped (Cox et al. 2012), an increase of the ratio of reverse to strike-slip mechanisms with event magnitude has been seen (as shown in Fig. 4.7 in this study, see also Cox et al. 2012, their Fig. 8), and the dip-slip rates inferred geodetically and geologically differ by 3–5 mm/yr (Beavan et al. 2004; Beavan et al. 2010a; Norris and Cooper 2001, and references therein). Similarly, Hardebeck and Hauksson (2001) interpreted the strike-slip, 1992 M_W 6.2 Big Bear earthquake and the primarily strike-slip faulting regime of the young San Bernardino Mountains, as an indication that the oblique convergence there is partitioned into strike-slip and thrust earthquakes. The orientation of the thrust faults perpendicular to the maximum horizontal compressive stress direction enables them to fail in the strike-slip regime.

Figure 7.1. Factors influencing the depth of the brittle–ductile transition zone as shown by Sibson (1984) (his Fig. 7). High geothermal gradients, high water content in the quasi-plastic (QP) regime, low fluid pressure in the frictional (FR) or low strain-rates in the QP-regime lower the thickness of the frictional regime. In principle, thrust faulting decreases the frictional zone, however Sibson (1984) reports that in combination with other factors such as higher fluid pressure in the FR regime or lower geothermal gradients the QP-FR transition occurs at larger depth in thrust regimes in comparison to strike-slip regimes. For normal faulting regimes the QP-FR transition theoretically deepens compared to strike-slip regimes but effectively the transition zone becomes lower due to high geothermal gradients associated with extensional settings.

Due to pronounced variations in the thickness of the seismogenic zone, the depth extent and therefore the size of $M \geq 5.5$ earthquakes adjacent to the Alpine Fault will vary with distance perpendicular to the fault (Boese et al. 2012). The largest events in the immediate vicinity of the Alpine Fault (≤ 17.6 km) exhibit magnitudes of M_W 5.4–5.8. At distances ≥ 25 km from the fault, the magnitude of the major events lie in the range M_W 6.1–6.7 (Fig. 1.6). The area between the Whataroa and Wanganui rivers poses a hazard of a $M \geq 6.5$ earthquake at 15–30 km from the Alpine Fault.

The cumulative strain-rate released by the microseismicity within ~ 40 km of the Alpine Fault within one year is equivalent to an M_L 4.5 earthquake, which accounts for $\leq 1\%$ of

the accumulated strain-rate determined geodetically. The microseismicity distribution may correlate with high shear-strain rates (Fig. 6.16), suggesting fatigue of faults near failure in the tectonic stress field.

Inversion of the focal mechanism solutions for the tectonic stress has shown that the orientation of the Alpine Fault with respect to the maximum horizontal compressive stress direction requires the fault to be weak to fail in (large) earthquakes. Paleoseismic evidence for large earthquakes every 200–400 yr has been identified in several geomorphological features (Adams 1980; Wells and Goff 2007; Sutherland et al. 2007, and references therein), trenches across the active fault trace (Berryman et al. 1992; Langridge et al. 2010), forest (re-)establishments and tree-ring growth (Wells et al. 1999). The orientation of the Alpine Fault relative to the stress field suggests that the fault sustains a low ratio of shear to effective normal stress. Modelling of the stresses on the fault plane (Section 4.4) shows that the weakest scenario has been obtained for near-lithostatic fluid pressures and similar absolute values of the near-vertical intermediate stress S_2 and the near-horizontal least compressive stress S_3 . If the Alpine Fault acts as a permeability boundary for fluids, as suggested for faults in general (e.g. Audin et al. 2002), near-lithostatic fluid pressures within the fault zone are quite realistic. With the stress inversion performed, we cannot obtain information about the stress amplitude. However, elevated stress ratios were obtained from focal mechanism inversion in the study area, indicating similar S_2 and S_3 values at shallow depths (Fig. 4.8 and Table 4.1).

The Alpine Fault appears to be the second major strike-slip fault that exhibits tremor. While tremor on the San Andreas Fault underlies the surface fault trace at 15–40 km depth, it is not clear whether tremor is confined to the downdip extension of the Alpine Fault or extends beyond it in the lower crust and lithospheric mantle. Becken et al. (2011) suggested that tremor results from brittle failure in a ductile deformation regime due to high fluid pressures causing embrittlement. The coincidence of the tremor depth with the high-attenuation and low-velocity zone (Scherwath et al. 2003; Eberhart-Phillips et al. 2008), the bright seismic reflections extending from 15 to 35 km depth (e.g. Kleffman et al. 1998; Stern et al. 2007, and references therein) and the inferred high fluid-pressure zone (Stern et al. 2001; Stern et al. 2007) support this interpretation. Trapping of fluids in the ductile shear zone and transport towards the Alpine Fault as observed in the resistivity structure (Wannamaker et al. 2002) 60 km north of the tremor region may account for the generation of high fluid pressure adjacent to the Alpine Fault.

7.4. Suggestions for future work

Several of the subsidiary topics introduced in Chapter 6 have potential for further research. The mechanisms of deep earthquakes could be determined from inversion of P-wave polarities given a broader station distribution in the Southern Alps region (as obtained in 2012 with the installation of the WIZARD array and the ALFA array) and sufficient recording time (since approximately 1–2 deep events occur per year). The arrivals of the events studied here were impulsive and P_n polarities could be obtained for stations at distances of ≤ 280 km from the earthquake. Alternatively, mechanisms could be obtained using waveform inversion, if the events are recorded by at least three broadband stations. A superficial search for deep events in the catalogue of O' Keefe (2008) revealed at least three recorded events deeper than 25 km. These events were not studied due to uncertain polarities of the CAIF stations (O' Keefe 2008).

The tomography results could be greatly improved with more earthquake data recorded by stations to the north of the SAMBA array (as will be provided by the WIZARD array in the next few years). Of particular interest will be the velocities in the high-resistivity/low-attenuation structure to the north. With longer recording periods of station COVA, the area of resolution could also be extended to the south. As the attenuation model by Eberhart-Phillips et al. (2008) proved of importance in this study regarding the earthquake and tremor locations, a high-resolution local attenuation model for Q_P and Q_S obtained from spectral fitting of the earthquakes recorded in this study would improve the interpretation of elevated pore pressures and earthquake zones.

The swarms analysis could be extended by studying source parameters of the events. The smallest earthquake in the swarm could be used as a Green's function to remove the impulse response of the Earth. While the triggering threshold of earthquakes was exceeded only by the two large events, tremor could have been triggered by all events listed in Table 5.3 because the threshold for triggered tremor appears to be much lower (1 kPa as determined by Guilhem et al. 2010). It would also be interesting to analyse earthquakes of low-frequency recorded at station MTFO. If these events are low-frequency earthquakes they could be used to search for low-frequency events embedded in the tremor signal as identified by Shelly et al. (2006, 2007) in the subduction zone of Japan and on the San Andreas Fault near Cholame (Shelly et al. 2009; Shelly 2009).

A systematic analysis of the abundance of recorded signals related to ice movement would also be of interest. These events could be compared to the collapse of the glacier

terminal at Fox Glacier (8 January 2009 4:20 pm NZ local time) and the collapse of the face of the Tasman glacier (30×10^9 kg) that was triggered by the M_W 6.3 Christchurch earthquake (R. Dykes pers. comm., 2011). A check for seasonality in the occurrence of these events would help to identify and characterise them. In Iceland, an annual cycle has been observed for icequakes with most of the events occurring in autumn when the glaciers are thinnest and start to regrow, so that the ice load is increasing (Tryggvason 1973). Some of these signals may also result from rockfalls that occur predominantly along the Main Divide (S. Cox pers. comm., 2011). These events occur in “swarms” on cold days following a warm period when freeze-thawing occurs. A large rockfall ($\geq 100\,000$ m³ mass movement) occurs approximately every 100 years. The last major rockfall occurred on the southeastern side of Mt. Cook in December 1991 causing an equivalent seismic signal to a M_L 3.9 earthquake (McSaveney 2002).

Distinct secondary phases have been noticed on GOVA, EORO, FRAN, WHAT and POGR that could be studied to identify reflectors at depth. Preliminary modelling of the most prominent arrivals recorded by station WHYM suggests that these are not reflected from the Alpine Fault. Instead, they may indicate the presence of heterogeneous local structures that play a role in controlling the shallow maximum depth of the seismogenic zone in this region.

A. Response functions

To understand the response of a seismometer to ground movement, one first has to understand the instrument itself and the recording system as a whole. A purely mechanical (short-period) seismometer is a velocity transducer consisting of a magnet and a coil (Fig. A.1). Ground movement either moves the magnet or the coil, inducing a current in either case. The instrument output is a voltage from the coil, which is proportional to the external acceleration (e.g. Havskov and Alguacil 2004).

Figure A.1. Schematic representation of a passive seismometer, a so-called velocity transducer, from Havskov and Ottemöller (2010)

The system response of a seismometer describes mathematically how the recorded voltage

relates to the ground motion. It contains information about the inertial response of the suspended mass/coil to external motion, damping of the movement (so that excessive motion near the resonance frequency is suppressed), coil resistance, manipulations applied to the voltage output (e.g. amplification or gain), the digitalisation of the voltage and it should also account for reversed polarity (in general, a positive velocity causes a positive voltage Havskov and Alguacil 2004). To understand the inertial response of the motion of the mass/coil, which is called the phase response, the equation of motion is needed:

$$m(\ddot{u} + \ddot{z}) + d\dot{z} + kz = 0, \quad (\text{A.1})$$

with the vertical ground motion $u = u(t)$, the displacement of the mass/coil $z = z(t)$, the damping d and the spring constant k .

Using $\omega_0 = \sqrt{k/m}$ and the damping constant $h = d/2\omega_0 m$ equation A.2 becomes

$$\ddot{z} + 2h\omega_0\dot{z} + \omega_0^2 z = -\ddot{u}. \quad (\text{A.2})$$

At high frequencies, the acceleration term dominates, so $\ddot{z} \sim -\ddot{u}$ and the motion of the mass/coil is of reversed polarity to the ground motion (Havskov and Alguacil 2004). This is because of the inertia of the mass/coil, which tends to remain stationary, so that the motion is phase-shifted by π . At low frequencies, the relative displacement of the mass/coil is proportional to the ground acceleration:

$$z = -\frac{1}{\omega_0^2} \ddot{u} \quad (\text{A.3})$$

There is no phase shift at low frequencies because $\ddot{u} \sim -u$. The phase response at the resonance frequency is $\pi/2$ (Havskov and Alguacil 2004).

The digitiser converts the measured analogue voltage into a discrete number for digital data storage. This sampling is done according to the pre-defined sampling rate. Usually, the data also acquires a GPS time stamp during this process. The measured voltage is amplified, which is expressed by the gain. For the Reftek Data Acquisition System, used in this study, the recorded signal is amplified by 32 (high gain setting). An input voltage of $1.589 \mu\text{V}$ at unity gain and 49 nV at high gain is converted to the same count value.

The displacement amplitude spectrum of a damped oscillator can be expressed as:

$$A_d(\omega) = c \frac{(i\omega - z_1)(i\omega - z_2)(i\omega - z_3) \dots}{(i\omega - p_1)(i\omega - p_2)(i\omega - p_3) \dots}, \quad (\text{A.4})$$

where c is a gain factor, z are the zeros and p the poles (Havskov and Ottemöller 2010). The zeros, as the name suggests, describe the zero crossing of the displacement amplitude function. The poles describe the decay behaviour of exponential functions which are components of the displacement amplitude function and together define the complete displacement amplitude function. The phase shift increases with the number of poles (Havskov and Ottemöller 2010). The absolute value of the displacement spectrum is

$$A_d(\omega) = \frac{G\omega^3}{\sqrt{(\omega_0^2 - \omega^2)^2 + 4h^2\omega_0^2\omega^2}}, \quad (\text{A.5})$$

where G is the generator constant, a measure of the sensitivity of the coil to ground velocity. However, the seismometer response function is usually given as the amplitude response to velocity:

$$A_v(\omega) = \frac{G\omega^2}{\sqrt{(\omega_0^2 - \omega^2)^2 + 4h^2\omega_0^2\omega^2}}, \quad (\text{A.6})$$

The (velocity) response of a seismometer flattens at frequencies above the natural frequency f_0 . Below f_0 , the sensitivity of the instrument is reduced.

In summary, all the information about how the input displacement of the ground is manipulated (sensor, amplifier, filter and digitizer) to finally become the waveform of the seismogram, is stored in the response function.

A.1. Response functions of the SAMBA stations

To create a response file using SEISAN (Havskov and Ottemöller 2000), the following input has to be provided:

- Natural period of the seismometer $T_0=1/f_0$ (Table A.1 and A.2)
- The damping ratio of the seismometer h . This parameter was always set to 0.7 because it was assumed that the damping under load and not the open circuit damping h_0 is required
- The generator constant of the sensor G in [V s/m] (Table A.1 and A.2)
- The recording media gain in [counts/V], which was calculated to be $1/1.589 \times 10^{-6} = 629327$ for unity gain and $1/49 \times 10^{-9} = 20408164$ for high gain
- The digitiser sample rate, which is 200 samples/s (Section 1.4.2)

- The amplifier gain in [dB]. As no preamplifier was used we set this value to 0.

The poles and zeros were calculated with SEISAN and are stored in GSE format. Response files for the GeoNet stations were extracted from the dataless SEED volume provided by GeoNet¹. SEISAN can directly read SEED response information of poles and zeros and calculate the corresponding response function.

