THE PROBABILITY AND CONSEQUENCES OF THE NEXT ALPINE FAULT EARTHQUAKE, SOUTH ISLAND, NEW ZEALAND

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Frontispiece – Landsat image Pel 104, taken late in the winter of 1974, showing the area of Westland between the Karangarua River in the southwest and the Robinson River in the northeast. The Alpine Fault forms the obvious straight line at the western boundary of the Southern Alps. This satellite image shows the abrupt elevation difference across the fault marked by the pattern of snow cover, the change in the density of valley dissection, and the fault bounded alluvial basins confined to the relatively down-thrown northwest side of the fault.
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PROBABILITY AND CONSEQUENCES OF THE NEXT ALPINE FAULT EARTHQUAKE, SOUTH ISLAND, NEW ZEALAND.

Detailed paleoseismic investigation of the Alpine Fault, South Island, New Zealand, has been undertaken at locations which bracket the central and north sections of the fault, between the Hokitika and Ahaura River. A total of seven trenches and pits have been excavated at four localities along approximately 75 kilometres of the fault. From these excavations a total of 16 radiocarbon dates provide age constraints on the timing of the most recent two earthquakes.

This trenching demonstrates that the most recent rupture occurred after 1660 AD, and most probably around 1700 – 1750 AD. There is consistent evidence for this event in the trenches in the central section of the fault. The surface rupture has extended into the north section of the fault as far as the Haupiri River area, which is 25 km northeast of the Alpine Fault junction with the Hope Fault. An earlier event at around 1600 AD can be recognised throughout the study area, and this is the most recent event in the trench locations north of the Haupiri River.

An updated record of landslide and aggradation terrace ages is consistent with two earthquakes over this period, but this does not significantly refine the estimates of their timing. However, the analysis of indigenous forest age in Westland and Buller reveals two periods of synchronous regional forest damage at 1625 ± 15 AD and 1715 ± 15 AD. I infer that these two episodes of forest damage correspond to the two earthquakes revealed in the trenches for this same time period. Analysis of growth rings in trees which are old enough to have survived these earthquakes indicates that the most recent event occurred in 1717 AD. The growth ring anomalies also indicate a northeast earthquake limit near the Haupiri River. The most recent 1717 AD event
appears to have been a synchronous rupture for a distance of over 375 km, from Milford Sound in the south Westland section of the fault, northeast to the Haupiri River. Based on the forest disturbance record, the earlier earthquake at 1625 ± 15 AD had a rupture length of at least 250 km, but further work is required to determine the southwest and northeast limits of this event.

A range of methods is used here to estimate the probability of the next earthquake occurring on the central section of the Alpine Fault and all the calculated probabilities are relatively high. The most robust method, that of Nishenko and Buland 1987, suggests a conditional fifty-year probability in the order of 65 ± 15%. A sensitivity analysis indicates that the conditional probabilities of rupture are not significantly affected by assumptions regards the exact timing of the last earthquake, or even the number of most recent earthquakes, and conditional fifty-year probabilities of rupture remain at around 50% or higher.

Based on the previous earthquake events, the next Alpine Fault earthquake is likely to have a Moment Magnitude of 8 ± 0.25, and will have a widely felt regional impact. Very strong ground shaking will occur in the epicentral area of the Southern Alps and central Westland. For most of the central South Island the ground shaking is likely to be stronger than that experienced in any other historical earthquake.

Landslides and liquefaction will cause the greatest immediate damage to the natural environment, and in the longer-term increased sediment loads will cause aggradation, channel avulsion, and flooding in the numerous rivers which drain the epicentral region. There will also be substantial and widespread damage to the built environment, in some cases at a considerable distance from the epicentre. Because of the rugged nature of the topography of the central South Island, and the expected regional extent of the earthquake shaking, one of the greatest problems during the post earthquake recovery phase will be difficulty in communication and access.
Chapter 1

INTRODUCTION

1.1 BACKGROUND

The Alpine Fault is the longest active fault in New Zealand with an overall length of more than 800 km. It also has the highest average long term slip rate of any active fault in New Zealand and a total cumulative dextral separation of 460 - 480 km. Despite this there have been no significant moderate or large earthquakes on the Alpine Fault since European settlement in New Zealand and the earthquake data over the last 150 years show that only a small percentage of the strain accumulation across the Alpine Fault has been released seismically. All the indications are that the Alpine Fault is locked and has the potential to rupture in large earthquakes.

An Alpine Fault earthquake is likely to be a regional event felt in all districts of the South Island and in particular has potential to cause serious damage in Westland, Buller, Canterbury and Otago. The importance of possible future Alpine Fault earthquakes in the South Island has been recognised by engineers, geologists and hazard planners for some time (for example Adams 1980; Smith and Berryman 1986; Elder et al. 1991; Pettinga et al. 1998; Stirling et al. 1999) but despite this there has been surprisingly little paleoseismic investigation of the fault. In 1995, when the paleoseismic investigation trenching for this thesis commenced, these were the first paleoseismic excavations along the central and northern section of the Alpine Fault.

It is critical to establish the paleoseismic history of the Alpine Fault as a first step in assessing the probability of a future Alpine Fault earthquake. It is
equally important to consider the likely consequences. Local authorities, emergency management, and infrastructure providers require a realistic scenario for emergency planning, lifeline vulnerability studies and hazard mitigation.

In this thesis I apply a range of geological and paleoseismic investigation methods to determine the likelihood of future Alpine Fault earthquakes and then base an assessment of the likely effects of future earthquakes on the inferred characteristics of past events. Further research will no doubt improve our knowledge of both the probability and the hazard, but this thesis provides a fundamental first step in the assessment and mitigation of Alpine Fault earthquakes in the central South Island.

1.2 OBJECTIVES

This thesis has the following objectives:

• to define the record of Late Holocene earthquakes (and in particular the last 2 – 3 events) on the central and northern sections of the Alpine Fault using a range of paleoseismic indicators.

• to estimate the probability of the next Alpine Fault earthquake

• to bracket the likely range of magnitude and intensity of the next earthquake event

• to consider the probable consequences of the next Alpine Fault earthquake

This chapter is an introduction to the Alpine Fault describing its location, plate tectonic setting, structure, seismicity, and the available geodetic information. Following this is a brief outline of the thesis methodology and a description of the contents of the subsequent chapters.
Following this is a brief outline of the thesis methodology and a description of the contents of the subsequent chapters.

1.3  INTRODUCTION TO THE ALPINE FAULT

1.3.1  Definition and division into sections

The Alpine Fault is part of a transform fault zone through continental crust linking the west dipping subduction zone of the eastern North Island with the east dipping subduction zone of southern Fiordland. Figure 1.1, from Walcott (1998), shows the major plates of the southwest Pacific and the plate tectonic setting of the Alpine Fault.

The Alpine Fault onshore in the South Island is approximately 650 km long and extends from Blenheim, on the northeast coast, to Milford Sound on the southwest coast (Figure 1.2). Berryman et al. (1992) include in their definition of the Alpine Fault the section along the Wairau valley. This northeastern section of the fault, which begins at Blenheim and extends to the Tophouse Saddle near Lake Rotoiti, is also sometimes referred to as the Wairau Fault which can be misleading. At the southwest end, near Milford Sound, the Alpine Fault has recently been mapped extending further offshore approximately 200 km to the Puyssegur Trench area (Barnes, pers. comm., 1999). This increases the overall length of the Alpine Fault to approximately 850 km.

The fault appears clearly on satellite images of the South Island (refer frontispiece photo – Landsat image Pel 104 taken 1974) with a very distinct, apparently straight trace at approximately 055° from Milford Sound to the so called “big bend” near Murchison. This double bend is complete by the Matakitaki River, from where the fault is again straight until the Tophouse Saddle. The transition to the Wairau valley is marked by a 15° strike change to a more easterly strike of around 070°.
Figure 1.1 – The plate boundary of the Pacific and Australian plates with the Alpine Fault (labeled) forming the onshore boundary through the central and southern South Island (from Walcott 1998). The Pacific plate has rotated counterclock-wise relative to the Australian plate about poles (solid squares), the positions of which have changed progressively with time from 5d (17.5 Ma) to 3a (5.89 Ma) thereby increasing the component of shortening. The solid circle is the current NUVEL 1A pole, obtained from seafloor spreading information (De Mets et al. 1994), while the open diamond is the instantaneous pole from direct measurement of current plate motion based on the Global Positioning System (GPS) as determined by Larson et al. 1997.
Berryman et al. (1992) divide the fault into geographic sections (Figure 1.2). They define the most north–eastern extension of the fault in the Wairau Valley as the Wairau section; an adjacent north Westland section from Tophouse Saddle southwest to the Taramakau River; a central Westland section from the Taramakau River to the Haast River; and a South Westland section extending from Haast to Milford Sound. In view of the recent offshore mapping it is now appropriate to add an offshore Fiordland section.

Figure 1.2 – Division of the Alpine Fault into four sections after Berryman et al. (1992), with the addition here of a Fiordland section based on Barnes, pers. comm., 1999. These sections are not necessarily fault rupture boundaries. Figure adapted from Berryman et al. (1992).
These sections were originally defined by Berryman et al. 1992 on the basis of geomorphology and structural style. At that time the authors proposed these sections may also be fault rupture segments but noted the data available to make this inference is very limited.

Bull (1996) also divides the central and northern Alpine Fault into sections using inferred rupture segment boundary. He suggests a rupture boundary exists at the Taramakau River, where the Hope Fault splays off the Alpine Fault, and again at the beginning of the "big bend" of the Alpine Fault. He proposed the names Cook segment for the straight central section, and Brunner segment for the much shorter section to Springs Junction. His inferred segment boundaries are based on obvious structural discontinuities i.e. the splaying at the Hope Fault junction and the curve in strike at the "big bend".