Natural frequency f_0 [Hz]	Intrinsic sensitivity G [V s/inch] [V s/m]		Open circuit damping h_0	DC resistance R_C [Ω]	Moving mass m [g]
2	1.58	62.205	0.61	2500	23
4	1.60	62.992	0.27	2500	23

Table A.1. Specification for the GeoSpace Technologies HS-1-LT sensor (from specification sheets number S11230 and S11510).

Sensor type	Natural frequency f_0 [Hz]	Transduction G [V s/inch] [V s/m]		Open circuit damping h_0	Coil resistance R_C [Ω]	Suspended mass m [kg]
L-4C	1	7.02	276.378	0.28	5500 ²	0.5
L-22D	2		88		4300	

Table A.2. Specification for the Mark Products sensors.

Station	Normalisation factor A_0	Norm. freq. f_0 [Hz]	Gain	Complex zeros z	Complex poles p
FOZ	2.30426×10^6	1	2042.4	0, 0	-160, -80, -0.01178-0.01178i, -0.01178+0.01178i, -180
WVZ	2.30426×10^6	1	2050.6	0, 0	-160, -80, -0.01178-0.01178i, -0.01178+0.01178i, -180
RPZ	46.2282	1	994.8	0, 0, 146.5	-73.2-37.6i, -73.2+37.6i, -0.00589-0.00589i, -0.00589+0.00589i

Table A.3. Specification for the GeoNet sensors Guralp CMG-3ESP of FOZ and WVZ, and Guralp CMG-3TB of RPZ obtained from the Dataless SEED file provided by GeoNet.

¹ obtained from <ftp://ftp.geonet.org.nz/seed/RESPONSE/>

² This coil resistance has been measured by myself for the used Marks Products L4-3D sensors number 602 and 733. Note that the sensitivity of 166.54 V s/m listed by IRIS is for the sensor with coil resistance 2000 Ω

Specifications provided for the different sensor are listed in Table A.1, A.2 and A.3. The amplitude and phase response are plotted in Figures A.2 and A.3. The latter figure clearly shows that the response function provided by GeoNet for RPZ is wrong, because it reverses the correct polarity of the instrument. We can confirm that the polarity of the RPZ sensor is correct by comparing polarities of all stations from recordings of two teleseismic events (Section B). This wrong response has been noted before (Behr 2011). It does not affect the focal mechanisms or the magnitude determinations, since polarities are picked on raw seismograms and magnitudes are determined as half the maximum peak to peak amplitude value on the deconvoluted seismograms, and a sign change does not change this value.

Figure A.2. Amplitude and phase response of the Guralp sensors at the GeoNet stations FOZ, WVZ (CMG-3ESP) and RPZ (CMG-3TB). Note that the response as provided by GeoNet is plotted in green (which shows a reversed polarity) and the corrected response function is shown in black.

Figure A.3. Amplitude and phase response of the GeoSpace Technologies HS-1-LT sensors employed at the SAMBA stations, the Mark Products sensors used as addition surface sensor (L4-3D) to supplement the borehole sensor and the Mark Products sensors (L22-3D) used for the DFDP10 array.

B. Polarity tests

The consistency of station polarities is a prerequisite for focal mechanism solutions. The polarity of all sensors is usually tested in the lab or with a “cluster” test in the field. Another possibility for testing of the polarity is to compare the first onset polarity at all stations for a teleseismic event. Serendipitously, we recorded two teleseismic earthquakes with exceptionally clear P-wave onsets, so we could use these earthquakes to check the polarity. Therefore, polarity-reversals are exactly known for our dataset. The only station which shows an emergent onset and does not reveal a clear polarity for the teleseismic earthquakes is station REYN. For this reason and due to inconsistent polarities found for this station, we excluded all polarity picks from that station from the focal mechanism analysis. Both events clearly show that the polarity is reversed for the short-period Mark Products L4-3D sensors, which were deployed at this time at station WHAT2 and POGR2.

One of the Mark Products L4-3D sensors was later installed as a surface sensor at station COVA, so this station has a reversed polarity in comparison with the other SAMBA stations. All stations of the DFDP10 array, deployed between January and May 2010, were equipped with short-period Mark Products L22-3D sensors. It was suspected that these sensors have reversed vertical-component polarities, too. One teleseismic event of intermediate depth (Fig. B.3) shows reversed polarities at stations ERE, WNQ, BON, NOL, and DRC, but is ambiguous for the remaining DFDP10 stations.

Several quarry blasts have been recorded by the SAMBA and the combined SAMBA/DFDP10 array. All polarities are upwards for the SAMBA stations and downwards for the DFDP10 stations, consistent with the inferred polarity reversal of the DFDP10 stations (Fig. B.4 and 3.13).

Figure B.1. Waveforms of the raw (unfiltered, no instrument response removed) Z-component seismograph traces of all SAMBA and GeoNet stations. The teleseismic earthquakes which produced these clear onsets are the $M_W=6.6$ of 4 October 2009 (10:58:03.6 GMT, lat= 6.77° , lon= 123.56° , depth=626 km) and the $M_W=6.9$ event of 28 August 2009 (1:51:25.9 GMT, lat= -7.09° , lon= 123.46° , depth=634 km, see Fig. B.2). Note that the travel-time for station COSA for the latter event had to be corrected by 24.235 s due to GPS uncertainties.

Figure B.2. Locations and focal mechanisms of the events shown in Fig. B.1 (black) and Fig B.3 (grey)

Figure B.3. Waveforms of the raw (unfiltered, no instrument response removed) Z-component seismograph traces for all SAMBA, DFDP10 and GeoNet stations. The teleseismic earthquakes which produced these onsets are the $M_W=6.2$ earthquake of 15 February 2010 (21:51:47.8 GMT, lat= -7.22° , lon= 128.72° , depth=126 km) and the $M_W=6.1$ earthquake of 8 March 2010 (09:47:11.6 GMT, lat= 19.33° , lon= 144.71° , depth=456 km)

Figure B.4. Waveforms of the Z-component traces of all SAMBA and GeoNet stations which recorded a quarry blast near station WHAT. Clear onsets are upwards (compressional) for all stations. Waveforms of the vertical seismograph traces of all SAMBA, DFD P10 and GeoNet stations that recorded a quarry blast in a similar location to that in Fig. B.4. Clear onsets for the DFD P10 stations are downwards (reversed polarity).

Date	Time (UTC)	Latitude	Longitude	Depth [km]	M _W
4 October 2009	10:58	6.77°	123.56°	626	6.6
28 August 2009	01:51	−7.09°	123.46°	634	6.9
15 February 2010	21:51	−7.22°	128.72°	126	6.2

Table B.1. Origin time and hypocentre locations from the USGS Preliminary Determination of Epicentres (PDE) of teleseismic earthquakes with impulsive P-arrivals recorded by the SAMBA/DFDP10 stations.

C. Station recording times

Figure C.1 displays the recording time in days for the SAMBA and DFDP10 stations. Note that for GeoNet stations FOZ, WVZ and RPZ only those days from Julian day 125 of 2010 onwards are shown which were analysed. Horizontal lines indicate 1 January of each year for orientation. Small numbers above each trace indicate Julian days when recording began or stopped. See Section 1.4.1 for station names and locations.

Figure C.1. (*facing page*) Station recording time of the SAMBA, GeoNet and DFDP10 stations between November 2008 and December 2010. Small numbers above the traces indicate when the recording began or stopped.

C. Station recording times

D. Noise signals

In this appendix some of the spurious signals caused by instrumentation or other identifiable noise sources are listed. This is by no means a complete catalogue of the recorded noise signals, but it may be useful as a reference in future seismicity studies. The reason why these signals are so prominent in the SAMBA data may be due the high gain setting.

D.1. Solar controller

The data recorded by the SAMBA stations exhibits prominent spikes, although the same equipment and station setup have been used before in other VUW seismicity studies. The regular occurrence of these spikes was found to be strongly related to sunshine hours and therefore the source of these spikes could be identified in the power system. Solar controllers apply pulse-modulation-width charging to recharge the battery. Different solar controllers were used at the SAMBA stations (Table D.1), which caused different spikes in the data (Fig. D.1).

Solar controller	Spike generation	Recurrence period of spikes [s]	Characteristics/ comments
SunSaver 6	yes	37 s	amplitude variation depending on sun intensity
SunSaver 10	yes	37 s	
ProStar-15	yes	6 min	double spike up down with 1.3 s spacing in between
ProStar-30	yes	6 min	
Steca Solsum	yes	1.5 s	low in amplitude
Powertech	no		devices failed after less than one year, cannot be purchased in New Zealand
Flexcharge NC 12	no		

Table D.1. Solar controller used in this study and their recurrence period and characteristics of spikes generated.

D.2. Other noise sources

Other anomalous signals that have been observed are shown in Figures D.2, D.3 and D.4. When water leaked into the box containing the solar controller, sudden one-sided amplitude changes (either positive or negative) were observed (Fig. D.2). A frequency and amplitude change of the signal was observed on one channel at station WHAT, which was found to have a “dry” soldering joint. Similar signals were observed on the N-components of DFDP10 station BLO and GeoNet station WVZ. These channels recorded only the high frequencies of the signal and the amplitude of the signal amounts to only 10% of that of the other horizontal component. At station REYN, the E-component, which turned out not to be grounded at the sensor, also showed a reduced amplitude/amplification and is characterised by jumps in the signal.

Lightning and thunder, as shown in Fig. D.3, can be seen on all stations except those installed in boreholes. This signal is characterised by a spike followed by an emergent noisy signal with varying time delay according to the distance from the storm. These records usually show noisier traces associated with wind and rain.

Internal noise, identified at station COSA, resembles the spikes of unknown origin at stations COSA and COVA but persisted when the sensor and GPS cables were removed from the Reftek. On closer inspection, these spikes were found to occur every second with a duration of 0.1 s and always show the same form and amplitude. The recorded signal of a helicopter passing by the stations lasts for a few minutes and typically oscillates with a period of 0.05 s. The signal resembles a beating wave. An electric fence recorded on a borehole sensor generates a “ringing” spike every 1.3 s with only a positive or negative amplitude. In this case, the steel casing of the borehole acted as an antenna transferring the signal to the sensor. Grounding of the casing prevented further recording of this signal. The amplitude of the spike shows a variability associated with the current. After a new fencing system was installed at the property on which the sensor is located, the spikes were much larger in amplitude.

Figure D.3. Noise generated by lightning and thunder (top) and internal noise generated by the Reftek (bottom).

D. Noise signals

Figure D.4. Noise generated by helicopter (first row) and electric fence (second and third row). The figures in the right column give a closeup view of those in the left column

E. Assessment of individual station performance

The following two tables (E.1 (a) and (b)) discuss the overall performance of the SAMBA and DFDPI0 sites, noise sources near the sensors and other problems that occurred. For the mountain sites requiring helicopter access, brief comments regarding the possibility of telemetry are added.

Station name	Performance and problems
COSA	good recording during night times, but noisy during the day due to road traffic, sensor not accessible any more, since site was covered by at least 5 m of sediments, but no equipment damage, internal noise
EORO	site close to earthquake swarms and low noise levels for high frequencies, sometimes noisier due to change in course of nearby stream, faulty N-channel, site accessibility difficult (15 min walk), sunshine hours marginal in winter
MTFO*	continuously recording site, little snow, intensive kea damage, telemetry possible (line of sight and cell phone reception)
WHAT	noisy during daytime due to machinery used in nearby quarry, recorded several quarry blasts, damage to solar panel holder due to strong wind, once cows in quarry
WHYM*	reliable, continuously recording site, close to earthquake cluster, clear waveforms although in a boulder of a landslide, full sun and little snow, telemetry problematic (in valley), visible from walking track
POCR	performance ok, due to low-period noise, but good in addition to station FRAN, electric fence noise, cows
LABE*	site recorded many avalanches, lots of snow, but solar panel free of snow due to strong winds, initial solar panel setup blown away, kea damage, telemetry possible (line of sight and cell phone reception)
GOVA	reliable, continuously recording site, telemetered to lakeside, clear simple waveforms, low noise levels, recorded all earthquakes even from large distances
FRAN	good site, sometimes noise from rock crusher in riverbed nearby when worked on
REYN*	problematic/poor site, one broken channel, very emergent signal, site covered by snow in winter, landslide hazards, removed in September 2010 due to flat batteries, telemetry possible to upper Whataroa river
COVA*	site performance ok, site covered in snow during winter, area too sheltered

Table E.1. (a) Observations of site conditions and assessment of the station performance. Sites marked with star require helicopter access.

Station name	Performance and problems
BLO*	close to large cluster of earthquakes, site directly on solid rock, however, no soil to bury sensor, strong wind and lots of snow, sensor frozen to ground in autumn, faulty E-component, kea damage
BON*	good site, few places to put sensor because of steep topography and cell phone towers, no outcrop because of vegetation, sensor buried in soil, kea damage
DRC	sensor buried in moraine on farm, noisy, likely to have cows/deer on field, swampy, little sun, GPS had problems getting satellite contact
ERE	noisy due to nearby farm, sensor buried in soil, waveforms emergent
GCK	sensor buried in river gravels, performance alright, sometimes noisy (woodworks), now several sensors installed at DFDP1 boreholes
GHU*	good site, sensor buried in moraine, important site because close to earthquakes, kea damage
NOL*	site performed better than expected, despite in moraine, no bedrock in area close by, kea damage
ONE	noisy but signal not too bad for farmland, huge station term of 0.7s
POE	site next to farm track in river gravels, sometimes noisy
VBV*	buried in moraine close to bivouac, quiet site, sheltered valley, little sun, maybe snow in winter, kea damage
WHB	in river gravels on farm, cows, noisy, picks up quarry blasts
WNQ	sensor buried in gravels, little quarry noise, little sun

Table E.1. (b) Assessment of the performance of DFDP10 seismometer sites. Sites marked with star require helicopter access.

F. Noise characteristics of the SAMBA stations

The analysis of power spectral densities shows that the N-component of the station EORO, the E-component of the station REYN and the first horizontal component of WHAT are faulty. These channels do not properly record frequencies below their natural frequencies of 2 Hz. This has been recognised and fixed at station WHAT which had a “dry” soldering joint. For station REYN this problem could not be fixed as this channel appears to be ungrounded at the sensor (see also section D.2). For EORO, the malfunction of the channel is not particularly obvious in seismograms and has not been addressed yet.

Figure F.1. Power spectral density for “low” (left) and “high” (right) noise conditions for the deep SAMBA stations, see Fig. 1.10 (c) for more detail

Figure E.2. (a) Power spectral density for “low” (left) and “high” (right) noise conditions at the GeoNet stations, see Fig. 1.10 (c) for details.

G. Travel-time residuals of SAMBA and DFDP10 stations

Figure G.1. (a–d, following pages) Travel-time residuals for P (blue) and S-waves (red) versus event number for the combined SAMBA/GeoNet stations between 24 March and December 2009 when the SAMBA station network was complete and for the DFDP10 stations between January 2010 to April 2010. All values were reduced by the median of the distribution. Dashed red and blue lines show the range of one and two standard deviations. Travel-time residuals larger than two standard deviations were considered as outliers.

(a) Travel-time residuals between 24 March and December 2009 for the SAMBA stations. Note that due to data recording problems (described in Section 1.4.3) not all stations were recording in the beginning. For station COSA, residuals were corrected as described in Section 3.1.3

(b) Travel-time residuals between 24 March and December 2009 for SAMBA and GeoNet stations.

H. Comparisons of focal mechanism solutions

Figure H.1. (a)
P-wave polarities
and focal mechanism
solutions for earth-
quakes within region
4. See Fig. 3.22 (a)
for details.

Figure H.1. (b)
Probability density
contours of P- (or-
ange) and T-axes
(green) of the focal
mechanism solution
for earthquakes
within region 4.

Figure H.2. (a) P-wave polarities and focal mechanism solutions for earthquakes within region 6. See Fig. 3.22 (a) for details.

I. Slip, frequency, rupture and recording range of various-sized earthquakes

Moment Magnitude M_W	Average slip $\bar{\mu}$	Rupture length or width L or W	Rupture area $A \sim L \times W$	Relative frequency	Recording range
8	4 m	100 km (650 km/15 km)	10^4 km^2	Nyr^{-1}	$\sim 20 \text{ km}$
7	1 m	30 km (60 km/15 km)	10^3 km^2	10 N yr^{-1}	
6	40 cm	10 km	18 km^2	10^2 N yr^{-1}	
5	10 cm	3 km	10 km^2	10^3 N yr^{-1}	
4	4 cm	1 km	1 km^2	10^2 N yr^{-1}	
3	1–4 cm	300 m	10^5 km^2	10^5 N yr^{-1}	
2	0.4–1 cm	100 m	10^4 m^2	10^6 N yr^{-1}	
1	1–4 mm	30 m	10^3 m^2	10^7 N yr^{-1}	
0	0.4–1 mm		30 m^2	10^8 N yr^{-1}	
–1	0.1–0.4 mm		3 m^2	10^9 N yr^{-1}	
–2	0.04–0.1 mm		0.3 m^2	$10^{10} \text{ N yr}^{-1}$	$>3 \text{ km}$
–3	10–40 μm		0.03 m^2	$10^{11} \text{ N yr}^{-1}$	1.5 km
–4	$\leq 10 \mu\text{m}$		0.003 m^2	$10^{11} \text{ N yr}^{-1}$	0.75 km
					0.45 km
					$\leq 30 \text{ m}$

Table I.1. Source dimensions, frequency of occurrence and recording range for earthquakes of different magnitude. Table modified from Sibson (1989) and extended to earthquakes with $M_W < 0$

		Stress (MPa)		Period
Solid Earth tides		10^{-3}		12 h
Ocean tides		10^{-2}		12 h
Hydrological loading		10^{-3} – 10^{-1}		days–years
Glacier loading		10^1 – 10^2		10^3 years
	10^2 km	10^3 km	10^4 km	
Static stress changes, M8	10^{-1}	10^{-4}	10^{-7}	NA
Dynamic stress changes, M8	3	0.06	0.001	20 s

Table I.2. Table 2 from Manga and Brodsky (2006) showing amplitudes and periods of stress changes from different sources compared with static and dynamic stresses imposed by earthquakes.

J. Hypocentre locations

Hypocentre locations for the period November 2008 to December 2009 are available from the supplementary material Table 1 of Boese et al. (2012). The remaining events from January to April 2010 are listed below.

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100101	2:16	48.89	-43.414	170.236	4.52	0.03	3	6	264	-0.04
20100101	8:58	12.28	-43.532	170.576	-1.86	0.08	5	11	238	0.95
20100101	8:58	12.13	-43.538	170.575	5.14	0.10	5	12	238	0.95
20100101	17:06	52.01	-43.551	170.579	6.08	0.15	11	24	101	3.01
20100101	17:07	4.61	-43.556	170.583	7.60	0.14	8	21	101	2.71
20100101	21:21	21.24	-43.956	169.752	23.03	0.09	4	12	306	2.31
20100102	16:34	26.48	-43.262	170.754	8.47	0.17	6	19	158	2.48
20100104	1:08	25.32	-43.248	170.374	2.02	0.13	6	16	297	1.85
20100104	10:02	26.34	-43.262	170.752	8.57	0.20	5	18	157	2.86
20100105	1:02	56.23	-43.146	170.921	11.42	0.17	5	15	233	1.96
20100105	3:03	41.65	-43.330	170.802	9.27	0.11	5	14	165	1.78
20100105	5:19	55.09	-43.581	169.880	9.27	0.18	7	17	218	2.16
20100105	6:02	56.09	-43.588	169.880	7.88	0.13	5	13	222	1.33
20100105	6:55	8.35	-43.587	169.863	10.03	0.12	5	13	230	1.39
20100105	8:23	44.65	-43.467	170.372	0.08	0.09	1	9	222	NA
20100105	12:01	22.71	-43.587	169.874	9.41	0.20	8	19	224	1.65
20100105	12:01	33.85	-43.604	169.870	7.92	0.09	2	7	235	1.49
20100105	12:13	34.55	-43.585	169.877	9.41	0.11	5	14	222	1.20
20100105	13:08	12.39	-43.584	169.882	7.67	0.12	7	17	218	1.70
20100105	13:09	43.42	-43.591	169.870	8.92	0.11	6	14	229	1.24
20100105	15:19	8.56	-43.584	169.881	8.99	0.16	7	17	219	1.64
20100106	4:35	53.77	-43.584	169.878	8.78	0.12	5	13	220	1.60
20100106	6:16	54.42	-43.382	170.283	5.28	0.04	3	7	212	0.76
20100107	13:04	44.24	-43.591	170.623	7.95	0.14	10	22	129	2.36
20100107	16:29	28.90	-43.597	170.617	8.61	0.12	7	18	132	1.71
20100108	8:18	1.38	-43.467	170.277	1.78	0.12	8	18	67	0.87
20100108	9:27	19.38	-43.562	170.385	2.92	0.13	9	19	141	1.37
20100108	19:31	44.31	-43.585	170.251	10.00	0.10	2	5	172	1.10
20100108	20:34	24.48	-43.692	170.043	12.63	0.02	2	8	232	1.60
20100109	0:25	29.32	-43.528	170.055	3.06	0.02	2	6	240	NA
20100110	15:37	15.35	-43.542	170.297	2.61	0.07	5	14	140	0.93
20100110	20:42	44.22	-43.254	170.765	10.41	0.12	5	11	162	2.00

J. Hypocentre locations

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100110	20:54	1.53	-43.456	170.400	6.42	0.18	7	17	125	1.85
20100111	4:38	0.60	-43.888	169.736	13.18	0.10	6	14	303	1.96
20100111	11:44	34.81	-43.592	170.238	2.68	0.14	10	20	176	1.94
20100117	22:39	1.77	-43.377	169.772	57.56	0.12	4	8	318	NA
20100112	11:34	47.38	-43.454	171.061	13.84	0.16	10	21	213	2.35
20100113	16:51	43.51	-43.569	169.860	8.50	0.12	6	13	218	1.28
20100114	5:27	40.36	-43.538	170.590	6.56	0.11	10	21	104	1.42
20100115	21:34	47.72	-43.418	170.174	5.85	0.11	10	21	94	2.15
20100116	11:02	31.26	-43.484	170.355	6.03	0.17	11	23	86	1.95
20100117	11:41	13.41	-43.389	170.737	5.66	0.25	10	22	144	1.87
20100117	22:39	1.77	-43.377	169.772	57.56	0.12	4	8	318	1.45
20100118	2:40	44.41	-43.385	170.735	10.69	0.14	2	7	213	1.28
20100118	9:58	15.59	-43.818	169.657	15.13	0.11	6	14	308	1.34
20100118	17:23	27.02	-43.831	169.641	16.93	0.15	3	11	309	1.19
20100119	3:53	1.02	-43.821	169.655	11.24	0.08	6	16	308	2.23
20100119	8:20	47.06	-43.833	169.651	10.58	0.16	9	21	298	3.12
20100119	10:41	14.58	-43.546	170.284	-1.98	0.12	6	15	148	1.22
20100120	12:11	20.52	-43.455	170.345	2.19	0.05	6	12	93	1.00
20100120	12:11	23.98	-43.455	170.353	3.34	0.09	3	10	104	1.00
20100120	17:30	3.21	-43.536	170.610	-1.93	0.07	5	10	281	1.59
20100121	14:16	56.85	-43.824	169.734	17.90	0.16	7	15	299	1.50
20100121	20:12	29.58	-43.825	169.744	15.96	0.09	6	14	298	1.80
20100121	23:44	7.36	-43.472	170.227	10.38	0.10	8	17	86	1.07
20100121	23:44	7.38	-43.470	170.227	10.17	0.08	7	16	85	1.07
20100122	1:29	27.54	-43.262	170.376	0.01	0.08	3	8	293	1.09
20100122	7:38	32.49	-43.393	170.742	3.75	0.22	4	12	145	1.36
20100122	9:12	20.95	-43.456	170.352	3.22	0.14	8	18	104	1.26
20100123	9:50	39.59	-43.879	170.598	14.95	0.12	7	17	246	1.62
20100123	9:52	25.60	-43.256	170.862	7.46	0.23	7	19	190	1.82
20100123	16:38	8.19	-43.551	169.992	5.18	0.09	5	13	94	0.68
20100124	4:43	19.50	-43.496	170.344	2.54	0.07	5	12	121	0.64
20100124	5:02	0.03	-43.291	170.908	8.05	0.06	2	7	250	1.35
20100124	6:31	46.38	-43.310	170.941	10.45	0.17	8	20	201	1.80
20100124	8:46	29.88	-43.250	170.859	-1.97	0.21	6	18	190	2.43
20100124	22:03	37.72	-43.289	170.918	5.94	0.27	6	17	199	1.65
20100125	2:45	45.99	-43.298	170.819	-1.93	0.14	4	8	228	1.30
20100125	3:26	19.54	-43.646	170.111	4.62	0.12	5	12	199	0.85
20100125	17:50	27.97	-43.479	170.278	1.71	0.09	5	13	78	1.06
20100125	18:17	20.77	-43.247	171.335	10.00	0.22	2	7	272	2.01
20100125	19:35	11.35	-43.504	170.356	5.14	0.04	4	9	131	0.82
20100125	22:48	37.95	-43.474	170.266	-1.97	0.11	6	15	76	1.09
20100126	9:53	33.24	-43.547	169.990	5.89	0.14	6	16	99	1.45
20100126	20:11	12.14	-43.581	169.880	7.67	0.08	4	9	198	0.92
20100127	3:16	2.34	-43.278	170.362	-0.01	0.23	5	14	176	NA
20100127	11:57	36.88	-43.144	170.881	7.60	0.16	4	16	225	1.82
20100127	20:30	30.08	-43.170	170.757	2.85	0.00	1	4	244	1.31
20100128	9:14	28.53	-43.572	171.136	14.81	0.09	6	15	232	1.71