For descriptive convenience this thesis adopts the original geographic division into sections, as outlined by Berryman et al. (1992), but makes the clear distinction that these are not rupture segments. One of the conclusions of this thesis is that rupture lengths and terminations have probably varied considerably during past surface rupture events (refer Chapter 6) and that the most recent Alpine Fault earthquake rupture extended across at least three of these geographic sections.

The work outlined in this thesis has been carried out mainly on the northern and north central sections of the Alpine Fault, extending from the Hokitika River to the Ahaura River (Figures 1.3 and 2.2 in the map pocket). However, some specific investigations, generally to follow up on earlier work by others, has extended to sites as far south as the Karangarua River and north to Springs Junction.
Figure 1.3 – Location diagram of Westland showing the principal faults, rivers and main locations referred to in this thesis.

1.3.2 Regional geology and total fault offset

The Alpine Fault separates the quartzofeldspathic dominated terranes of the eastern province, including the associated Haast Schist terrane, from Paleozoic granites and meta-sediments of the Buller terrane. Figure 1.4 summarises the basement geology of the South Island adopting the terrane classification of Bradshaw 1989.
Figure 1.4 – The basement geology of the South Island, New Zealand adopting the terrane classification of Bradshaw 1989. Map courtesy of Anekant Wandres.
The Alpine Fault is recognised as one of the major strike-slip faults in the world. Wellman (1955a) first noted the lateral shift of the Permian rocks of the Brook Street terrane by the fault from Fiordland to northwest Nelson (Figure 1.4). This is an apparent offset along the fault of approximately 460 - 480 km. While there has been some disagreement over this estimate (see for example Suggate 1963), and others have proposed alternative oblique-slip models resulting in some apparent offset rather than a pure strike-slip separation (Campbell and Rose 1996), most geologists accept at least 350 km of strike-slip movement has occurred along the Alpine Fault but the timing of this movement is still in question. Whereas early workers favoured a Cretaceous initiation (Suggate 1963; Wellman and Cooper 1971), plate tectonic considerations and implications based on the sedimentary record from Cenozoic basins adjacent to the fault currently favour an early Miocene inception (Carter and Norris 1976; Norris et al. 1978; Kamp 1986; Cooper et al. 1987; Walcott 1998).

Wesnousky (1989,1990) introduced the concept of cumulative geologic offset as the major control on the evolution of strike-slip faults and this concept is further discussed in Stirling et al. 1996. In summary these authors demonstrate that, based on Californian, Japanese and Turkish examples, seismological evolution takes place as the fault plane is smoothed by successive offsets. As a result of this structural and seismic evolution the size of major earthquakes increases and the relative frequency of intervening smaller earthquakes decreases.

Table 1.1 presents a summary of the cumulative offset of the major strike-slip faults for which this can be estimated. It is important to note that the Alpine Fault has the greatest cumulative offset of the faults for which this data is known and exceeds all others by a very large margin.
<table>
<thead>
<tr>
<th>Fault and location. (California, USA unless noted otherwise).</th>
<th>Cumulative offset (Kilometres)</th>
<th>Total fault length (Kilometres)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newport – Inglewood</td>
<td>0.2 – 10</td>
<td>60</td>
</tr>
<tr>
<td>Whittier – Elsinore</td>
<td>10 – 15</td>
<td>330</td>
</tr>
<tr>
<td>San Jacinto</td>
<td>24</td>
<td>230</td>
</tr>
<tr>
<td>Garlock</td>
<td>64</td>
<td>240</td>
</tr>
<tr>
<td>Calaveras</td>
<td>24</td>
<td>220</td>
</tr>
<tr>
<td>San Andreas</td>
<td>Approx. 250</td>
<td>1000</td>
</tr>
<tr>
<td>N. Anatolia Turkey</td>
<td>25 – 45</td>
<td>900</td>
</tr>
<tr>
<td>Neodani, Japan</td>
<td>3 – 5</td>
<td>100</td>
</tr>
<tr>
<td>Atera, Japan</td>
<td>7 – 10</td>
<td>60</td>
</tr>
<tr>
<td>Atotsugawa, Japan</td>
<td>3</td>
<td>60</td>
</tr>
<tr>
<td>Porters Pass Fault Zone, New Zealand *</td>
<td>&lt; 2</td>
<td>100</td>
</tr>
<tr>
<td>Hope Fault, New Zealand *</td>
<td>20</td>
<td>200</td>
</tr>
<tr>
<td>Alpine Fault, New Zealand</td>
<td>&gt; 350 and probably 460 – 480 km</td>
<td>650 km on land (850 km inclusive of the offshore section)</td>
</tr>
</tbody>
</table>

**Table 1.1**: A comparison of cumulative offset for major strike-slip faults in California, Turkey and New Zealand. (data from Wesnousky 1989 and Stirling *et al.* 1996, with additional data denoted by [*] from Cowan *et al.* 1996).

It is conceivable that the extremely large cumulative offset of the Alpine Fault is one of the explanations of the relatively low instrumental seismicity along the fault which is noted in the next section. This may also explain why the Alpine
Fault seismicity most closely resembles sections of the San Andreas fault, the fault with the next largest known offset (Leitner et al., submitted; Eberhart-Phillips 1995). This also implies that when Alpine Fault earthquakes do occur they will have longer rupture lengths than might be expected for other faults and that segmentation may play a relatively minor role in arrests of rupture propagation.

1.3.3 Plate motion, crustal structure and seismicity

De Mets et al. (1990) propose the NUVEL 1 global kinematic model of plate motion. This model, modified in 1993 as NUVEL 1A to incorporate new information on dating (De Mets et al. 1994), has become the standard reference for plate kinematics (see for example Walcott 1998). The NUVEL 1A pole position for the Pacific Plate is shown in Fig 1.1. At most locations along the central section of the Alpine Fault (for example Franz Josef, 43.5°, 170°) this pole location predicts plate rates of around 37 mm/yr at an azimuth of 071°. For most of the central section of the Alpine Fault this can be resolved into components of approximately 36 mm/yr parallel to the average strike of the Alpine Fault and 10 mm/yr normal to it.

More recently estimates of pole position from the Global Positioning System (GPS) have become available (Larson et al. 1997) which differ in position from the NUVEL 1A pole by around 5 degrees of latitude further south (Figure 1.1). Because the pole is relatively close to New Zealand a small change in its position results in a large change in the computed relative motion along the Alpine Fault plate boundary. The GPS pole indicates a plate rate of 44 mm/yr at an azimuth of 77 degrees. This in turn increases the fault parallel rate to around 43 mm/yr and the fault normal rate (shortening) to approximately 15 mm/yr.

It is likely that the true rate is somewhere within this range. For example Walcott (1998) adopts a long-term average rate based on the position
computed from the finite rotation of anomaly 3a (5.9 Ma) [Figure 1.1]. This is intermediate between the Nuvel 1A and GPS rates with approximately 40 mm/yr of fault parallel motion and 13 mm/yr of fault normal shortening.

Models for the crustal structure in the South Island (for example Norris et al. 1990) show the Alpine Fault as the westward boundary of a zone of distributed deformation formed by the oblique convergence of continental crust of the Pacific and Australian plates. A two-sided deforming wedge model has been proposed which views the Alpine Fault as dipping east under the Southern Alps and Canterbury, accommodating crustal delamination of the upper 20 - 25 km of the Pacific plate on a ductile detachment, with associated west dipping back thrusting. Figure 1.6 shows the model originally proposed by Norris et al. (1990) and subsequently modified by Pettinga et al. 1998.

![Diagram](image.png)

**Figure 1.5** – Schematic diagram of the two sided deforming wedge model representing the oblique continent-continent collision zone of the Australia-Pacific plate boundary across the central South Island. Note the Alpine Fault forming the western limit of the deforming orogen and the contact with the Australian plate which can be viewed in terms of critical wedge mechanics as the indentor. Model proposed by Norris et al. (1990). Figure modified from this by Pettinga et al. (1998). The red colour represents lower crustal rocks.
Support for this model comes from recorded microseismicity under the Southern Alps (for example Reyners 1987; Eberhart – Phillips 1995) and the existence of gravity anomalies (Allis 1986; Woodward 1979). Leitner et al. (submitted) also outline supporting evidence from recorded seismicity, but note this relatively simple model may only be realistic for the section of the Southern Alps south of Mount Cook, which is the area beyond the influence of the Marlborough fault system.

![Diagram of crustal structure in central South Island](image)

**Figure 1.6** – A cross-sectional model of the crustal structure in the central South Island after Kleffman et al. 1998. Note the crustal thickening in response to shortening which commences approximately 80 km east of the Alpine Fault. The Alpine Fault appears as a broad southeast dipping fault zone.
Two comprehensive geophysical transects have recently been completed across the central South Island near Mt Cook and the Rangitata River. The results of the Mt Cook seismic lines are summarised in Kleffman et al. 1998. They show that greywacke crust, which is approximately 25 km thick at the east coast of the South Island, begins to thicken approximately 80 km from the Alpine Fault to reach a maximum depth of at least 35 km (Figure 1.6). The Alpine Fault appears as an easterly dipping zone, approximately 7.5 km wide, to a depth of around 25 km. They infer this wide zone has a low seismic velocity but cannot directly measure this.

Crustal thickening of the Pacific Plate, by up to 20 km, has also been identified by Stern et al. 1997 and Wilson and Eberhart-Phillips 1997. The recent work indicates a probable average dip for the Alpine Fault of around 45 degrees but lower dips of 35 – 40 degrees have also been proposed (Stern et al. 1998). A dip for the fault of around 45 degrees on geological grounds had previously been proposed by Sibson et al. 1979 and Norris et al. 1990.