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100128	16:07	42.46	-43.594	170.623	9.11	0.28	13	26	131	2.37
20100128	16:35	22.58	-43.130	170.916	12.08	0.17	2	14	238	2.12
20100128	17:44	24.87	-43.549	170.240	6.15	0.11	6	15	157	1.19
20100129	1:53	44.90	-43.373	170.804	-0.96	0.05	2	6	161	1.37
20100129	17:47	55.27	-43.493	170.324	4.43	0.09	8	18	104	1.14
20100130	2:29	40.27	-43.492	170.323	4.27	0.10	7	16	103	1.21
20100130	2:29	46.93	-43.491	170.322	4.34	0.10	3	11	102	0.69
20100130	2:40	23.51	-43.492	170.322	3.79	0.09	5	13	103	1.08
20100130	2:42	14.05	-43.494	170.323	4.36	0.11	7	17	104	1.25
20100130	4:00	39.45	-43.493	170.323	4.50	0.10	8	17	103	1.15
20100130	7:51	25.04	-43.038	171.213	21.09	0.21	7	19	286	2.44
20100130	10:58	35.79	-43.504	170.328	3.23	0.05	5	13	109	1.06
20100130	10:58	39.57	-43.503	170.328	3.58	0.06	5	15	109	1.06
20100130	14:49	8.03	-43.504	170.329	4.08	0.12	8	18	110	1.15
20100130	14:59	15.57	-43.457	170.395	7.05	0.21	9	19	123	1.77
20100131	2:29	33.62	-43.442	170.126	5.00	0.19	8	19	118	1.10
20100131	11:47	36.61	-43.425	170.053	8.45	0.11	10	23	177	2.44
20100131	13:04	17.68	-43.418	170.713	2.54	0.18	10	21	137	1.75
20100201	4:24	50.68	-43.266	170.361	0.95	0.19	3	12	292	0.74
20100201	19:19	8.14	-43.206	170.636	-1.97	0.19	4	13	145	2.15
20100201	19:19	8.19	-43.200	170.625	-1.97	0.20	7	17	146	2.15
20100201	19:27	37.39	-43.248	170.561	25.87	0.03	3	7	299	1.78
20100201	22:25	58.35	-43.220	170.846	12.84	0.02	2	5	239	1.69
20100201	22:39	55.01	-43.970	169.939	15.68	0.13	5	14	297	1.88
20100201	23:25	58.33	-43.202	170.628	-1.93	0.08	3	8	159	1.78
20100202	1:28	22.88	-43.239	170.885	12.80	0.13	4	12	199	2.12
20100202	2:55	58.73	-43.460	170.233	7.33	0.06	6	12	83	0.99
20100202	4:26	44.85	-43.462	170.229	8.09	0.13	6	13	80	1.74
20100202	9:01	17.39	-43.582	170.623	8.00	0.21	12	24	123	3.54
20100202	10:09	5.20	-43.258	170.752	7.05	0.23	5	13	157	1.46
20100202	20:55	18.70	-43.474	170.541	9.61	0.20	9	20	103	2.11
20100202	20:59	55.58	-43.475	170.533	7.22	0.10	3	8	258	1.35
20100202	21:11	33.23	-43.470	170.551	8.99	0.16	6	14	172	1.64
20100203	3:38	28.59	-43.314	170.828	7.22	0.01	2	6	173	1.43
20100203	11:14	39.37	-43.577	170.614	9.27	0.13	8	20	119	1.73
20100203	14:35	35.56	-43.494	170.420	2.54	0.11	7	15	169	1.02
20100203	16:33	19.10	-43.431	170.385	3.15	0.16	7	17	164	1.13
20100203	23:21	46.90	-43.169	170.797	11.24	0.02	2	5	229	1.47
20100204	3:38	35.18	-43.234	170.763	0.01	0.03	2	5	211	1.60
20100204	10:48	9.04	-43.344	170.330	-1.97	0.16	9	20	114	1.59
20100204	13:36	24.52	-43.469	170.549	7.19	0.18	6	16	104	1.39
20100204	16:30	14.67	-43.413	170.713	-1.97	0.15	5	14	137	1.73
20100204	19:29	56.51	-43.473	170.541	7.22	0.07	2	8	223	1.18
20100205	0:37	41.23	-43.501	170.352	27.40	0.00	2	4	360	0.66
20100205	7:01	39.10	-43.464	170.552	6.08	0.16	6	16	105	1.46
20100205	14:27	56.04	-43.501	170.510	-1.97	0.11	5	12	208	1.05
20100205	23:07	30.08	-43.296	170.782	-1.97	0.23	5	12	163	1.57

J. Hypocentre locations

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100206	8:44	55.21	-43.268	170.807	-1.97	0.22	7	16	225	2.26
20100206	8:54	0.38	-43.264	170.794	-1.97	0.21	7	18	170	1.72
20100206	8:54	55.67	-43.265	170.792	-1.97	0.22	5	13	169	1.61
20100206	9:48	37.70	-43.261	170.792	6.67	0.21	6	17	169	1.93
20100206	11:55	3.56	-43.262	170.803	-1.97	0.25	5	16	172	1.71
20100206	18:56	52.56	-43.270	170.809	-1.93	0.19	6	17	226	2.28
20100206	19:04	18.05	-43.270	170.815	-1.97	0.19	7	17	227	2.05
20100207	6:58	31.35	-43.268	170.807	-1.97	0.21	4	8	225	1.27
20100207	19:48	22.56	-43.257	170.763	8.19	0.06	3	6	212	1.46
20100207	22:26	1.52	-43.202	170.894	19.56	0.03	1	6	210	1.86
20100208	7:19	52.89	-43.536	170.668	6.11	0.08	4	12	270	1.57
20100208	19:41	20.08	-43.706	170.035	9.34	0.10	5	12	238	1.05
20100209	4:21	17.63	-43.543	169.971	7.41	0.13	9	19	119	1.66
20100209	5:33	19.63	-43.546	169.972	8.64	0.09	4	9	143	0.59
20100209	5:33	35.99	-43.543	169.974	7.33	0.11	5	11	118	0.83
20100209	8:26	11.89	-43.508	170.503	7.41	0.20	10	22	95	2.01
20100209	8:50	46.78	-43.498	170.286	3.79	0.22	11	23	79	2.43
20100209	11:23	30.94	-43.542	169.973	7.40	0.15	8	17	119	1.21
20100209	11:42	9.76	-43.542	169.974	7.64	0.16	10	20	119	1.77
20100209	13:48	17.30	-43.430	170.683	9.16	0.06	3	6	269	1.43
20100209	14:57	1.48	-43.270	170.806	-1.97	0.19	5	12	225	1.37
20100209	14:57	47.18	-43.269	170.831	-1.93	0.12	2	7	232	1.29
20100209	19:39	27.97	-43.310	170.905	10.55	0.09	6	16	192	2.00
20100209	19:57	9.12	-43.259	170.840	8.30	0.19	5	16	183	2.54
20100210	6:28	1.46	-43.171	170.341	18.80	0.02	2	4	265	1.57
20100211	3:39	58.53	-43.155	170.928	11.17	0.20	1	9	231	2.08
20100211	18:58	7.38	-43.311	170.362	9.79	0.06	2	4	195	1.29
20100211	20:05	16.78	-43.449	170.241	6.89	0.14	6	14	71	1.13
20100211	21:50	10.58	-43.138	170.948	11.97	0.26	4	15	241	2.36
20100212	13:32	59.74	-43.207	170.595	6.94	0.09	1	11	151	1.74
20100213	2:59	14.24	-43.203	170.640	4.17	0.11	4	10	163	1.56
20100213	11:36	45.31	-43.417	170.711	-1.97	0.18	9	21	136	2.54
20100213	19:32	13.17	-43.584	170.620	5.73	0.21	10	22	124	1.97
20100214	1:04	26.64	-43.152	170.901	7.01	0.23	4	8	226	1.75
20100214	7:27	43.41	-43.535	170.425	3.69	0.09	6	14	161	1.24
20100214	8:41	32.43	-43.460	170.285	6.70	0.14	4	13	65	1.09
20100214	13:26	0.92	-43.209	170.866	-0.34	0.01	1	4	246	1.13
20100214	23:42	35.65	-43.172	170.787	10.90	0.01	2	5	181	1.66
20100215	9:40	30.90	-43.257	170.821	6.60	0.15	3	7	229	1.24
20100215	13:30	29.97	-43.525	170.690	-1.97	0.15	7	17	121	1.68
20100215	15:56	56.70	-43.528	170.302	4.83	0.12	8	18	118	1.25
20100216	11:22	7.21	-43.134	170.971	12.73	0.32	5	16	246	1.89
20100216	20:48	17.56	-43.777	170.510	3.96	0.03	2	4	322	1.01
20100216	21:02	45.67	-43.401	170.746	-1.97	0.16	5	11	217	1.54
20100217	0:32	32.60	-43.425	170.218	8.30	0.15	7	14	133	0.99
20100217	1:09	12.64	-43.403	170.779	-1.97	0.16	8	18	152	1.88
20100217	2:10	51.29	-43.155	170.935	10.00	0.22	3	9	233	1.82

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100217	5:57	26.01	-43.128	170.964	15.26	0.23	3	10	246	1.83
20100217	7:21	29.56	-43.152	170.929	12.08	0.19	2	12	233	2.63
20100217	7:28	57.28	-43.155	170.923	11.94	0.17	2	13	230	3.15
20100217	7:31	38.24	-43.147	170.937	15.06	0.05	2	5	277	1.31
20100217	7:55	29.53	-43.159	171.003	-0.96	0.08	3	6	284	1.57
20100217	8:10	49.08	-43.301	170.810	-1.97	0.23	2	7	226	1.50
20100217	8:29	55.82	-43.437	170.384	5.02	0.10	7	15	163	1.27
20100217	8:33	26.74	-43.169	170.986	-1.93	0.21	5	13	279	2.21
20100217	9:02	54.61	-43.720	170.649	9.13	0.11	2	9	204	1.62
20100217	10:03	59.29	-43.150	171.029	8.05	0.21	5	15	289	2.53
20100217	14:32	53.06	-43.266	170.790	-1.97	0.22	5	10	168	1.84
20100218	6:26	40.51	-43.264	170.797	-1.93	0.16	4	10	170	1.93
20100218	11:02	48.48	-43.417	170.714	-1.97	0.19	7	15	211	NA
20100218	2:02	53.01	-43.552	170.578	6.39	0.08	3	8	264	NA
20100219	2:02	52.99	-43.552	170.580	6.39	0.08	3	8	265	1.46
20100219	10:03	36.96	-43.505	170.690	8.12	0.06	3	9	272	1.36
20100219	17:58	20.80	-43.565	170.532	6.70	0.10	9	19	135	1.73
20100219	23:04	6.56	-43.564	170.223	6.42	0.06	4	10	177	0.77
20100220	2:31	30.94	-43.171	170.888	8.19	0.21	3	13	217	3.27
20100221	2:27	26.17	-43.421	170.701	9.34	0.19	7	14	134	1.55
20100221	15:59	12.22	-43.034	171.215	12.25	0.28	6	16	287	2.29
20100221	18:08	32.78	-43.385	170.310	4.76	0.13	4	11	122	1.02
20100221	18:08	38.92	-42.912	171.447	6.84	0.37	5	15	307	3.27
20100222	5:59	9.41	-43.507	170.501	6.84	0.22	8	18	95	1.95
20100222	6:00	41.76	-43.506	170.509	4.83	0.12	6	13	207	1.47
20100222	10:47	56.58	-43.253	170.866	-1.97	0.24	8	16	191	1.65
20100222	11:41	12.70	-43.503	170.513	-1.97	0.12	5	12	209	1.16
20100222	13:25	40.09	-43.384	170.313	6.22	0.13	5	12	125	1.20
20100222	20:26	57.67	-43.171	170.868	8.75	0.23	3	9	211	1.90
20100223	7:30	1.32	-43.616	169.779	9.58	0.13	5	12	317	1.72
20100223	11:49	7.11	-43.252	170.789	-1.93	0.14	3	8	169	1.58
20100223	22:22	0.00	-43.304	170.501	0.98	0.05	3	7	256	NA
20100223	22:25	8.40	-43.282	170.362	-0.01	0.12	2	10	111	NA
20100224	22:17	49.07	-43.550	170.672	9.23	0.02	2	5	210	1.63
20100225	3:42	21.61	-43.358	170.615	-1.93	0.14	3	7	178	1.60
20100225	11:53	18.49	-43.671	171.113	9.61	0.06	6	15	247	2.43
20100225	14:04	34.43	-43.363	170.799	-1.97	0.20	5	12	161	1.51
20100226	1:30	16.01	-43.505	170.503	6.29	0.20	8	17	96	2.30
20100226	20:08	46.85	-43.537	170.609	5.56	0.08	5	10	252	1.33
20100226	20:52	58.13	-43.411	170.441	5.04	0.13	5	10	193	1.05
20100227	6:20	23.82	-43.576	170.640	11.42	0.09	3	11	149	1.69
20100227	6:41	13.03	-43.578	170.639	12.04	0.08	3	10	148	1.61
20100227	8:06	15.64	-43.092	170.789	12.04	0.26	7	18	230	2.81
20100227	18:55	11.76	-43.571	170.639	9.30	0.08	5	10	270	1.47
20100228	3:35	39.34	-43.256	170.875	-1.93	0.11	3	7	244	1.31
20100228	5:51	39.85	-43.104	170.804	10.55	0.08	1	6	263	1.65
20100228	10:18	35.28	-43.095	170.770	11.62	0.18	7	18	214	2.69

J. Hypocentre locations

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100228	16:35	11.81	-43.243	170.883	-1.97	0.18	7	17	198	2.34
20100228	16:40	32.99	-43.249	170.897	-1.93	0.09	3	7	250	1.64
20100228	17:53	22.84	-43.085	170.822	12.08	0.12	3	9	285	1.77
20100228	18:20	59.02	-44.033	170.103	15.54	0.14	9	19	272	2.76
20100228	20:24	8.48	-43.537	170.712	8.47	0.01	2	6	170	1.75
20100228	23:42	59.59	-43.544	171.118	11.83	0.14	7	16	225	2.27
20100301	2:56	22.63	-43.296	170.827	-1.93	0.14	3	8	175	1.76
20100301	13:41	43.92	-43.172	171.048	12.28	0.09	2	5	287	2.01
20100301	13:43	23.59	-43.168	171.017	17.34	0.12	3	7	284	2.15
20100301	18:12	31.84	-43.246	170.875	-1.93	0.12	2	6	245	1.60
20100302	7:39	32.52	-43.169	170.857	8.23	0.24	6	16	208	2.43
20100302	12:43	14.89	-43.266	170.858	-1.93	0.15	2	6	187	1.65
20100302	12:57	28.79	-43.236	170.832	-0.96	0.03	2	5	234	1.44
20100302	13:32	37.22	-43.592	170.620	8.61	0.17	11	22	129	2.89
20100302	21:50	21.88	-43.046	170.900	6.94	0.14	3	7	275	2.11
20100302	22:23	53.45	-43.492	170.339	5.80	0.19	8	16	116	2.01
20100303	23:24	55.10	-43.493	170.341	5.75	0.18	6	15	117	1.38
20100304	4:45	43.47	-43.798	170.653	8.02	0.16	10	21	216	2.30
20100304	6:50	38.98	-43.544	170.546	4.17	0.13	6	14	225	2.09
20100305	19:57	56.51	-43.609	170.435	4.21	0.13	8	18	147	1.74
20100305	22:28	37.03	-43.491	170.273	2.89	0.21	6	16	81	2.27
20100305	22:49	3.77	-43.494	170.277	3.34	0.21	9	20	72	2.35
20100307	1:15	6.18	-43.400	170.703	10.41	0.13	5	11	275	1.97
20100307	6:30	58.60	-43.451	170.639	11.10	0.03	2	6	257	0.85
20100307	7:06	38.02	-43.444	170.654	10.03	0.17	4	12	123	1.42
20100307	20:38	0.06	-43.401	170.779	-1.97	0.19	6	15	153	NA
20100307	23:41	41.39	-43.244	170.863	8.75	0.18	4	10	192	1.60
20100308	19:43	35.21	-43.254	170.890	-1.93	0.18	5	12	248	1.91
20100308	21:23	45.31	-43.155	171.144	13.01	0.29	5	16	262	2.59
20100312	7:09	7.47	-43.485	170.599	5.31	0.21	8	19	111	2.04
20100312	10:18	21.05	-43.153	170.871	6.94	0.24	1	8	218	1.64
20100312	22:03	29.22	-43.188	170.855	11.73	0.20	2	7	202	1.50
20100312	23:48	30.63	-43.550	170.596	8.02	0.11	7	16	148	1.51
20100313	6:57	38.66	-43.047	170.882	8.61	0.18	2	10	275	2.35
20100313	8:18	2.42	-43.308	170.786	-1.86	0.15	3	7	220	0.96
20100313	13:12	13.10	-43.825	170.723	14.05	0.19	6	15	225	1.96
20100314	1:23	20.25	-43.161	170.911	10.83	0.18	2	12	226	2.50
20100314	11:45	43.66	-43.254	170.592	-1.97	0.24	4	14	129	1.44
20100314	19:15	17.57	-43.268	170.829	-1.93	0.18	3	9	179	1.61
20100314	19:33	38.67	-43.198	170.530	11.17	0.00	2	4	228	1.01
20100314	22:33	12.14	-43.138	170.884	15.92	0.20	2	9	228	1.84
20100315	7:37	4.80	-43.436	170.401	4.22	0.20	7	15	170	0.85
20100315	9:52	7.92	-43.143	170.895	13.29	0.24	3	11	229	1.26
20100317	15:10	51.57	-43.310	170.826	-1.97	0.17	3	8	173	1.09
20100317	15:48	22.43	-43.274	170.750	-1.97	0.22	4	11	208	1.24
20100318	9:19	45.13	-43.308	171.490	22.82	0.01	1	5	285	2.61
20100318	10:33	31.11	-43.186	170.883	-1.93	0.06	3	10	255	1.93

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100318	16:03	32.61	-43.192	170.971	18.04	0.11	1	6	229	1.44
20100319	6:14	5.99	-43.149	170.792	-1.45	0.09	2	8	232	1.32
20100319	9:55	31.56	-43.547	170.558	10.31	0.17	7	17	99	1.23
20100319	10:20	27.01	-43.507	170.650	5.14	0.17	10	21	117	1.46
20100319	10:50	6.28	-43.538	170.570	7.67	0.18	7	16	102	1.32
20100319	12:48	26.82	-43.542	170.566	10.19	0.25	9	20	101	1.55
20100319	15:05	37.13	-43.538	170.617	3.75	0.06	3	11	155	1.10
20100319	19:08	6.33	-43.261	170.599	-1.10	0.10	2	8	162	1.04
20100319	21:13	26.15	-43.402	170.780	-1.97	0.12	4	10	153	1.20
20100319	21:36	49.10	-43.438	170.711	-1.97	0.17	8	17	135	1.92
20100320	17:28	39.39	-43.196	170.974	16.89	0.20	6	16	229	2.41
20100320	18:22	38.73	-43.196	170.994	17.07	0.00	2	5	276	1.69
20100321	23:42	45.46	-43.101	170.955	14.26	0.20	3	12	255	2.91
20100322	13:05	49.48	-43.049	171.033	13.01	0.28	5	16	276	2.88
20100322	14:40	42.37	-43.486	170.677	-1.97	0.23	5	13	124	1.82
20100325	4:49	48.15	-43.112	170.827	6.77	0.18	6	16	265	2.67
20100325	21:25	10.27	-43.533	170.634	5.25	0.17	7	16	111	1.85
20100326	6:57	14.42	-43.533	170.684	4.86	0.10	5	12	119	1.70
20100327	12:09	18.64	-43.325	170.990	-1.93	0.08	3	8	264	1.54
20100327	12:32	26.52	-43.161	170.879	11.28	0.32	3	12	218	1.71
20100328	10:46	8.49	-43.080	170.812	9.48	0.25	7	18	251	2.14
20100329	10:36	52.90	-43.539	170.627	8.28	0.22	10	21	109	2.27
20100329	16:58	19.35	-43.530	170.629	3.58	0.16	6	16	111	1.68
20100329	18:50	54.35	-43.469	170.416	6.15	0.01	1	5	171	1.48
20100329	18:50	56.52	-43.467	170.422	5.80	0.13	5	12	174	1.48
20100329	18:51	9.39	-43.455	170.457	2.92	0.06	3	6	191	NA
20100329	19:23	26.91	-43.537	170.629	8.00	0.19	9	19	110	2.15
20100330	1:54	7.38	-43.466	170.416	6.86	0.18	9	20	87	2.18
20100331	8:23	18.38	-43.541	170.634	8.02	0.19	7	15	214	1.52
20100331	10:34	5.88	-43.109	171.037	6.94	0.18	2	9	298	2.27
20100401	2:57	57.86	-43.380	170.710	-1.97	0.18	4	10	139	1.49
20100401	11:18	27.02	-43.518	170.647	10.55	0.14	6	13	165	1.60
20100401	11:18	27.10	-43.513	170.650	8.16	0.14	5	12	166	1.60
20100401	11:49	40.87	-43.036	171.045	21.64	0.22	1	9	280	1.74
20100401	22:14	41.26	-43.183	170.901	-1.86	0.09	1	6	260	1.69
20100401	22:14	41.06	-43.182	170.913	-1.93	0.13	2	7	263	1.69
20100402	16:04	21.74	-43.133	170.827	12.04	0.30	3	13	214	2.55
20100402	16:36	11.05	-43.499	170.379	4.05	0.17	10	20	117	1.97
20100402	19:34	8.91	-43.572	171.131	14.71	0.03	3	8	231	1.74
20100402	21:02	16.93	-43.497	170.379	2.68	0.19	9	18	151	1.75
20100402	23:42	54.19	-43.402	170.797	-0.99	0.10	4	11	157	1.71
20100403	5:57	53.70	-43.499	170.379	3.44	0.14	9	18	151	1.74
20100403	7:11	9.54	-43.496	170.382	3.58	0.12	6	15	152	1.29
20100403	7:17	23.13	-43.553	170.669	10.97	0.03	1	5	273	1.77
20100403	7:22	49.67	-43.493	170.387	4.71	0.16	8	18	92	2.92
20100403	11:12	47.22	-43.494	170.385	4.53	0.17	9	19	92	1.87
20100403	12:48	26.13	-43.494	170.389	3.58	0.10	6	14	155	1.36

J. Hypocentre locations

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100404	19:13	13.92	-43.520	170.648	11.00	0.14	7	15	164	1.63
20100405	11:29	20.46	-43.498	170.383	4.14	0.15	10	20	118	1.92
20100405	12:04	38.31	-43.239	170.878	8.47	0.14	5	14	197	1.84
20100405	17:10	22.98	-43.493	170.289	1.15	0.15	9	19	77	1.98
20100405	22:15	25.59	-43.495	170.382	3.44	0.11	6	15	118	1.47
20100406	13:49	42.46	-43.177	170.830	8.05	0.16	1	9	197	1.82
20100407	17:55	12.56	-43.352	170.657	-1.97	0.23	6	13	129	1.31
20100408	10:06	32.34	-43.160	170.820	11.76	0.16	3	14	199	2.64
20100408	18:43	33.47	-43.099	171.096	5.90	0.18	1	6	303	1.35
20100408	20:40	20.87	-43.496	170.383	3.51	0.11	6	15	153	1.38
20100408	22:40	6.31	-43.493	170.385	3.44	0.14	4	10	154	1.05
20100409	16:59	49.29	-43.352	170.730	-1.97	0.18	5	11	208	1.00
20100409	19:36	55.90	-43.490	171.057	14.33	0.08	4	13	210	1.51
20100409	22:04	37.18	-43.187	170.856	-1.93	0.07	2	7	247	1.31
20100410	5:51	7.92	-43.143	170.921	9.58	0.11	2	7	275	1.62
20100410	8:31	3.95	-43.297	170.806	-1.97	0.17	4	8	225	1.11
20100410	9:17	26.60	-43.172	171.253	21.37	0.17	2	9	271	2.14
20100410	11:24	24.73	-43.494	170.382	4.55	0.18	9	20	92	1.83
20100410	16:40	20.33	-43.381	170.707	-1.97	0.22	7	17	139	2.35
20100410	23:16	10.05	-43.096	171.409	0.01	0.15	1	6	318	1.82
20100411	4:42	19.97	-43.256	170.896	-1.03	0.05	2	5	249	1.63
20100412	7:35	1.09	-43.129	171.016	-1.10	0.07	2	6	311	1.57
20100413	18:35	11.35	-43.223	170.560	6.98	0.17	5	15	152	2.00
20100415	22:47	36.15	-43.242	170.754	0.01	0.13	2	6	208	1.66
20100416	15:40	47.29	-43.100	171.074	-0.96	0.22	3	9	302	2.05
20100416	18:03	3.56	-43.129	170.866	16.24	0.00	1	4	228	1.19
20100417	12:29	56.53	-43.517	170.662	10.62	0.03	2	5	293	1.13
20100417	17:59	36.82	-43.471	170.382	0.46	0.11	2	10	157	1.52
20100418	13:50	44.22	-43.314	170.807	-1.97	0.14	5	13	168	1.42
20100418	14:19	52.57	-43.386	170.762	0.08	0.08	3	8	219	1.26
20100418	14:43	34.82	-43.315	170.804	-1.97	0.15	3	9	167	1.33
20100418	17:37	38.46	-43.319	170.820	-1.97	0.14	6	12	229	2.84
20100419	9:04	57.44	-43.101	170.942	14.85	0.17	2	12	253	2.04
20100419	10:42	20.99	-43.410	170.691	-1.97	0.21	6	14	133	1.38
20100421	4:02	56.48	-43.420	170.224	8.94	0.18	8	17	103	1.62
20100421	15:42	1.96	-43.661	170.681	9.48	0.17	8	18	173	1.65
20100422	16:54	51.42	-43.245	170.856	5.07	0.00	1	4	308	1.53
20100423	11:17	24.38	-43.245	171.012	21.16	0.01	3	6	317	1.69
20100424	8:49	9.92	-43.548	170.596	10.31	0.06	3	8	248	1.20
20100424	17:31	3.33	-43.501	170.698	7.88	0.13	6	16	126	2.09
20100430	16:16	21.00	-43.077	170.803	6.98	0.29	3	9	253	1.91
20100501	2:53	28.60	-43.464	170.618	6.74	0.12	6	13	251	1.75
20100828	15:18	54.49	-43.506	170.148	4.07	0.09	8	16	110	0.97
20100828	15:19	12.19	-43.508	170.148	4.53	0.12	9	18	112	2.57
20100828	15:20	13.19	-43.506	170.148	4.07	0.09	7	16	110	1.13
20100828	15:20	13.19	-43.507	170.151	4.00	0.09	7	15	111	1.13
20100828	15:32	21.05	-43.510	170.153	5.18	0.08	7	15	113	0.87