The implications of this dip in relation to future large Alpine Fault earthquakes include a likely offset in earthquake epicentre eastward and away from the fault trace geometrically related to the dip and depth of focus for the event. Given that moderate and large earthquakes tend to nucleate at the base of the seismogenic zone (Scholz 1990; Kramer 1996) a 45 degree dip will result in an offset of the earthquake epicentre east by a distance approximately equal to the depth of the seismogenic zone.

In addition the potential fault rupture area above the brittle - ductile transition (which governs the release of earthquake energy, see for example Kramer 1996) is increased proportionally by the dip of a fault, thereby increasing the seismic moment. For example a 45 degree dip increases the seismic moment in comparison to an equivalent length of vertical fault by approximately 40%.
However, for the Alpine Fault this effect may be offset somewhat by a relatively shallow depth to the brittle – ductile transition, at least in local areas. The region of highest uplift along the central section of the fault is associated with a high thermal gradient (Allis and Shi 1995; Koons 1987; Shi et al. 1996). Thermal modelling (Allis and Shi 1995; Batt and Bruan 1999; Shi et al. 1996), data from fluid inclusion studies (Craw 1988; Craw et al. 1994; Holm et al. 1989; Jenkins et al. 1994), zircon reset ages (Tippett and Kamp 1993) and heat flow measurements (Funnel and Allis, submitted) all predict a thermally weakened crust. This implies a shallow depth to the brittle - ductile transition which, based on the various models, has been estimated to be as shallow as 5 - 8 km (Holm et al. 1989; Jenkins et al., 1994; Shi et al., 1996). This in turn has led to speculation that the central section of the Alpine Fault has insufficient volume of brittle rock to accumulate the elastic strain necessary to generate a large earthquake (Walcott 1998; Stern et al. 1998).

Recorded seismicity is the most reliable guide to the depth of the brittle – ductile transition and is preferable to inferences based on modeling. Seismicity in the central South Island is nearly all confined to the crust (Allis and Shi 1995; Reyners 1988; Reyners et al. 1983; Rynn and Scholz 1978; Scholz et al. 1974) with the exception of a small number of 50 – 100 km deep earthquakes beneath the Southern Alps (Reyners, 1987). Figure 1.7 shows the general pattern of seismicity across New Zealand for the period 1990 – 1994. Deeper events are confined to the subduction zones to the north of the Alpine Fault (Hikurangi Trough) and south near Fiordland (Pusegur Trench).

The traditional view of recorded seismicity along the Alpine Fault has been of anomalously low levels in relation to the importance of the structure. This led to the suggestion by Adams (1980) that it is a "seismic gap" comparable to other sections of plate boundary where large earthquakes have been predicted after relatively low recorded seismicity (e.g. Sykes 1971; Sykes et al. 1981; Ward and Page 1989; Scholz 1990).
Figure 1.7 – New Zealand seismicity at depths of 39km or less for the period 1990 – 1994. From Aitken and Webb (1998).
Figure 1.8 - Three years of seismicity with a range in magnitude from $M = 2$ to $M > 5$ for the period 22 March 1991 to 30 April 1994 for events with depths less than 40 km. Note the occurrence of small earthquakes along the fault. The cross-section H-H' has been constructed by projecting epicentres recorded over the length of the fault to a single cross-section and shows the depths of events of various magnitudes (circles = $M = 2 - 3.9$; stars = $M = 4 - 4.9$; filled diamonds = $M \geq 5$). The Alpine Fault is at zero on this section and this data suggests a fault plane dipping east. Figure from Eberhart-Phillips (1995).
However, recent improvements in the seismograph network and the deployment of portable networks demonstrate that more seismicity is occurring than was being recorded in the old network (Eberhart-Phillips 1995). Figure 1.8 shows the recorded seismicity along 450 km corridor parallel to the Alpine Fault over the three years between 1991 and 1994.

Over this time period there were 9 earthquakes per year along the fault ranging from magnitude 2.5 to 3.6. Focal depths for all recorded events range from 10 km at the Alpine Fault to more than 20 km at distances of up to 60 km east of the fault, and the data suggest an east dipping fault plane.

While the level of recorded seismicity is still relatively low Eberhart – Phillips (1995) noted that it is comparable to the Mojave section of the San Andreas Fault, which last ruptured in 1857, and is estimated by Sieh et al. (1989) to have generated at least 10 large earthquakes in the last 1400 years.

While the work of Eberhart – Phillips (1995) suggests a brittle – ductile transition along the Alpine Fault at depths of 10 km or more, Walcott (1998) criticises this evidence on the basis that the study may be excessively influenced by recorded events at the extreme ends of the study region (i.e. north and South Westland sections of the fault).

However, important new constraints for depth to the brittle – ductile transition zone come from recent precise determination of earthquake depth (Leitner and Eberhart - Phillips 1998; Leitner et al. submitted). Four complementary data sets are used to evaluate the seismicity of the central South Island. The Southern Alps Passive Seismic Experiment (SAPSE) gives unprecedented high-quality earthquake locations but over a relatively short time period (6 months from November 1995 – April 1996). The upgraded New Zealand National Seismic Network and Lake Pukaki network provide insight into the long-term seismicity over the last 8 years. In addition the September 1997 Mt Cook earthquakes, consisting of two main events of $M_{l}$ 5 and numerous
associated aftershocks, was recorded by both the national network and temporary aftershock deployment of three stations. This provides useful information in the area of the Southern Alps with the estimated highest uplift.

Leitner et al. (submitted) conclude:

- the maximum depth of crustal seismicity is uniform over large parts of the South Island at about 10 – 12 km

- The SAPSE data, which include 60 earthquakes in a band 5km northwest to 15km southeast of the surface trace of the Alpine Fault, outline its seismogenic zone. The data indicate the seismogenic depth of the Alpine Fault is about 10 – 12 km. The seismic data do not confirm wide spread high temperatures at the Alpine Fault or the deepening of the 350° isotherm, as suggested by Shi et al. 1996. While some local areas with a more shallow brittle – ductile transition can be recognised, these constitute a relatively small proportion of the total fault length.

- The seismicity rate on the Alpine Fault is comparable to locked sections of the San Andreas Fault and there is potential for large earthquakes on the Alpine Fault.

At this stage it is not possible to completely resolve the question of the likely focal depth for large Alpine Fault earthquakes, but on the basis of the micro-earthquake evidence discussed above a focal depth of around 10 km is adopted for the modeling presented in Chapter 7.

1.3.4 Near-surface Alpine Fault structure and long-term slip rates

The dominant fault-parallel component of plate motion, in combination with a significant shortening normal to the fault, results in an obliquely transpressive sense of movement at the fault trace. A combination of dextral strike-slip
movement and oblique thrusting develops in the near surface (i.e. in the upper few kilometres) where the relative amounts of shortening and strike-slip movement vary slightly with subtle changes in fault strike. For example Norris and Cooper (1995) show segmentation of the fault at a detailed scale of several kilometres, into dominantly strike-slip segments with a more east - west strike, and relatively more north aligned oblique thrust segments. They suggest this is a near surface phenomena resulting from perturbations in the stress field in the upper 1 - 4 km created by valley incision (Figure 1.9).

**Figure 1.9** - Conceptual view of near surface segmentation of the Alpine Fault into alternating right lateral strike-slip fault traces and oblique thrust traces typical of the fault near Franz Josef (after Norris and Cooper 1995).
Horizontal Fault Parallel Slip Rates

The rate of horizontal slip parallel to the Alpine Fault is not well constrained. This reflects the absence of suitable offset markers and piercing points which can still be recognised on both sides of the fault. Either deposition on the relatively downthrown western side, or erosion on the uplifted eastern side, tends to remove one or other part of the matching offset. Plate tectonics implies the rate cannot be much more than 40 mm/yr at the most active central section and this maximum assumes all the horizontal plate motion is concentrated on the Alpine Fault. Given the presence of active faults with strike-slip movement further to the east, in particular the Porters Pass Fault Zone (Cowan et al. 1996), it appears that at least some of the relative plate motion is accommodated away from the Alpine Fault. This is supported by the recent GPS (Global Positioning System) data discussed below in section 1.3.5.

There are a number of minimum fault parallel slip rate estimates which indicate at least 25 mm/yr of movement along the central section (for example Berryman et al. 1992; Cooper and Norris 1995). More recently Norris and Cooper (1997) have derived a slip rate over the last 20,000 years of 22 - 30 mm/yr at Waikukupa near Franz Josef. Combined with the earlier minima the likely range reduces to 25 - 30 mm/yr.

The alternative way to estimate the slip rate for the central section of the fault is sum the average slip on the Hope Fault with estimates of the slip on the north Alpine Fault. Estimates of total slip for the Hope Fault zone since the end of the last glaciation are approximately 25 ± 10 mm/yr (Cowan 1989; Cowan 1990; Van Dissen and Yeats 1991) and all of this movement appears to transfer to the Alpine Fault between the Taramakau River and the Toaroha River.

The north Alpine Fault average slip is estimated by Berryman et al. (1992) to be 10 ± 2 mm/yr at Inchbonnie which is at the extreme southern end of the north section. During this thesis work I have obtained an average slip rate of 6
± 2.5 mm/yr for this section of fault at the Haupiri River, 25 km further north, but where more Alpine Fault movement may have transferred to parts of the Marlborough fault system (Chapter 2, section 2.3). It should be noted that both these estimates of fault parallel slip rate are from very young features (i.e. 1300 years and 2000 years respectively) and may not necessarily be representative of the long term rates. However, this remains the best available data.

The sum of these north Alpine Fault and Hope Fault slip rates is 35 ± 12 mm/yr and 31 ± 12.5 mm/yr respectively. Unfortunately the large errors associated with the Hope Fault estimates limit the value of this approach, but it does suggest that the upper limit of 30 mm/yr for the Alpine Fault is the most realistic end of the range.

This field evidence of slip rates collected over the last decade indicates that the Alpine Fault accounts for 75 - 90% of the horizontal component of the long term plate motion rate and refutes a line of reasoning first proposed by Walcott (1979). Walcott suggested that the Alpine Fault accommodates less than one third of the plate motion, the remainder being taken up by aseismic permanent strain in the Southern Alps. Walcott discounted regular earthquakes on the Alpine Fault on the basis that the observed fault offsets of young surfaces were not large enough. At that time only 6 - 13 mm/yr of horizontal fault parallel slip could be demonstrated (Berryman 1979).

Despite the progressive recognition by field workers that the slip rate is actually much higher (see for example the discussion in Berryman et al. 1992) the notion that insufficient elastic strain accumulates along the fault to produce large earthquakes persists in the thinking of some geologists.

**Vertical Fault Slip Rates**

The Southern Alps are the consequence of the compressive component of plate motion operating across the fault and there have been a number of
models of regional uplift proposed (see for example Walcott 1979; Wellman 1979; Norris et al. 1990; Walcott 1998). The rate of uplift of the Southern Alps has been estimated by Walcott (1979) and Wellman (1979) at 22 mm/yr and 17 mm/yr respectively although these estimates were based on the early estimates of a more northern pole of plate rotation.

Uplift along the Alpine Fault plays a major part in the Southern Alps uplift rate, and probably the largest part. Suggate (1968) estimated uplift rates of around 12 mm/yr at Paringa but subsequent investigation indicates locally high uplift rates in parts of this area and a lower background regional rate of 7 – 8 mm/yr (Simpson et al. 1994). The recent estimates of vertical uplift rate at the Alpine Fault have all been much less than the Southern Alps estimates and consistently less than 10 mm/yr. For example Bull and Cooper (1986); Berryman et al. (1992); Simpson et al. (1994); and Yetton and Nobes (1998) all estimate rates of around 6 - 8 mm/yr for the central section. Vertical uplift rates are discussed in more detail in Chapter 2 (section 2.4.1) and again in Chapter 5 (section 5.3).

1.3.5 Geodetic and GPS surveys

Resurveys of the 100 year old geodetic survey networks (Walcott 1978; Walcott 1984; Wood and Blick 1986) indicate elastic strain is accumulating over a broad zone in the rocks on each side of the Alpine Fault at rates consistent with the plate tectonic predictions. It is appears this accumulated elastic strain represents potential energy in storage for the next Alpine Fault earthquake.

Berryman et al. (1992) highlight the surprising consistency of the magnitude of the strain rate across the fault from South Westland to the Wairau section, despite the implied reduction in slip rate to the north. They also note that these rates alone, in conjunction with the definite absence of Alpine Fault earthquakes since at least 1840, suggest the observed strains can only
continue to accumulate for one or two hundred more years before rock strength is exceeded and an earthquake occurs.

Aseismic creep has not been observed along any part of the Alpine Fault. The Alpine Fault trace crosses sealed roads in at least 10 locations and a concrete monitoring wall was constructed across the trace at Springs Junction in 1964 (Beanland 1987). Beaven et al. (1998) report that small scale survey networks crossing the Alpine Fault indicate no significant near-surface aseismic fault slip on the central Alpine Fault over the past 25 years.

There are early reports in the New Zealand Geological Survey files (Bowen 1957) of power lines requiring loosening over a 20 year period in the vicinity of the fault trace at Rocky Point, near Inchbonnie. However, no deformation of the two adjacent roads each side of this site which cross the fault was recorded. The poles concerned are in the active bed of the Taramakau River and it is also possible gravel movement was causing this.

More recent enquiries with TransPower Ltd, who have inherited the transmission network, indicate no further problems in this area. However, it should always be recognised that in many locations aseismic movement of the fault would not be recognised due to the actively changing landforms and the extensive forest cover.

Recently the first GPS (Global Positioning System) resurveys of the original triangulation network have been carried out between Christchurch and Hokitika (Pearson et al. 1995). The observed strain rates over a period from 1978 to 1992 are once again consistent with the De Mets et al. (1994) plate velocities, and indicate about two thirds of the plate motion is being taken up as elastic strain in the vicinity of the Alpine Fault. Further strain is accumulating east of the Alpine Fault in the Porters Pass - Amberley Fault Zone (Cowan et al. 1996), but this in turn may ultimately join the Alpine Fault, possibly in the vicinity of the Whataroa River (Anderson and Webb 1994).
More recently Beaven et al. (1998) and Beaven et al. (submitted) have reported on the GPS resurveys from 1994 to 1998. Their results confirm more than 70% of the plate strain is occurring within a band from 5 km northwest to 20 km southeast of the Alpine Fault. However, significant strain continues a further 60 km southeast to the Southern Alps foothills and the Porters Pass Fault Zone. Figure 1.10 shows the plate velocities as measured by GPS for the period since 1994 to 1998.

![Figure 1.10 - Relative velocities measured by GPS resurvey for the period 1994 – 1998 with a reference frame of a fixed Pacific Plate. Note the Alpine Fault appears to be locked and the relative strain is primarily confined to a zone which commences at approximately 20 km southeast of the fault. Additional more minor relative strain continues to the Canterbury foothills, a distance of around 60 km. Data from Beaven pers. comm. (1999).](image-url)
The surface displacements in the high strain rate area are well fitted by a model in which 65 - 75% of the relative plate motion is accommodated by slip below 7 - 9 km depth on a southeast dipping Alpine Fault, with material above this behaving elastically and thus storing elastic strain in the region of the Alpine Fault (Beaven et al. 1998).

1.3.6 Field evidence that Alpine Fault ruptures are accompanied by earthquakes

Because of the absence of historic earthquakes and the early views of Walcott (1979) regarding aseismic permanent strain, the view has persisted among some geologists that the Alpine Fault moves in some way which does not cause earthquakes. The historic and geodetic evidence against aseismic fault creep is outlined above, and this is the only known alternative to seismic slip. However, there are also two strong additional arguments which indicate that when Alpine Fault movement does occur it is associated with earthquakes.

The first is the abundance of pseudotachylite, a dark glassy material produced by friction melt during seismic slip. Wallace (1976) first recorded this in the Alpine Fault zone and later Sibson et al. (1979) described the distribution in more detail, noting it as "the product of seismic slip on discrete planes". It is particularly abundant in the fault rocks of the central section of the fault.

The second argument for past seismic events is the presence of liquefaction induced features in the trenches excavated adjacent to, and across the fault. The first trench to show this is one of the trenches in this study (the Kokatahi 2 trench, refer section 3.2.2) but Wright (1994) earlier noted liquefaction induced folding and clastic dykes in tilted terrace gravels 500m from the fault near this same location. Recent trenches excavated across the Alpine Fault trace at Haast also contain evidence of liquefaction in the surficial sediments at the fault zone (Berryman, pers. comm. 1999). Liquefaction is caused by an abrupt
increase in pore water pressure associated with shear stress loading at a rate which does not allow drainage of near surface saturated fine sands and coarse silts. As such it is a very good indicator of large earthquakes, as opposed to slow aseismic deformation during which drainage is possible.

Many of the trench face logs presented later in Chapter 3 depict intense deformation (i.e. contorted beds and tight folding) and clear truncation of the sedimentary units. This is strongly suggestive of sudden rupture, as opposed to slow progressive deformation, but is not as definitive as the presence of liquefaction structures.

1.4 HISTORIC EARTHQUAKES AND PREVIOUS PALEOSEISMIC INVESTIGATIONS

Possible historic and pre-European Earthquakes

There has definitely been no rupture on any section of the Alpine Fault in the 160 year record since 1840. While it is possible an Alpine Fault earthquake may have gone unrecognised between 1800 and 1840, sealers and whalers were regularly working in the region which makes this unlikely. Sealers working from Dusky Sound, Fiordland, did record a series of strong earthquakes in 1826 (McNabb 1907). Severe landscape disturbance was observed following these events northeast of Dusky Sound at the Cascade River area (50 km southeast from Haast, refer Figure 1.2) as the following description by R. Taylor records:

"Beyond Cascade Point the whole coast presented a most shattered appearance, so much so that its former state could scarce be recognised. Large masses of the mountains has fallen, and in many places the trees might be seen under water"

McNabb 1907, page 262
Despite the relatively close proximity of the shaking and damage to the South Westland section of the Alpine Fault these earthquakes are generally attributed to events in the offshore Fiordland area (Norris pers. comm. 1996; Berryman pers. comm. 1997). This view is supported by paleoseismic research on the Alpine Fault near Milford Sound (Cooper and Norris 1990) which suggests that the most recent Alpine Fault earthquake event at this most southern on-land location occurred between 1650 and 1725 AD. Given the recent recognition of the submarine continuation of the Alpine Fault offshore from Milford Sound a further 200 km (Barnes pers. comm. 1999) the possibility of the 1826 earthquakes being associated with this submarine section of the fault cannot be discounted.

Enquiry through the Ngai Tahu liaison officer (Department of Conservation, Hokitika) indicates no known Maori oral tradition of a large earthquake in the Arahura area near Hokitika, despite the traditional importance of the area for Pounamu (greenstone) gathering. There may have been early tribes in the area from the sixteenth century (e.g. the Waitaha) but Ngai Tahu became progressively established from the seventeenth to eighteenth century (Belich 1996).

The absence of an oral tradition of earthquakes does not preclude an event having occurred after Ngai Tahu began using the area, because it is possible the impact was minor for the early gathering parties, and a permanent local population may not have established for some time. The oral record may also be incomplete, particularly for events well before 1800 AD. Potton (1987) notes that Brunner's census of 1847 recorded a total of only 99 Maori living north of latitude 44° N (Cascade Point, south of Jackson Bay and 50 km south of Haast). Raids by northern tribes a generation earlier had decimated the local inhabitants which may also have wiped out many oral traditions. However, the absence of an oral tradition does at least suggest that the last large
earthquake did not occur in the 2 - 3 generations immediately prior to the first European contact in the 1840's.