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100828	17:53	5.73	-43.508	170.150	4.90	0.10	7	16	112	0.97
20100828	20:22	21.29	-43.542	170.139	2.78	0.10	8	17	129	1.08
20100828	20:22	33.74	-43.541	170.136	4.97	0.07	2	8	145	0.42
20100828	23:48	36.45	-43.163	170.931	11.38	0.11	4	11	229	2.02
20100829	8:14	26.85	-43.540	170.139	4.00	0.08	8	17	128	1.06
20100829	8:23	30.78	-43.466	170.185	6.44	0.08	8	17	75	1.46
20100829	17:30	22.90	-43.442	170.679	12.46	0.12	4	11	210	1.10
20100829	21:37	8.95	-43.577	170.521	8.57	0.02	3	7	230	0.69
20100829	22:59	19.94	-43.579	170.523	8.57	0.09	9	18	148	1.62
20100829	23:07	21.88	-43.575	170.521	8.12	0.01	3	8	230	0.89
20100830	7:47	15.46	-43.509	170.153	5.59	0.08	5	13	112	0.52
20100830	17:58	51.66	-43.523	170.283	4.64	0.13	8	17	103	1.13
20100830	21:30	43.22	-43.507	170.147	2.75	0.11	9	18	111	1.43
20100831	5:24	50.88	-43.509	170.154	4.62	0.08	5	10	148	0.53
20100831	6:33	48.72	-43.507	170.152	4.00	0.10	8	17	111	1.11
20100831	6:48	16.64	-43.506	170.154	3.30	0.10	7	16	110	1.00
20100831	7:28	10.33	-43.509	170.155	3.48	0.08	8	17	112	1.03
20100831	8:29	27.14	-43.510	170.153	3.79	0.08	6	12	113	0.78
20100831	8:36	51.40	-43.509	170.152	3.44	0.11	9	18	113	1.64
20100831	8:49	20.62	-43.582	170.526	9.13	0.10	10	20	147	1.83
20100831	9:29	33.04	-43.555	170.559	7.67	0.11	8	17	256	1.48
20100831	13:26	48.81	-43.574	170.523	7.53	0.06	7	14	232	1.09
20100831	13:27	16.54	-43.574	170.524	7.46	0.08	6	13	233	0.80
20100831	13:34	4.34	-43.702	170.123	5.70	0.08	7	15	229	1.24
20100831	13:38	7.25	-43.577	170.526	8.57	0.07	7	16	234	1.37
20100831	13:41	18.28	-43.568	170.539	7.60	0.06	5	11	244	0.73
20100831	13:42	57.03	-43.533	170.552	6.25	1.01	5	12	251	0.78
20100831	13:56	29.19	-43.452	170.364	5.45	0.07	6	15	123	1.52
20100831	14:08	4.51	-43.576	170.521	8.61	0.03	3	7	231	0.44
20100831	14:08	32.38	-43.500	170.265	3.75	0.06	3	6	144	0.20
20100831	20:22	9.81	-43.501	170.278	-1.97	0.10	6	13	81	0.64
20100831	20:22	22.62	-43.502	170.280	-1.97	0.10	5	12	82	0.87
20100831	22:09	42.57	-43.578	170.524	9.23	0.01	2	5	233	0.31
20100831	22:12	46.20	-43.519	170.268	3.44	0.09	5	11	86	0.33
20100901	8:52	33.51	-43.706	170.122	5.70	0.06	7	16	231	1.24
20100901	23:39	59.91	-43.568	170.548	8.68	0.01	1	4	250	0.45
20100902	4:01	56.83	-43.518	169.991	5.40	0.11	7	15	132	0.96
20100903	7:44	52.15	-43.572	170.528	7.67	0.05	5	11	236	1.20
20100903	8:20	1.67	-43.406	170.756	-1.97	0.13	8	17	147	1.65
20100903	14:30	27.99	-43.227	170.835	15.54	0.19	4	12	187	1.83
20100903	17:13	17.84	-43.478	170.195	3.27	0.01	2	5	271	1.47
20100903	17:13	22.22	-43.475	170.201	4.00	0.06	3	8	155	1.47
20100903	17:52	23.18	-43.346	170.347	10.55	0.04	2	5	263	NA
20100903	19:09	29.43	-43.706	169.847	8.40	0.01	1	4	309	NA
20100903	19:12	44.66	-43.686	169.850	9.51	0.04	2	5	300	NA
20100903	19:18	42.31	-43.478	170.066	0.98	0.00	1	4	172	NA
20100903	19:22	22.30	-43.491	170.192	8.78	0.13	5	11	163	1.95

J. Hypocentre locations

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20100903	20:22	3.52	-43.679	169.861	9.06	0.11	7	15	262	2.55
20100903	20:25	16.28	-43.552	170.584	8.05	0.06	2	6	267	1.82
20100903	20:27	40.60	-43.449	170.363	7.78	0.00	2	5	256	NA
20100903	20:48	38.69	-43.444	170.369	7.85	0.00	1	4	259	NA
20100903	20:51	35.55	-43.444	170.369	7.85	0.01	1	4	259	NA
20100903	21:01	10.33	-43.479	170.285	5.56	0.00	1	4	182	NA
20100903	21:01	10.22	-43.432	170.391	9.72	0.04	1	5	215	NA
20100903	21:01	37.13	-43.515	170.283	-1.10	0.03	1	4	285	NA
20100903	21:20	4.87	-43.518	170.353	4.73	0.11	5	13	128	1.32
20100903	21:20	42.82	-43.499	170.370	3.48	0.07	4	8	162	1.01
20100903	21:26	9.33	-43.506	170.369	7.33	0.02	3	7	141	NA
20100903	21:43	57.57	-43.578	170.524	8.68	0.01	1	4	233	NA
20100903	21:50	10.81	-43.578	170.518	8.19	0.05	3	7	228	1.33
20100903	22:00	23.02	-43.500	170.363	7.01	0.00	1	4	302	1.22
20100903	22:01	53.20	-43.524	170.327	-1.93	0.08	1	5	119	0.53
20100903	22:11	1.99	-43.427	170.393	9.23	0.01	1	4	286	NA
20100903	22:22	39.38	-43.538	170.462	-0.96	0.04	2	5	198	NA
20100903	22:34	48.73	-43.464	170.206	7.22	0.07	3	8	149	1.02
20100903	22:48	59.38	-43.339	170.354	10.55	0.10	4	9	270	1.64
20100903	22:49	59.66	-43.495	170.319	-1.93	0.07	1	5	177	NA
20100903	22:50	10.66	-43.509	170.367	1.61	0.08	4	11	140	1.14
20100903	23:50	52.79	-43.495	170.393	5.42	0.04	3	8	166	0.79
20100904	0:43	52.25	-43.482	170.378	3.96	0.01	1	4	309	NA
20100904	0:56	39.79	-43.455	170.271	4.59	0.04	4	10	115	1.39
20100904	0:56	49.97	-43.466	170.266	5.00	0.01	2	5	246	0.78
20100904	2:59	9.86	-43.481	170.367	3.30	0.10	6	12	143	1.54
20100904	3:01	21.81	-43.463	170.258	4.14	0.08	5	10	99	1.49
20100904	4:08	31.13	-43.619	169.793	7.12	0.08	6	13	278	2.22
20100904	4:14	55.46	-43.500	170.364	2.33	0.08	6	13	138	1.29
20100904	4:24	6.91	-43.338	170.901	9.41	0.09	4	12	188	2.64
20100904	4:24	26.77	-43.479	170.206	6.70	0.07	5	10	109	2.14
20100904	4:40	9.59	-43.498	170.329	3.37	0.10	4	11	109	1.27
20100904	4:53	11.79	-43.503	170.329	-1.97	0.13	4	10	110	0.98
20100904	5:11	29.89	-43.500	170.385	4.86	0.02	2	5	157	1.37
20100904	5:15	12.01	-43.492	170.377	5.14	0.00	1	4	156	NA
20100904	5:55	54.09	-43.492	170.352	4.34	0.11	4	10	127	0.89
20100904	6:02	12.83	-43.470	170.387	6.18	0.01	1	4	320	NA
20100904	6:42	16.62	-43.487	170.376	4.45	0.07	1	5	307	NA
20100904	6:59	29.24	-43.460	170.551	5.14	0.03	2	7	228	1.03
20100904	7:23	40.40	-43.535	170.637	11.17	0.03	1	5	281	NA
20100904	7:32	51.20	-43.492	170.367	4.62	0.08	2	7	160	1.00
20100904	8:00	17.75	-43.494	170.368	3.89	0.08	3	9	143	0.96
20100904	8:07	7.14	-43.491	170.369	4.55	0.10	3	9	144	1.01
20100904	8:37	32.12	-43.494	170.345	0.29	0.07	2	6	290	NA
20100904	10:53	56.84	-43.453	170.410	5.66	0.09	4	9	282	0.50
20100904	10:59	54.19	-43.456	170.252	7.60	0.08	2	7	152	NA
20100904	11:00	29.43	-43.475	170.239	4.31	0.04	2	5	258	0.65

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100904	11:09	40.67	-43.469	170.594	-0.96	0.01	1	5	243	NA
20100904	11:34	50.86	-43.460	170.247	6.35	0.15	7	15	98	2.29
20100904	11:38	44.74	-43.442	170.262	7.60	0.11	4	8	219	0.96
20100904	11:47	31.49	-43.437	170.388	6.01	0.14	6	13	164	1.03
20100904	12:18	40.70	-43.492	170.380	4.41	0.07	2	7	156	0.82
20100904	12:38	40.11	-43.492	170.382	4.45	0.07	1	6	155	0.76
20100904	13:01	2.88	-43.397	170.368	-0.96	0.04	1	4	241	-0.25
20100904	13:13	12.31	-43.427	170.391	5.00	0.01	1	4	233	0.03
20100904	13:21	37.93	-43.490	170.375	4.79	0.00	1	4	307	0.05
20100904	13:23	7.68	-43.450	170.353	4.45	0.01	1	4	239	0.19
20100904	13:55	5.41	-43.554	170.591	5.42	0.11	2	6	248	1.23
20100904	14:13	4.66	-43.490	170.389	5.11	0.08	4	9	156	0.80
20100904	14:18	51.06	-43.308	170.786	-1.86	0.10	3	9	220	1.47
20100904	14:18	50.96	-43.310	170.796	-1.93	0.12	4	10	223	1.47
20100904	14:39	51.93	-43.451	170.360	4.59	0.06	1	5	256	0.38
20100904	14:59	10.67	-43.422	170.359	8.12	0.00	1	4	210	NA
20100904	15:00	11.46	-43.343	170.387	4.45	0.05	3	7	203	0.86
20100904	15:07	32.76	-43.490	170.203	5.70	0.07	2	5	276	0.64
20100904	18:23	47.03	-43.435	170.394	5.94	0.12	7	15	168	1.47
20100904	18:29	24.70	-43.427	170.400	5.14	0.05	1	5	247	0.22
20100904	18:41	12.16	-43.487	170.359	5.45	0.10	3	8	160	NA
20100904	18:42	34.36	-43.489	170.357	5.28	0.09	3	9	135	0.84
20100904	18:50	2.08	-43.487	170.347	0.36	0.09	0	4	287	NA
20100904	18:53	44.14	-43.434	170.393	6.01	0.11	5	13	167	1.30
20100904	19:33	48.60	-43.492	170.357	4.59	0.06	3	6	296	NA
20100904	19:54	3.23	-43.462	170.381	7.29	0.00	1	4	315	NA
20100904	20:04	1.30	-43.495	170.364	5.42	0.00	1	4	301	NA
20100904	20:11	25.38	-43.502	170.344	3.10	0.14	8	17	121	1.24
20100904	20:11	38.45	-43.450	170.354	4.59	0.00	1	4	241	0.48
20100904	20:37	38.79	-43.494	170.356	4.73	0.05	3	7	253	0.49
20100904	21:27	45.64	-43.649	169.914	9.48	0.11	4	10	280	1.26
20100904	23:32	11.02	-43.484	170.370	3.93	0.10	3	7	262	0.86
20100904	23:51	26.20	-43.506	170.372	4.07	0.20	8	16	145	1.54
20100905	0:12	21.04	-43.489	170.385	5.28	0.05	3	7	308	0.79
20100905	1:17	11.89	-43.558	169.660	10.62	0.06	5	10	319	1.71
20100905	1:18	57.08	-43.547	169.669	12.32	0.07	6	13	317	1.97
20100905	2:07	25.13	-43.427	170.391	5.94	0.05	3	6	233	0.82
20100905	2:07	43.40	-43.437	170.389	6.16	0.14	8	17	165	1.74
20100905	2:07	43.26	-43.433	170.399	6.77	0.16	6	14	125	1.74
20100905	11:11	30.84	-43.491	170.389	5.25	0.06	4	9	156	1.54
20100906	16:01	11.81	-43.599	169.981	5.73	0.09	4	10	117	0.98
20100906	16:01	27.07	-43.607	169.994	6.49	0.05	2	6	216	0.62
20100906	20:17	40.92	-43.493	170.339	3.34	0.08	4	9	157	0.52
20100906	20:51	25.39	-43.492	170.374	4.97	0.09	5	11	149	0.77
20100906	21:15	40.87	-43.491	170.372	4.34	0.08	4	9	147	0.72
20100906	21:31	29.51	-43.502	170.360	1.81	0.00	1	4	155	NA
20100907	1:45	39.86	-43.512	170.274	3.75	0.02	2	6	142	0.38

Date yyyymmdd	Time hrmin	Sec [s]	Latitude [°]	Longitude [°]	Depth [km]	RMS [s]	STN	NPH	GAP [°]	Mag
20100907	9:12	31.17	-43.560	170.563	7.50	0.10	2	8	236	1.15
20100907	12:58	11.46	-43.562	170.545	7.05	0.15	5	12	227	1.24
20100907	13:16	19.07	-43.565	170.532	7.33	0.15	4	9	219	1.15
20100907	20:24	8.48	-43.577	170.515	8.61	0.02	2	5	226	0.80
20100907	21:31	7.58	-43.513	170.259	6.94	0.07	3	7	119	0.49
20100908	0:19	41.93	-43.508	170.362	3.06	0.01	2	6	157	NA
20100908	3:54	58.58	-43.298	170.803	10.13	0.14	3	9	224	1.52
20100908	5:12	50.86	-43.447	170.358	0.74	0.06	2	6	218	-0.18
20100908	7:37	30.28	-43.501	170.327	5.21	0.05	3	8	165	0.28
20100908	9:25	30.36	-43.500	170.107	-0.75	0.01	1	4	346	NA
20100908	13:07	31.30	-43.448	170.535	8.85	0.11	5	11	224	1.07
20100908	14:09	42.04	-43.565	170.543	8.47	0.14	4	9	226	0.81
20100908	19:01	25.70	-43.559	169.932	6.49	0.12	4	10	251	NA
20100908	19:06	35.56	-43.564	169.918	7.74	0.10	4	9	260	2.33
20100908	23:32	27.24	-43.542	170.322	-0.02	0.15	3	7	189	0.19

Table J.1. Hypocentre locations for all events recorded between January and April 2010.