*Previous paleoseismic investigations*

Paleoseismic investigation of the central and north Alpine Fault has been relatively neglected to date considering the fundamental importance of the fault to the seismology, geological structure and earthquake hazard in the South Island. This reflects in part the poor exposure of the fault, the very rapid active processes which tend to mask it, and the difficulty of working in such heavily forested terrain. This thesis is primarily a paleoseismic investigation and I briefly discuss here the work which has been done by others prior to this study on the past earthquake history of the Alpine Fault. In Chapter 6 (section 6.5) there is a more detailed comparison of their conclusions with the results of this thesis.

*Adams (1980)*

The earliest research on the paleoseismic history of the Alpine Fault is the work of Adams (1980). He collects together ten \(^{14}\text{C}\) dates from landslides and aggradation terraces in central and South Westland, some dates from landforms which he has identified, and some dates from previous researchers. He presents the apparent coincidence of some of the radiocarbon dates as indirect evidence of the timing of large earthquakes on the Alpine Fault (Figure 1.11).

Two or more dates which coincide within the radiocarbon dating error are considered by Adams to be sufficient evidence to infer a series of earthquakes at approximately 500 year intervals over the last 2000 years, with the most recent event around 550 years ago. However, Adams noted that the record may still be incomplete and that future dating may reveal intermediate age earthquakes.
Figure 1.11 - Adams (1980) radiocarbon date distribution from aggradation terraces and landslides in Westland over the last approximately 2000 years. Note the relatively small number of dates and the inferred earthquakes (*) on the central Alpine Fault at around 500 year intervals with the most recent around 550 years ago (1450 AD).

Cooper and Norris (1990)

Cooper and Norris (1990) investigated Alpine Fault paleoseismicity along the South Westland section of the fault near Milford Sound. This involved the $^{14}$C dating of material excavated from sag ponds near the fault scarp and estimates of the age of trees which appear to have lost their crowns as a result of earthquake shaking. The tree age estimates are based on circumference as opposed to the more reliable increment corer method which samples actual tree rings.

They conclude that the last large earthquake in this area due to movement of the Alpine Fault occurred in the period between 1650 AD and 1725 AD. An incomplete record suggests a possible earlier event around 2000 years ago.
This work of Cooper and Norris (1990) has the important advantage of being a more direct investigation than the previous work in the central area of Adams (1980), and the subsequent work of Bull (1996).

**Bull (1996)**

Bull infers a quite different pattern of past earthquakes on the central Alpine Fault to that of Adams (1980). Bull's approach is based on the lichenometric dating of rockfalls. These rockfall sites were all well east of the fault, the closest being approximately 18 km away, and the majority more than 25 km. The dating method is based on the assumption that lichen growth rates are uniform on new rock surfaces created during rock falls. Lichens of the genus *Rhizocarpon* subgenus *Rhizocarpon* apparently grow in the Southern Alps at the same uniform growth rate of c. 0.17 mm/yr in the altitude range between 400 and 1600m, independent of substrate, microclimate or aspect.

While the dating technique has been criticised (refer for example Oelfke and Butler 1985; McCalpin 1996), Bull is able to demonstrate a correlation between the historical earthquake record for the Hope and Kakapo faults and the lichen record.

Bull infers regional peaks in lichen size modes of 43 mm, 84 mm and 125 mm are the result of earthquakes on the Alpine Fault at 1748 ± 10 yrs AD, 1489 ± 10 yrs AD and 1226 ± 10 yrs AD respectively. A less distinct size mode at 166 mm also suggests an event around 967 ± 10 yrs AD. The implied recurrence interval is a remarkably constant 261 ± 14 years (Bull 1996).

**Other current work**

During final compilation of this thesis, and after completion and distribution of the EQC funded project upon which this thesis is based (Yetton et al. 1998), the results of two subsequent paleoseismic investigations have been
published. Both are discussed in more detail in Chapter 6 (section 6.5). These are:

- Wright et al. (1998) and Wright (1998) who outline recent paleoseismic investigations at the Waitaha River, in the central Alpine Fault, which have involved trenching of sag ponds and dendrochronology. They infer at least 5 ground rupturing earthquakes in the last 1500 years, four within the last 900 years and the most recent around 1720 ± 5 AD.

- Berryman et al. (1998) describes trenching in the Haast area along the South Westland section of the fault. They report clear evidence of three ground rupturing earthquakes in the last 900 years, each of around 8 m of dextral slip. They consider the last event was probably at 1718 ± 5 AD.

### 1.5 Methodology and thesis outline

This thesis presents the results of an investigation of four independent potential paleoseismic indicators in the central and northern sections of the Alpine Fault. These include:

- seven paleoseismic trenches and pits excavated in various locations across the central and north sections of the Alpine Fault from which radiocarbon dates have been obtained.

- radiocarbon ages collected from aggradation terraces and mass movement deposits.

- forest stand ages indicating periods of synchronous regional forest disturbance and re-establishment.

- synchronous tree ring disturbance in long living tree species which have survived previous Alpine Fault earthquakes.
After a general description in Chapter 2 of the Alpine Fault trace and trenching sites from the Hokitika River to the Ahaura River, these four independent paleoseismic indicators are discussed individually in Chapters 3 - 5. They reveal a consistent paleoseismic record for the last 750 years. This is summarised in Chapter 6, along with the available information on the locations and lengths of rupture of previous earthquake events. In Chapter 7 this paleoseismic record is then used to calculate the probability of the next Alpine Fault earthquake. Inferences are made regards the magnitude and shaking intensity of the previous earthquakes in Chapter 8, and these are used to bracket the likely seismic characteristics of the next earthquake. In Chapter 9 the general consequences of the next earthquake are discussed. Finally Chapter 10 outlines the thesis conclusions and general recommendations.

This thesis builds on an earlier report (Yetton et al. 1998) prepared for the Earthquake Commission (EQC) and the local authorities and infrastructure providers who have assisted to varying degrees with research funding. All those providing financial support are listed in full in the acknowledgements. Various updates of progress during this research are summarised in Wells et al. (1999); Yetton (1998a); Yetton (1998b); Yetton and Wells (1998); Yetton and Nobes (1998); Yetton (1997) and Yetton (1996).
Chapter 2

NATURE OF THE ALPINE FAULT TRACE FROM THE HOKITIKA RIVER TO THE AHAURA RIVER

2.1 INTRODUCTION

The Alpine Fault forms the northwest range front of the Southern Alps. It separates this rising mountain system from isolated hills on the relatively downthrown northwest side, which are partially buried by fluvial and fluvio-glacial sediments. Despite high rates of erosion and deposition associated with the rising range front, the Alpine Fault is marked by a discontinuous Holocene and Late Holocene active trace along most of its length.

This chapter describes the Holocene and Late Holocene Alpine Fault trace over an 85 km distance from the Hokitika River in the central Westland section of the fault, to the Ahaura River in the north Westland section (Figure 2.1, back pocket). The fault trace in this area has been mapped and examined in detail as part of this project to select suitable sites for more detailed paleoseismic investigation.

The aim of this field work was not to carry out structural or general geological mapping of the Alpine Fault, but instead to focus on locating and defining the most promising paleoseismic sites. Limited parts of the area have already been mapped from a structural and geological perspective (for example Wright 1994; Rattenbury 1986; Rattenbury 1987; Angus 1984). Berryman et al. (1992) describe local details of the trace between the Taramakau River and Haupiri...
River while more regional geological mapping of the general area is included in Bowen 1964; Gregg 1964; and Warren 1967.

2.2 VARIATIONS IN FAULT TRACE CHARACTER ALONG STRIKE

The Alpine Fault between the Hokitika River and Ahaura River forms a prominent, forest covered, northwest facing range front trending approximately 050° – 060°. The Southern Alps rise very steeply on the southeast side and within 5 – 10 km of the fault these mountain ranges commonly reach elevations of 1500 – 2000 m.

On the northwest side the relief alternates between alluvial fan surfaces at elevations of around 100 – 200 m, and higher isolated hills, some of which reach over a 1000 m in elevation.

Figure 2.1 (back pocket) shows the Holocene and Late Holocene active trace of the Alpine Fault based on the geomorphic expression in the field and in aerial photos. In most cases this surface trace has formed in Holocene and Late Holocene fluvial and colluvial materials so that in most locations there is no bedrock definition of the fault.

In Figure 2.1 the trace defines a fault zone with an average strike of approximately 050° for a 35 km distance between the Hokitika River (GR J33/465000) and Lynch Creek (GR K33/745240). In some areas this section of the fault trace has similar characteristics to the fault trace further southeast as described by Norris & Cooper 1995. They observed segmentation of the fault trace on a scale of 1 – 10 km, with oblique thrust sections striking 020° – 050°, that are linked by dextral strike-slip fault traces striking between 065° and 090°. They attribute this segmentation to the influence of deeply incised valleys in the hanging wall that disturb the stress field to depths of 1 – 4 km.
In Figure 2.1 this type of variation in trace character is most apparent in the area around the Toaroha and Kokatahi Rivers. These two rivers join near the fault trace and combine to form the largest valley along this section of the fault. The details of this segmentation, and the nature of the fault trace in this area, are discussed in more detail in the next section.

The fault trace exhibits a significant change of average strike at Lynch Creek. The strike changes by approximately 10° from around 050° to around 060°, and this average strike is then maintained through the area northeast of Lynch Creek for more than 50 km to the Ahaura River.