K. Focal mechanism solutions

Focal mechanism solutions for the period November 2008 to April 2010. Cluster- and event-numbers refer to those events shown in Fig. 3.22, H.1–H.3

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20081113	12:53	28.94	158	79	-41	
20081130	11:05	32.05	237	69	151	
20081208	17:34	22.48	341	77	45	2/1
20081213	11:23	2.96	338	87	7	2/2
20081214	5:30	43.85	68	60	135	
20081214	9:56	5.13	57	49	119	
20081217	5:39	6.45	160	79	-2	2/3
20081219	14:53	39.20	339	83	24	2/4
20081220	15:22	41.33	256	69	167	1/1
20081221	4:10	11.60	310	73	-146	5/1
20081223	1:36	50.07	201	38	76	
20081225	7:18	40.81	267	49	161	
20081225	7:42	55.09	0	78	29	
20081227	22:35	41.09	254	82	92	1/2
20090110	8:44	0.43	202	69	30	
20090111	14:28	52.48	252	78	-162	
20090127	12:33	17.44	271	69	-158	
20090205	8:19	36.70	248	55	157	
20090211	5:28	7.08	224	78	0	
20090224	19:43	21.29	201	38	57	
20090225	8:32	31.99	259	81	-179	2/5
20090225	21:42	18.05	335	87	16	2/6
20090315	9:20	6.27	202	70	-13	
20090316	17:32	25.74	12	72	11	2/7
20090413	1:05	12.02	74	81	169	
20090419	15:20	57.83	159	68	7	
20090503	13:28	31.16	89	86	176	3/1
20090507	7:27	46.39	288	78	-153	5/2
20090507	17:20	5.30	287	80	-153	5/3
20090508	0:34	38.36	151	48	-44	
20090513	4:27	42.30	233	57	165	2/8
20090524	5:17	59.58	15	56	7	2/9
20090526	9:57	9.30	236	74	169	3/2
20090601	2:58	0.14	153	81	6	
20090618	5:11	18.57	197	43	107	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20090618	7:25	9.29	106	63	-61	2/10
20090622	1:10	48.47	63	16	-60	
20090622	4:06	6.23	286	48	-96	
20090622	14:07	1.80	181	67	93	
20090623	1:52	0.04	56	77	161	
20090623	1:53	22.24	57	77	160	
20090623	2:59	21.22	158	70	13	
20090627	0:13	5.76	259	83	-167	
20090628	18:23	21.95	35	86	160	1/3
20090630	17:43	37.06	240	71	178	2/11
20090701	8:47	20.91	55	65	128	
20090701	19:36	29.80	162	79	12	
20090703	18:36	16.53	13	66	15	2/12
20090705	1:53	54.10	158	89	-25	2/13
20090705	18:34	18.13	306	81	7	
20090707	7:57	44.36	14	57	47	2/14
20090707	16:29	25.00	180	64	22	2/15
20090708	6:18	8.91	303	70	12	
20090708	9:24	6.44	75	74	153	2/16
20090708	14:15	7.31	284	40	-152	
20090709	4:37	10.26	112	41	-65	
20090709	13:02	47.97	248	38	124	2/17
20090709	13:26	7.77	258	37	121	2/18
20090711	23:32	18.62	51	68	122	
20090712	6:03	45.62	74	81	165	
20090714	1:35	15.19	46	66	113	
20090714	6:00	46.95	55	72	162	
20090714	9:58	52.30	321	73	-30	
20090715	7:35	16.46	33	84	86	
20090715	10:59	19.90	330	82	0	2/19
20090715	17:29	22.09	64	67	-166	2/20
20090715	17:46	47.11	160	84	-2	2/21
20090715	19:25	11.34	331	80	-2	2/22
20090715	20:28	55.51	158	69	-34	3/3
20090716	6:54	12.34	73	66	-174	2/23
20090716	9:25	1.31	38	67	127	5/4
20090716	14:32	4.42	173	72	1	3/4
20090716	14:49	55.19	325	74	-21	2/24
20090717	10:21	53.03	76	81	164	3/5
20090717	15:31	15.39	334	72	5	1/4
20090718	8:43	57.80	338	32	43	2/25
20090718	9:45	1.75	63	64	-168	2/26
20090718	16:42	33.42	96	50	-135	1/5
20090719	18:22	59.04	231	82	-168	
20090720	11:56	26.79	167	76	-17	5/5
20090720	13:13	7.86	355	87	29	5/6
20090724	15:17	16.17	94	79	-137	2/27

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20090724	15:24	1.40	194	64	46	2/28
20090726	14:55	12.35	248	75	-164	2/29
20090727	8:07	18.53	320	54	-21	2/30
20090729	11:13	47.29	339	83	15	
20090729	11:41	43.41	240	68	179	2/31
20090731	0:29	45.31	242	63	-173	2/32
20090801	2:27	55.44	84	83	-151	1/6
20090801	5:17	17.01	191	64	16	
20090801	19:57	13.61	240	84	-179	
20090803	8:43	21.71	285	80	-155	5/7
20090806	16:45	45.26	255	64	-161	3/6
20090807	6:44	41.18	161	89	3	
20090808	15:41	42.68	251	86	141	1/7
20090808	17:36	57.43	307	18	-26	
20090811	16:28	42.10	17	55	38	2/33
20090813	7:07	12.56	148	46	20	5/8
20090814	16:24	5.75	152	38	-28	5/9
20090815	12:42	50.42	18	80	22	1/8
20090815	18:30	30.55	149	45	21	5/10
20090816	21:56	56.40	102	31	90	1/9
20090818	1:51	2.42	357	71	33	1/10
20090818	10:34	43.91	285	87	18	5/11
20090819	10:50	28.00	217	53	-13	
20090820	13:49	9.06	51	32	81	
20090820	18:39	2.49	234	65	171	
20090820	21:25	9.91	256	58	148	2/34
20090821	11:33	36.06	223	60	-177	
20090822	0:38	20.77	319	79	13	1/11
20090826	11:51	12.34	93	60	-158	
20090828	1:47	33.92	52	64	-160	
20090828	6:55	10.57	76	89	179	2/35
20090904	20:24	7.70	144	67	-26	2/36
20090905	11:48	32.51	244	60	-146	2/37
20090905	18:47	10.91	178	83	17	
20090906	11:03	32.86	152	66	-40	5/12
20090906	12:30	33.46	47	80	179	
20090907	22:17	17.33	155	56	-28	5/13
20090908	19:20	47.68	292	66	-141	5/14
20090909	14:21	10.59	220	46	61	
20090912	5:38	12.46	144	82	1	2/38
20090912	11:19	36.18	144	85	1	2/39
20090913	10:54	18.80	10	53	102	
20090913	11:45	20.79	162	73	139	
20090914	19:32	52.90	44	74	-176	2/40
20090915	9:54	54.23	191	83	-4	
20090915	14:25	0.71	59	85	-158	2/41
20090916	7:59	6.12	48	68	124	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20090917	10:07	12.42	72	82	176	
20090918	21:00	48.49	141	38	28	3/7
20090919	4:07	43.81	45	62	148	
20090920	6:02	42.18	337	56	-43	1/12
20090921	13:30	22.32	60	78	-171	2/42
20090921	23:56	7.74	234	59	-166	2/43
20090922	6:48	1.98	12	85	45	
20090922	18:39	39.56	270	71	-153	
20090923	6:38	11.76	92	84	-171	
20090925	8:35	21.90	219	68	129	2/44
20090925	20:12	29.27	48	77	132	5/15
20090926	12:47	54.88	220	71	126	2/45
20090926	13:02	41.11	219	68	131	2/46
20090926	18:27	35.89	1	80	21	
20090926	22:23	5.26	90	74	162	
20090927	5:17	27.13	250	78	148	1/13
20090929	6:41	34.96	21	87	34	
20090929	16:07	51.96	327	58	17	1/14
20090930	2:25	13.96	148	32	39	5/16
20091001	14:37	43.31	320	74	23	2/47
20091002	2:52	43.43	158	66	2	5/17
20091002	10:57	2.21	55	78	177	2/48
20091002	10:57	23.85	221	70	126	2/49
20091003	3:34	35.10	202	85	-1	2/50
20091005	1:53	46.93	319	66	-33	2/51
20091005	19:47	44.45	128	20	-126	
20091009	21:29	6.05	219	70	129	2/52
20091010	18:06	38.43	328	77	-15	5/18
20091010	19:02	15.90	282	52	-63	
20091011	13:54	18.87	34	71	125	3/8
20091013	17:06	11.05	183	42	124	1/15
20091015	12:22	22.31	168	88	-2	2/53
20091016	8:13	42.70	307	77	-19	2/54
20091017	14:47	14.29	45	72	-173	2/55
20091018	9:10	43.22	307	76	-19	2/56
20091019	6:01	54.10	38	72	124	5/19
20091020	8:19	53.33	323	86	0	1/16
20091023	18:54	32.80	180	86	-1	2/57
20091024	12:53	2.79	73	78	-146	2/58
20091024	20:22	27.16	146	72	25	2/59
20091027	0:01	13.73	337	72	14	2/60
20091028	20:13	38.46	242	62	-148	2/61
20091028	20:32	22.17	45	87	176	2/62
20091101	7:07	46.83	317	56	9	
20091109	4:37	46.54	56	66	127	
20091109	13:59	4.96	60	76	164	
20091109	14:02	27.71	223	39	154	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20091109	14:09	59.45	156	68	17	
20091109	14:21	42.20	157	68	17	
20091110	8:13	15.44	106	84	-128	2/63
20091112	1:17	3.31	287	82	21	5/20
20091117	15:40	19.76	173	57	34	
20091117	23:07	37.85	235	27	-22	
20091118	7:22	4.34	54	50	94	2/64
20091118	14:13	23.78	286	82	21	5/21
20091118	19:37	12.29	69	81	90	
20091120	13:04	3.47	23	89	-14	5/22
20091124	2:00	18.06	53	84	169	
20091129	23:36	46.69	323	82	16	
20091202	4:37	19.89	154	81	-64	1/17
20091205	11:30	52.85	156	85	12	
20091205	11:40	6.97	69	77	-178	
20091205	17:00	5.56	100	84	171	5/23
20091205	19:32	49.22	81	89	174	2/65
20091209	19:39	39.16	1	83	-22	
20091215	13:01	18.46	189	79	8	
20091216	17:27	4.05	59	72	157	
20091217	3:46	58.24	329	75	-11	2/66
20091217	4:00	15.62	329	74	-10	2/67
20091217	10:27	20.12	327	69	-13	2/68
20091217	16:15	32.53	178	51	-23	3/9
20091221	16:26	6.85	126	72	6	
20091221	21:02	43.15	162	67	38	2/69
20091222	2:45	34.18	162	64	40	2/70
20091222	10:48	38.57	162	64	41	2/71
20091223	23:18	33.18	194	82	5	
20091225	4:22	39.67	64	85	166	
20091225	13:56	21.95	26	68	-99	
20091228	12:47	47.52	297	68	-132	5/24
20091229	10:11	53.37	60	76	162	
20091230	15:10	14.88	283	85	-167	5/25
20091231	15:55	7.26	145	48	14	2/72
20100101	17:06	52.01	70	82	150	
20100102	16:34	26.48	295	71	-131	5/26
20100104	10:02	26.33	294	72	-131	5/27
20100107	13:04	44.24	53	59	-174	
20100108	9:27	19.38	44	83	87	2/73
20100110	20:54	1.53	84	55	158	2/74
20100112	11:34	47.38	70	76	165	
20100115	21:34	47.72	338	67	13	
20100116	11:02	31.26	93	76	-152	2/75
20100119	3:53	1.02	328	33	-64	
20100119	8:20	47.05	117	49	-126	
20100121	23:44	7.36	230	76	143	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20100122	9:12	20.95	339	75	16	2/76
20100123	9:52	25.60	203	60	-147	5/28
20100124	6:31	46.37	103	80	-168	5/29
20100124	8:46	29.88	161	74	-17	5/30
20100126	9:53	33.24	122	84	-141	1/18
20100127	3:16	2.34	159	69	18	
20100127	11:57	36.89	83	29	156	5/31
20100128	9:14	28.53	250	65	-142	
20100128	16:07	42.46	58	60	148	
20100128	16:35	22.57	157	73	12	5/32
20100129	17:47	55.27	1	67	20	2/77
20100130	2:42	14.05	43	51	23	2/78
20100131	11:47	36.61	255	85	-149	1/19
20100131	13:04	17.68	175	59	-30	
20100201	19:19	8.19	4	59	-35	
20100201	19:19	8.14	126	49	-90	
20100202	1:28	22.88	302	51	-128	5/33
20100202	4:26	44.85	88	44	165	
20100202	9:01	17.38	101	62	-170	
20100202	10:09	5.20	89	87	-142	5/34
20100202	20:55	18.70	182	85	76	
20100202	21:11	33.23	13	60	156	
20100203	11:14	39.36	55	62	145	
20100203	14:35	35.56	290	74	15	2/79
20100203	16:33	19.10	147	67	33	2/80
20100204	16:30	14.68	102	74	-166	
20100205	7:01	39.11	51	27	178	
20100205	23:07	30.08	54	79	169	5/35
20100206	8:44	55.20	187	89	0	5/36
20100206	8:54	0.38	103	82	-172	5/37
20100206	8:54	55.66	280	81	-168	5/38
20100206	9:48	37.70	101	82	169	5/39
20100206	11:55	3.56	7	82	-7	5/40
20100206	18:56	52.56	285	84	-165	5/41
20100206	19:04	18.04	285	80	-170	5/42
20100207	22:26	1.53	183	57	13	5/43
20100208	19:41	20.09	316	38	-21	
20100209	4:21	17.63	286	58	-137	1/20
20100209	8:26	11.89	13	67	13	
20100209	8:50	46.77	86	62	159	2/81
20100209	14:57	1.48	171	87	7	5/44
20100209	19:39	27.97	139	35	-17	5/45
20100209	19:57	9.11	286	85	-159	5/46
20100211	3:39	58.53	144	66	-19	5/47
20100211	20:05	16.78	244	68	147	
20100212	13:32	59.74	181	71	20	
20100213	2:59	14.24	123	64	-30	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20100213	11:36	45.31	351	64	-27	
20100213	19:32	13.17	77	82	-179	
20100214	7:27	43.41	162	74	23	2/82
20100215	13:30	29.97	216	53	139	
20100215	15:56	56.70	213	71	123	2/83
20100216	21:02	45.66	174	86	21	
20100217	1:09	12.65	145	74	-14	
20100217	5:57	25.99	145	84	-42	5/48
20100217	7:21	29.56	146	73	-10	5/49
20100217	7:28	57.28	146	72	-8	5/50
20100217	14:32	53.06	309	78	-10	5/51
20100218	11:02	48.47	268	79	-165	
20100219	10:03	36.96	251	66	-144	
20100219	17:58	20.81	270	65	-147	
20100220	2:31	30.96	246	84	-162	5/52
20100221	2:27	26.17	285	75	-150	
20100221	15:59	12.23	190	77	62	
20100221	18:08	32.78	300	72	13	
20100222	5:59	9.41	234	77	106	2/84
20100222	10:47	56.58	287	58	-35	5/53
20100223	22:22	0.00	92	62	145	
20100223	22:25	8.40	156	79	15	
20100225	3:42	21.63	153	71	-22	
20100225	11:53	18.50	325	71	-38	
20100225	14:04	34.43	250	58	-143	
20100227	6:41	13.03	181	42	-14	
20100227	8:06	15.63	70	52	145	5/54
20100228	3:35	39.35	94	49	-103	5/55
20100228	10:18	35.28	74	54	146	
20100228	16:35	11.82	252	70	-161	5/56
20100228	16:40	32.98	181	59	-36	5/57
20100228	18:20	59.02	179	75	-16	
20100228	23:42	59.58	237	62	-146	
20100301	2:56	22.63	298	62	-148	5/58
20100301	13:41	43.91	212	71	-142	5/59
20100301	13:43	23.59	137	62	-36	5/60
20100301	18:12	31.85	106	61	-35	5/61
20100302	7:39	32.52	136	50	-11	5/62
20100302	13:32	37.22	72	76	179	
20100302	22:23	53.45	222	74	70	2/85
20100303	23:24	55.09	10	76	0	2/86
20100304	4:45	43.47	286	65	-160	
20100304	6:50	38.98	289	69	-142	
20100305	22:28	37.03	88	64	159	2/87
20100305	22:49	3.77	263	65	148	2/88
20100307	1:15	6.18	347	81	-12	
20100307	20:38	0.06	342	59	-38	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20100308	19:43	35.22	68	55	-141	5/63
20100308	21:23	45.29	324	76	-15	
20100312	7:09	7.46	260	15	-157	
20100312	10:18	21.02	58	62	-116	5/64
20100312	22:03	29.22	265	83	-157	5/65
20100312	23:48	30.63	51	61	123	
20100317	15:10	51.58	290	88	6	5/66
20100318	10:33	31.12	145	68	-26	5/67
20100319	9:55	31.56	40	64	25	
20100319	10:20	27.01	161	63	-31	
20100319	10:50	6.30	288	87	148	
20100319	12:48	26.82	290	83	155	
20100319	15:05	37.12	309	82	39	
20100319	21:13	26.15	325	74	-14	
20100319	21:36	49.10	158	67	23	
20100320	17:28	39.39	264	55	-124	5/68
20100322	13:05	49.47	341	81	15	
20100322	14:40	42.37	338	60	-32	
20100325	4:49	48.15	79	35	156	5/69
20100325	21:25	10.28	166	57	27	
20100326	6:57	14.42	268	62	-148	
20100327	12:32	26.52	285	66	161	5/70
20100328	10:46	8.50	185	63	37	5/71
20100329	10:36	52.90	158	59	14	
20100329	19:23	26.91	51	61	145	
20100330	1:54	7.38	197	65	150	2/89
20100401	11:18	27.10	246	83	179	
20100401	11:18	27.02	245	89	-158	
20100402	16:04	21.74	187	43	19	5/72
20100402	16:36	11.06	358	54	16	2/90
20100402	19:34	8.89	265	65	-139	
20100402	21:02	16.93	355	59	15	2/91
20100403	5:57	53.70	207	81	88	2/92
20100403	7:22	49.67	226	76	103	2/93
20100403	11:12	47.22	348	14	56	2/94
20100403	12:48	26.13	209	86	95	2/95
20100404	19:13	13.92	250	77	-146	
20100405	11:29	20.46	256	82	156	2/96
20100405	17:10	22.98	4	71	30	2/97
20100407	17:55	12.56	151	77	15	
20100408	10:06	32.32	228	71	-156	5/73
20100408	20:40	20.87	210	81	89	2/98
20100410	9:17	26.64	214	67	-143	
20100410	11:24	24.73	358	54	17	2/99
20100410	16:40	20.33	175	81	10	
20100411	4:42	19.97	54	49	-90	5/74
20100413	18:35	11.35	112	58	-38	