This change of strike also corresponds to a change in fault trace character. There is no longer any evidence of the segmentation pattern observed further southwest, and there is no clear relation between local fault strike and the style of deformation at the fault trace. Instead the fault trace maintains a consistent strike commonly separated by right step-overs on scales of 10 – 100’s of metres. For example at the Taipo River (GR K33/785270) a right step-over of about 100 m occurs between strands that overlap by about 1km. This pattern is repeated on a smaller scale 5 km further northwest at Inchbonnie (Berryman et al, 1992) and at several other locations in this general area.

The change in fault character in the area northeast of Lynch Creek is presumably linked to the change in strike. By aligning to a more east-west orientation in this area, the Alpine Fault trace is less oblique to the plate motion vector, and more closely parallel to the Marlborough Fault System. By contrast the increase in contraction across the fault southwest of Lynch Creek increases both the rate of uplift at the fault and the tendency for the fault to segment in response to incisions in the hanging wall.
2.3 SELECTION OF SITES FOR PALEOSEISMIC INVESTIGATION

In several locations within the study area large rivers cross the fault. These have created flights of aggradation and degradational terraces which provide a potentially useful late Pleistocene and Holocene chronology which allows a detailed study of the fault zone geometry and the paleoseismic history.

At such locations it is important to investigate the fault zone in sufficient detail to confidently identify the youngest traces on the Late Holocene fluvial surfaces where critical data include the timing of the most recent rupture event to provide estimates of elapsed time. It is also important to identify, and generally avoid, those sections of fault trace which may be associated with other subsidiary branch faults or less active back-thrusts.

Not all the large rivers provide suitable potential paleoseismic sites. In some cases, for example the Arahura River, active side streams and associated Late Holocene outwash fans have buried the fault trace at the critical locations. In other cases a fault trace may be present but there are no stratigraphic constraints to indicate a definite Late Holocene age.

Once an area which has a clear Late Holocene Alpine Fault trace has been identified, the key requirement is to determine the relative and absolute ages of the various geomorphic and stratigraphic markers which may be offset in the fault zone. This normally requires the presence of suitable organic material within the subsurface soil materials, either located in natural exposures or paleoseismic trenches, for radiocarbon dating.

In Figure 2.1 (map pocket) a total of 8 sites are shown which, after fault trace mapping, I had initially identified as having further potential for paleoseismic investigation. Of these 8 sites, trial excavations of the Late Holocene fault trace at Big Wainihinihi River, Taipo River, and the southwest bank of the Haupiri River, identified no subsurface datable material.
However, a total of seven excavations at five other locations shown in Figure 2.1 have all provided datable material. These are the Toaroha River and the Kokatahi River in the central Westland section of the Alpine Fault; and the northwest bank of the Haupiri River, Crane Creek and the Ahaura River in the north Westland section. A detailed description of the results of investigations at these five locations forms the basis for the paleoseismic research described in the next chapter (Chapter 3).

In the following section the active Alpine Fault trace which crosses the various fluvial surfaces at the confluence of the Toaroha, Kokatahi and Styx Rivers is further detailed in order to outline the key criteria included in a full paleoseismic analysis. This general area has provided detailed paleoseismic data of the Central Alpine Fault at two sites which are in close proximity (the Toaroha and Kokatahi River trench sites). It is also unique within the study area because of the semi-continuous length of the Late Pleistocene and Holocene traces, and the relative ease of trace definition provided by the cleared farmland.

2.4 AN EXAMPLE OF THE TRACE IN DETAIL – THE ALPINE FAULT TRACE NEAR THE JUNCTION OF THE TOAROHA, KOKATAHI AND STYX RIVERS

Figure 2.2 (map pocket) shows the active fault traces near the junction of the Toaroha, Kokatahi and Styx Rivers. Reconnaissance work on the Alpine Fault in this area was carried out by Bowen (1951a&b) and subsequently Berryman (1975). Geological mapping of this area has been carried out by Wright (1994).

The river system in this area has entrenched within a major late glacial - early post glacial aggradation surface (age approximately 14,000 – 15,000 years, refer section 2.5.1 for details of date) which is now approximately 70 m above modern river levels. A series of more than 17 degradational terraces have
been eroded at progressively lower levels associated with regional uplift along the Alpine Fault zone.

On the west side of the Toaroha River the most active trace of the Alpine Fault is represented by a clear scarp, with dextral strike-slip offset of young fluvial features, which extends with decreasing height across progressively younger fluvial surfaces. For most of its length this trace averages 070°. More detailed mapping and trenching in this area of scarp on the west side of the Toaroha River is described in Chapter 3 (Section 3.2.1 and Figures 3.1 - 3.4). The trenching indicates there are multiple fault strands within the scarp which splay and widen towards the ground surface. These are irregular but generally the fault zone is steeply dipping, with a significant number of west dipping strands.

Further east of the trenches the trace strike swings northwards to 045° and a broader more subdued trace continues towards the river with an inferred more oblique component.

There are other traces on this side of the Toaroha River (Figure 2.2). Another young trace branches from the most active trace and crosses most of the lower terraces, but then becomes indistinct in an area of track earthworks near the river. Further south, on a series of more elevated older terraces, there are several more traces which are not apparent on the younger fluvial surfaces. I infer these are related to a branch fault which appears to join the Alpine Fault in this area. There are other fault traces in the Kokatahi River immediately to the south of the Toaroha River, including an unusually straight truncation of the late glacial - early post glacial aggradation surface which may be a fault scarp, and which all appear to be associated with this branch fault.

The Alpine Fault trace between the Toaroha and Kokatahi River is intermediate in character between the strike-slip dominated faulting south of the Toaroha River and the more oblique thrust fault styles which predominate northeast of the Kokatahi River. A broad scarp in this area gently rises by 20 m over a
distance of 350 m. Individual traces can be discerned on the surface of this broad scarp, including one trace in swampy ground to the north.

GPR surveys on this scarp (Yetton and Nobes, 1998) indicate that bounding faults dip down to the northwest on the southeastern side of the fault, and dip down to the southeast on the northwestern side, forming a near-surface negative “flower” structure. Within these bounding faults are a series of at least seven subsidiary smaller faults which can be inferred from the GPR profile. This has formed through minor extensional collapse of the uplifted scarp in response to local fault zone complexity within the broadly transpressive environment.

Further northeast across the Kokatahi River the average trace strike swings to the north (average 040°) and the component of oblique thrust faulting increases. This fault style can be recognised in the field by more frequent curvature of the fault traces in plan view, and the uplift and tilting of the fault hanging wall towards the southeast on a range of scales (Figure 2.2). Trenching of the youngest fault traces in this area (Section 3.2.2, Figures 3.5–3.10) confirm the subsurface predominance of low angle southeast dipping fault strands.

The fault zone continues northeast across Dowricks Dilemma (a small creek) to the Styx River. In this area the dip of the bounding faults, and increased contractional component across the fault zone, combine to create a “pop-up” section of the hanging wall (labeled “Dowricks pop-up” in Figure 2.2). There appears to be a cross fault at Dowricks Dilemma itself (shown inferred), which is accommodating greater differential uplift at the southern end of this pop-up.

Further northeast, and approaching the Styx River, the fault strike once again swings more east-west (average 065°) and the main fault traces become distinctly more linear.
The general area described above, and included in Figure 2.2, demonstrates just how complex the Alpine Fault trace can be on a meso-scale, and also the importance in such areas of correctly identifying the relative age and activity level of the fault traces which are ultimately selected for paleoseismic investigation.

2.5 NEW ESTIMATES OF RATES OF VERTICAL AND HORIZONTAL FAULT OFFSET

In order to provide a framework for consideration of specific Late Holocene ruptures and associated earthquakes it is relevant to outline here new estimates of long term tectonic rates which have come from this thesis fieldwork. It is generally difficult in the Alpine Fault zone to identify landforms which are offset by the fault, and which might otherwise provide piercing points to reconstruct fault movement. This is because of the very high uplift and erosion rates on the southeast side of the fault, and the corresponding rapid burial and subsidence of landforms on the northwest side. However, at two locations within the field area, radiocarbon dates from paleoseismic investigations have provided new estimates for long-term rates of vertical and horizontal fault offset. These results are described below and compared with other published estimates for the Alpine Fault.

2.5.1 Late Holocene vertical offset at the Toaroha and Kokatahi River junction

On the true right of the Toaroha River the deformed surface forming the broad fault scarp described in the earlier section has been dated (Figure 2.2). The terrace surface east of the fault trace is elevated around 20m above modern river level. The terrace is underlain by fine to medium silty sandy schist gravels with a characteristic blue grey colour and strong induration. This type of gravel unit can be recognised throughout this area as forming a major late glacial - early post glacial aggradational surface approximately 70m above modern river
levels. A radiocarbon date was collected during fieldwork from near the top of this surface in the adjacent Styx River of 12,250 ± 90 yrs BP ¹ (WK 4015) and this indicates formation of this surface approximately 14,000 to 15,000 years ago.

Subsequent to this aggradational episode the river regime has reversed and a series of degradational terraces have been cut into this older aggradational gravel, most of which have a younger aggradational cap. The faulted terrace which forms this scarp belongs to this group. I have dated the younger erosional episode responsible for strath formation, from buried fibrous vegetation immediately above the strath, which in turn is overlain by 1.3 m of thin silty sands which form a terrace cap (refer Fig. 2.2 for sample location). This buried material has been dated at 2730 ± 90 yrs BP (WK 4915) which implies a calendric age before 2000 AD of 2420 to 3150 years i.e. 1150 BC – 420 BC (conversion conventions as for Wk 4015 above with two sigma calendric date limits).

These dates indicate the deeper gravels at this site have been undergoing deformation in the fault zone for at least 14,000 years, but a new younger time line has subsequently been superimposed on this early deformation by river erosion around 2800 years ago. To obtain a long-term uplift rate the vertical deformation of this Late Holocene surface must be defined.

To determine the true offset I have carried out both a level survey and a Ground Penetrating Radar (GPR) survey along the farm track (Yetton & Nobes, 1993).

¹ Throughout this thesis radiocarbon dates (as opposed to calendric dates) are given as years Before Present (yrs BP) with an associated error, indicating radiocarbon years prior to 1950. A radiocarbon year does not simply equate with a calendar year and the relationship between them has changed through time. I show the calibration graph and include a brief discussion of radiocarbon dating in Appendix 1. The error presented with the radiocarbon date is traditionally only one standard deviation but dates in this thesis have been calibrated adopting Vogel et al. 1993 and Stuiver & Reimer 1993, with two standard deviations (95% confidence limits) or greater reliability. WK designates the University of Waikato Dating Laboratory and the number following this (e.g. 4015) is the sample number.
This survey location was perpendicular to the main trace (refer Fig. 2.2). The GPR results indicate the true offset of the dated surface is the sum of the apparent vertical offset of 20m ± 0.1 and the additional 1.75 ± 0.5m of burial on the down-thrown side as measured in the GPR profile. This corresponds approximately to a total vertical offset of 21.75 ± 0.6m.

Using the full range of calendric age for the terrace (i.e. 2420 - 3150 yrs before 2000 AD), and this measured vertical offset, the uplift rate over the last approximately 3000 years at the Toaroha River is 8 ± 1.25 mm/yr. This estimate assumes the true values for the age and the offset could lie anywhere in the magnitude ranges. Using probability theory and the method of Geyh & Schliecher (1990), which recognises the slightly higher probability of the true values lying near the middle of each range, this estimate can be refined numerically to 7.8 ± 1 mm/yr.

Discussion

Although the ± 1 mm/yr error is a true approximation of the numerical uncertainties there are other geological factors which must be considered. Consider for example the difference to the estimate which is likely to apply immediately after the next Alpine Fault ground rupturing earthquake. Assuming an additional 1.5m of vertical movement in the next earthquake the apparent long-term uplift rate increases to 8.4 mm/yr.

Any significant time variation between the apparent age of the strath and the actual commencement of uplift will also affect the estimate. The age comes from the base of a 1.3 m thick sequence of overbank silts and sands deposited on the strath under flood conditions. Flood waters in large rivers such as the Toaroha, frequently reach 3 - 4 m above normal river level, and so can leave this type of deposit on relatively high surfaces. Therefore some uplift may have
already been underway when the vegetation was buried by the silt. If an arbitrary 3 m of early uplift is deducted from the derived 21.75 ± 0.6 m, then the calculated uplift rate reduces to 6.7 mm/yr.

Suggate (1968) and Simpson et al. (1994) determined uplift rates at the Paringa River (160 km southwest of Toaroha) from a localised up-thrust ridge adjacent to the fault of 11 and 13.7 mm/yr respectively. A lower figure of 7 - 8 mm/yr was obtained for regional post-glacial uplift rates which exclude the effects of local tilting.

The Toaroha uplift rate is consistent with their regional rate. The Toaroha site is not obviously located on any localised up-thrust feature. Such features do exist nearby, for example 1 km further northeast at Dowricks Dilemma (GR J33/586115). However, such areas are characterised by strongly oblique faulting, with associated back tilt of the hanging wall, which is normally obvious in the field on the up-thrown surface (refer Figure 2.2 for examples). In this case the level profile of the fault zone shows the smooth continuation of a typical fluvial gradient and no evidence of back tilt. In addition the GPR profile indicates a broadly extensional collapse structure, as opposed to a more contractional geometry (Yetton and Nobes, 1998).

Apart from the Paringa River, no other direct estimates of post-glacial uplift rates have been made along the central Alpine Fault. However, Bull & Cooper (1986) use inferred Pleistocene marine terraces to estimate uplift over a longer time range. They calculated an uplift rate of 7.8 mm/yr in the Fox - Franz Josef area, reducing to 5.5 mm/yr further northeast at Lake Kaniere.

The uplift rate here at Toaroha is a little higher, but broadly similar to their estimate. Lake Kaniere is another 10 km northeast of the Toaroha River and it
is possible the uplift reduces slightly over this distance. In addition the Bull & Cooper analysis assumes a constant uplift rate over the last 40,000 years since formation of the youngest inferred marine terrace. Minor fluctuations in uplift rate may occur when considering a much shorter time span at a single location such as this site.

This Toaroha River uplift rate does indicate that, if marine benches originally formed along the Pleistocene range front, then they would now be expected to be preserved at the general altitudes at which Bull and Cooper (1986) claim to recognise them.

The rate of uplift at Toaroha is less than estimates of the rate of uplift near the crest of the Southern Alps. Wellman (1979) estimates 17 mm/yr while Walcott (1979) predicts up to 22 mm/yr based on plate motion. The difference between these estimates and the observed offsets at the Alpine Fault has generally been attributed to distributed movement over a wide zone within the Alpine Schists east of the fault. If these Southern Alp uplift rate estimates are correct, then distributed movement in the order of 10 mm/yr is occurring away from the fault. However, Bull & Cooper (1986) report no suggestion of this pattern in the first 5 - 6 km of schist southeast of the fault in the Franz Josef area where they have examined the inferred marine terrace sequence in the most detail.

2.5.2 Late Holocene horizontal and vertical fault slip rates at the Haupiri River

The Haupiri River is 25 km northeast along fault strike from the Hope Fault junction at Inchbonnie (Figures 1.3 and 2.1). This location was first described by Munden (1951) and Wellman (1955b), who both noted a series of terraces showing progressive offset at the fault trace. They also described a 300mm high young fault trace, on an island in the middle of the river, which was in existence at that time. The site was subsequently visited in the later
reconnaissance trips of the Alpine Fault made by the New Zealand Geological Survey (Officers of the Geological Survey 1975 & 1985). Attempts to find the island and the young fault trace were unsuccessful and unfortunately it appears intervening floods have eroded this area away.

**Figure 2.3** – The Late Pleistocene and Holocene Alpine Fault trace at Haupiri River (GR K32/056437). The general location of this area shown in relation to the Alpine Fault is outlined in Figure 2.1 (map pocket). The area included in the box outline on the northeast side of the river in shown in more detail in the following Figure 2.4.
Figure 2.3 is a general map of the fault trace in this area and Figure 2.4 shows the northeast side of the river in more detail. The up-thrown side has a series of three terrace treads ranging in elevation between 3 m and 8.5 m above the current river level. The lowest tread has been trimmed by river erosion, and the highest tread is extensively modified by farm development. However, the middle tread, which is approximately 7 m above current river level, is relatively well preserved.

Four shallow channels in the tread are apparent on the up-thrown side, adjacent to a 1.5 m riser to the next higher surface on which there is a barn, sheep yards and various farm buildings. The main fault scarp crossing this middle tread surface is approximately 5 m high and strikes 060°. There are also two much lower subsidiary traces, crossing the channels at around 080°, which are both less than 0.75 m in height. Some minor dextral horizontal offset is implied by the channel pattern crossing these but this is not clear enough to quantify.

The adjacent ground on the down-thrown side of the main scarp is very swampy and densely forested. Despite local fan development associated with periodic flow in the various channels above the scarp, it is possible to discern the continuation of two of the channels and main riser on the down-thrown side. These provide piercing points for estimating the dextral horizontal offset of the 7 m terrace tread. The dextral horizontal offsets of the various features shown in Figure 2.4 is calculated by averaging the offset of the channel center lines and each channel side, and includes an estimate of the appropriate error based on the field expression of each feature.

The largest horizontal offset of 14 ± 2 m is at the main riser, which is probably the oldest feature of the terrace surface. The channels are more likely to have
Figure 2.4 – A more detailed map of an area of the Alpine Fault trace on the true right (northeast) side of the Haupiri River (GR K32/058439). Refer Figure 2.3 for location and also Figures 3.11 – 3.14 for paleoseismic investigation data.
been formed after the riser was cut, and to have remained active for longer, but
the offset of these is broadly similar at 12 ± 2 m and 11 ± 2 m respectively.

An age for this tread has been obtained by excavating a shallow 1m deep
trench in the location shown in Figure 2.4. A thin layer of topsoil (100 mm)
overlies overbank silt and sand which in turn overlie coarse Haupiri River
alluvial sandy gravel. Near the base of the flood deposited material at a depth
of 900 mm, and immediately above the gravel, are twigs and leaf fragments
within silty sand. This organic material is present for the length of the trench
and is below the base of the shallow channel (which the trench crosses). I
therefore infer that this was buried by the Haupiri River and provides the
approximate main tread age, slightly predating the channels. A sample of this
organic material (Wk 5377) yields a radiocarbon age of 2070 ± 130 yrs BP
which. The calendric age range at the 95% confidence level for this sample is
1750 to 2385 years before present (2000 AD) i.e. 385 BC to 250 AD.

To obtain the true range of horizontal slip rate the various uncertainties must
be included and calculation made over the full range of net offset and landform
ages. This indicates the average horizontal slip rate at the main fault trace over
the last approximately 2000 years is 6.5 ± 2.5 mm/year.

Discussion – horizontal rates

Once again the numerical errors are not necessarily the true errors. As noted
earlier in the discussion of the vertical estimates at the Toaroha River, the
overbank flood material on which the age is based may have been deposited
on the underlying gravels once some fault movement had already occurred.
This may lead to a slight over-estimate of the true rates.

A more important source of error comes the short time period which is
represented (approximate 2000 years). This potential error arises from possible
short-term temporal variations in the rate of fault activity, and also uncertainty
regards the current seismic cycle. For instance a rupture of 3 m in the next 12 months would increase the apparent average horizontal rate by around 1.5 mm/yr to approximately 8 mm/yr.