Date yyyymmdd	Time hrmin	Sec [s]	Strike [°]	Dip [°]	Rake [°]	Cluster/Event
20100417	17:59	36.82	258	88	-159	2/100
20100418	13:50	44.23	309	84	18	5/75
20100418	14:19	52.57	197	68	-21	
20100418	14:43	34.82	321	70	-20	5/76
20100418	17:37	38.46	228	75	-164	5/77
20100419	10:42	20.99	244	78	-172	
20100421	4:02	56.48	96	69	-141	
20100422	16:54	51.44	192	89	13	5/78
20100424	17:31	3.34	255	59	-152	
20100501	2:53	28.60	159	57	32	

Table K.1. Focal mechanism solutions for all events recorded between November 2008 and April 2010.

L. Waveform examples of earthquake sequences

Figure L.1. Earthquake swarm on 12 March 2009 as recorded on the vertical component by station WHYM.

Figure L.2. Earthquake swarm on 15 July 2009 as recorded on the vertical component by station WHYM. Note that the event on the 19 July are slightly shallower (5 km).

Figure L.3. Earthquake swarm on 16 July 2009 as recorded on the vertical component by station WHYM. Note that the later events in October are deeper (7.5 km) than the events at the beginning (5.5 km versus 6.5 km).

Figure L.4. Earthquake swarm on 12 September 2009 as recorded on the vertical component by station WHYM.

Figure L.5. Earthquake swarm on 22 September 2009 as recorded on the vertical component by station WHYM.

Figure L.6. Earthquake swarm on 17 December 2009 as recorded on the vertical component by station WHYM.

Figure L.7. Earthquake swarm on 30 January 2010 as recorded on the vertical component by station WHYM.

Figure L.8. Earthquake swarm on 28 August 2010 as recorded on the vertical component by station WHYM.

Figure L.9. Earthquake swarm on 03 September 2010 as recorded on the vertical component by station WHYM.

M. Overview of combined global studies of seismicity and resistivity

Seismicity in resistive and conductive zone boundaries (RCZB), generally in resistive side			
Region/fault zone	Earthquake	Reference	Explanations
NE Japan back arc	Intraplate earthquake zone	Ogawa et al. (2001)	Seismicity at RCZB in MT model
NE Japan	Miyagi earthquake (M6.5)	Mitsuhashi et al. (2001)	Microearthquakes just above the highly conductive deep zone in the overlying moderately resistive region
Japan/Itoigawa–Shizuoka tect. line		Ogawa et al. (2002)	Hypocenters in resistive region near the RCZB
Central Japan/Mt. Ontake	In Western Nagano Prefecture earthquake M6.8 area	Kasaya and Oshiman (2004)	Hypocenters in resistive region around the RCZB
Central Japan	Mid-Niigata prefecture earthquake (M6.8) area	Uyeshima et al. (2005)	The mainshock at the sediment cover/resistive basement border, aftershocks in sedimentary layer
Central Japan/Atotsugawa Fault	High seismicity in western, low seismicity (a gap) in central segment	Goto et al. (2005)	High seismicity in RCZB at resistive side, less conductive lower crust in active western segment
NE Japan Arc	High seismicity in central mountain range, no large earthquakes at Central Basin	Fujinawa et al. (2002)	High resistivity in high a seismicity region, conductive zone in seismically quiet region
USA/San Andreas, creeping Parkfield segment	Repeating microearthquakes and characteristic M6 events every 22 yr	Unsworth et al. (2000)	Hypocenters of microearthquakes are beneath the fault zone conductors and in resistive side
USA/San Andreas, creeping central segment at Hollister	Microseismicity on MT models	Bedrosian et al. (2002)	High seismicity at the border of fault zone conductor, inside the conductive zone
NW Taiwan/Puli Fault	Chi-Chi Mw 7.6 earthquake and Sanyi Puli seismic zone	Chen and Chen (2002)	Main seismicity in resistive layers at RCZB around Sanyi Puli conductivity anomaly, less activity in CA
India/Latur	Latur earthquake Mw 6.1, a stable continental region	Gupta et al. (1996)	Main and aftershocks in the resistive zone just above an upper crustal conductor.

NW Turkey/ North Anatolian Fault Zone	M7.4 1999 Izmit Earthquake Seismically active region, many microearthquakes and aftershocks	Tank et al. (2005)	The hypocenters of mainshock and aftershocks are located on the highly resistive side near RCZB
Seismicity in resistive zones			
Region/fault zone	Earthquake	Reference	Explanations
Ukrainian Carpathians	Intensive seismicity at Trans- carpathian deep fault forming a RCZB	Rokityansky and Ingerov (1999)	No seismicity over the Carpathian conductivity anomaly epicenters are on the resistive block
Central Japan/Itoigawa– Shizuoka Tectonic Line	Potential region for M8 class intraplate earthquake	Ogawa and Honkura (2004)	Seismicity clusters mainly in the resistive upper crust
Western India/ Koyna Fault	Koyna seismic zone/1967 Koyna earthquake	Sarna et al. (2004)	Koyna fault cuts the high resistive seg- ment in MT model and earthquakes occur in this segment around the fault
USA/San Andreas, creeping central segment at Holister	Microseismicity on MT models	Bedrosian et al. (2002)	A high seismicity clustering, in resistive Franciscan formation.
Tharace Basin/Terzili, Kuzey Osmancik, Kirkclareli Fault Zones	A stable area with low seis- micity, small earthquakes ($M > 4$) around Kirkclareli Fault Zone	Bayrak et al. (2004)	Hypocentral distribution of the earthquakes correlates with highly resistive Istranca Massif
NW Turkey/North Anatolian Fault Zone	Seismically active region. Many microearthquakes and aftershocks	Tank et al. (2003)	Aftershocks 1999 Izmit earthquake Fault Zone occur in the resistive zone.
West Anatolia, Turkey/ West Anatolian graben system	Seismically active region, rapid crustal deformation	Bayrak and Nalbant (2001)	Hypocentral distribution of the earthquakes correlates well with the most resistive element, the twin core structure of the MT section.
SW Anatolia, Turkey/ Fethiye Burdur Fault Zone	Seismically active region	Gurer et al., 2004a, Gurer et al., 2004b	The hypocenters of earthquakes are located in the highest resistivity zones or the lowest conductance zones in the upper crust

Seismicity in conductive zones			
Region/fault zone	Earthquake	Reference	Explanations
Central India	Jabalpur Earthquake M6.0	Waghmare (1997)	Epicentral region of main- and aftershocks correlates well with Jabalpur Paraswada conductivity anomaly
Central Tien Shan/ South Kazakhstan	Earthquake epicentres maximum in Hindu Kush area	Bragin et al. (2001)	Seismogenic zones coincide with subhorizontal areas of elevated conductivity
Northeastern Japan/ northern Miyagi Prefecture	Seismically active Miyagi Prefecture and microearthquakes	Ichiki et al. (1999)	Hypocenters of microearthquakes lie within conductive layer along the layering, near to the RCZB
NW Turkey/North Anatolian Fault Zone	Seismically active region, swarm activity presumably triggered by the Izmit Earthquake	Tank et al. (2003)	The swarm activity is confined in conductive zone. Possible 3-D sea effect due to Peninsula geometry

Table M.1. Comparison of seismicity and resistivity structure for tectonically active zones worldwide as reported by Güler and Bayrak (2007). Note that the abbreviation RCZB stands for resistive and conductive zone boundaries.

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