No other estimates of horizontal or vertical offset rates are available for the north section of the Alpine Fault. The closest estimate comes from 25 km southwest along strike at Inchbonnie at the junction with the Hope Fault. At this location Berryman et al. 1992 estimate rates of 10 ± 2 mm/yr horizontal offset and 6 ± 1 mm/yr vertical offset for the last 1300 years. However, as well as spanning a short time period, these are based on a relatively small number of greywacke weathering rind ages (Berryman, pers. comm. 1995). Although this dating method is widely adopted in the eastern South Island (see for example Chinn 1981; McSaveney 1992) it has yet to be proven as reliable in the more humid and warmer climate of Westland.

Rates of movement for the north Alpine Fault in this area can also be assessed from recent estimates of slip on the various Marlborough faults which join the Alpine Fault further to the north. It is likely that at least some of the estimated 4.7 mm/yr of slip on the Clarence Fault (Van Dissen & Nichol, 1998) joins the Alpine Fault southeast of Haupiri River, and therefore should be excluded. However, the Awatere Fault joins the Alpine Fault approximately 40 km northeast of Haupiri River. The Awatere Fault has an estimated slip rate of 6 ± 1 mm/yr over the last 20,000 years (Benson et al. 1998; Little 1998). In addition to slip transferred from the Awatere Fault there is also the component of slip along the Wairau section of the Alpine Fault, estimated at 4 ± 1 mm/yr by Berryman et al. 1992. This sums to 10 ± 2 mm/yr and suggests the Haupiri River rate obtained here for the last 2000 years, although of the right order, may still be an under-estimate.

By adding this Haupiri River rate with estimates of the average slip rate on the Hope Fault zone of 25 ± 10 mm/yr (Cowan 1989; Cowan 1990; Van Dissen and Yeats 1991; Pettinga et al. 1998) an estimate for the central Alpine Fault of
31.5 ± 12.5 mm/yr is obtained. As was noted in Chapter 1, although this estimate has a very large uncertainty, it does suggest that the true Alpine Fault rate is most likely to be at the high end of the 25 ± 5 mm/yr range of Norris & Cooper (1998), and is most likely in the order of 30 mm/yr.

Vertical uplift rates at the Haupiri River

The uplifted terrace tread at Haupiri River, from which the horizontal rate discussed above has been determined, forms a scarp which averages 5 ± 0.75 m in height. The channels on this tread, which can still be discerned on the down-thrown side, indicate that there has been little significant burial, and so at this site the scarp height approximates to the total vertical offset.

This scarp height, in conjunction with the tread age, equates to a vertical offset of 2.5 ± 0.8 mm/yr for the past approximately 2000 year period. This is a significant reduction in comparison to the uplift rate at the Toaroha River (8 ± 1.25 mm/yr). This substantial difference in the fault uplift rates is not reflected in the average elevation of the Southern Alps between the Toaroha River and the Haupiri River, which remains broadly similar (refer Figure 2.1). This suggests that progressively increased uplift on other faults, further northwest of the Alpine Fault, may be accommodating some of the difference between uplift rates at the Alpine Fault in the central and more northern area.

2.6 SUMMARY AND CONCLUSIONS

- The Alpine Fault forms a discontinuous Holocene and Late Holocene trace along the northwestern margin of the Southern Alps throughout the field area.

- The average strike of the trace between the Hokitika River in the south, and Lynch Creek in the north (near the Taramakau River), is approximately 050°. In some parts of this area the trace segments into oblique thrust
sections striking 020 – 050°, that are linked by dextral strike-slip traces striking 065 – 090°, and this is a similar pattern to that observed by others further to the south.

- Between Lynch Creek and the extension of the fault northeast to the Ahaura River (the northern limit of field work) the average fault strike changes to 060°. In addition local segmentation ceases and right stepovers become more common.

- In some locations within the field area large rivers which cross the Alpine Fault have created flights of terraces allowing the definition of Late Holocene fault traces. In particular the lowest terraces, closest to modern river levels, provide the best potential sites to date the most recent few ruptures of the fault.

- Eight locations were initially identified as having paleoseismic potential. Five of these locations have provided datable material for paleoseismic analysis in a total of seven excavations. The details of this are presented in the following chapter.

- Some of the radiocarbon dates have also provided new estimates of long-term vertical and horizontal fault offset to provides a framework for consideration of specific late Holocene ruptures. An uplift rate for the past approximately 3000 years of 8 ± 1.25 mm/yr has been obtained from the Alpine Fault scarp at Toaroha River. This is broadly similar to earlier estimates for the central Alpine Fault.

- At the Haupiri River, in the less active north Westland section of the fault, the uplift rate reduces to 2.5 ± 0.8 mm/yr over the past 2000 years. A horizontal slip rate of 6.5 ± 2.5 mm/yr can also be obtained over the same time span.
Chapter 3

PALEOSEISMIC EVIDENCE FROM FAULT TRENCHING

3.1 INTRODUCTION

This chapter presents the results of paleoseismic trenching and excavation at a series of sites located along approximately 75 km of the Alpine Fault trace. These sites were selected following the fault trace mapping previously outlined in Chapter 2. The principal criteria for selection of the sites is evidence of relatively fresh ground rupture in a narrow and well defined zone, preferably with evidence of subsidence and the associated accumulation of datable organic material on the down-thrown side. In addition the sites were chosen to provide a geographic range of location and to bracket the junction of the Hope and Alpine Faults which has been previously proposed as a possible fault segment boundary (Berryman et al., 1992).

Despite looking along 75 km of fault strike the number of potential sites are relatively few, reflecting the very rapid burial of the youngest trace by sedimentation and mass movement in this active range front environment. Much of this burial probably occurred relatively quickly after the most recent earthquake. In general the best sites are the relatively wide alluvial fans associated with the largest rivers. In such locations there is a combination of faulted young sediments forming distinct geomorphic surfaces at a reasonable distance from unstable slopes and active alluvial fans. The main limitation to scarp preservation in these areas is subsequent flood modification and river erosion.
The paleoseismic evidence presented in this chapter is direct evidence of previous fault ruptures. It is based on the radiocarbon dating of subsurface organic material in key stratigraphic positions in trenches dug by machine and smaller hand dug excavations.

This evidence is presented in chronological order, with the most recent rupture event presented first. The rupture events have been named after the locations at which the available paleoseismic evidence best constrains the rupture event timing.

### 3.2 PALEOSEISMIC TRENCH EVIDENCE FOR THE MOST RECENT EARTHQUAKE RUPTURE c. 1700 – 1750 AD – THE TOAROHA RIVER EVENT

#### 3.2.1 Toaroha River Site

The Toaroha River crosses the Alpine Fault very close to the Kokatahi and Styx Rivers near the south end of Lake Kaniere (see Figure 1.3). This area was first inspected at a reconnaissance level by Wellman (1955a) and later by the New Zealand Geological Survey (Officers of the Geological Survey, 1975; 1985). Wright (1994) included the area in his more detailed mapping work, which focussed primarily along the range front with only limited mapping of the young alluvial surfaces.

As discussed in Chapter 2 and shown in Figure 2.2 (map pocket), the Alpine Fault in this area is characterised by a complex pattern of segmentation, with multiple surface rupture traces and a wide zone of ground warping. This is associated with the change from a predominantly strike slip segment on the southwest side of the Toaroha River, to a more north striking oblique thrust near the Kokatahi River. Figure 3.1 shows a simplified map of the fault zone in the vicinity of the Toaroha and Kokatahi Rivers, with the Toaroha trench sites
shown near the west margin. The nearby Kokatahi River trench sites, which are only 1 km further to the northwest, are also shown.

Figure 3.1 - The Alpine Fault in the vicinity of the Toaroha and Kokatahi Rivers inland from Hokitika (GR J33/572102). Refer Figure 1.3 and 2.1 for the area location. Section 2.4 describes segmentation and possible branch faulting in this area and Figure 2.2 (map pocket) is a detailed map including this area.

In Chapter 2 the evidence for the existence of a branch of the Hope Fault joining the Alpine Fault zone in this Toaroha River area is noted. It is
Figure 3.2 - Trench location details for the two trenches on the true left of the Toaroha River (GR J33 572102). The trenches were located on the third terrace tread (Surface C), which the NW fault trace has ruptured at least twice. The trenches are near an area of swampy ground on the down-faulted NW side, which is denoted by the swamp-grass symbol.
possible the southern most fault trace on the true left of the Toaroha River is related to movements on this inferred branch fault, rather than the Alpine Fault, so this trace was avoided as a trench site. Instead the trenches were located on the most westerly trace, which has also been the most active in displacing the alluvial channels and risers. Figure 3.2 details the location of the two trenches at the Toaroha River site.

The two trenches were located close together on a terrace approximately 6 metres above current average river level. This terrace surface (Surface C) is the third tread above the current river flood plain (i.e. Surface A is the first tread above the modern river flood plain). The fault trace across Surface C has a subdued surface expression with approximately 1m of vertical relief. A small swampy area occupies the down-thrown side west of the cattle race and two trench sites were selected in this area in the hope of obtaining datable organic material in this swampy ground.

The surface evidence of horizontal displacement of channels on this trace (Fig. 3.2) indicates a range from 6m on the most recently modified old channels up to 16–17m, and possibly as high as 22m, in locations where the channels have been completely abandoned. Based on typical horizontal displacements in large earthquakes on other strike slip faults, and inferred single event displacements at other locations on the Alpine Fault, this suggests that at least two ruptures, and probably three, have created this trace.

The riser between Surface C and Surface B, which was created in conjunction with Surface B, shows 7m of dextral horizontal displacement. This could be a single event offset in the most recent event, but stream modification at the base of the riser may have reduced the combined offset of two events.

If only two ruptures have created the trace in Surface C, and the most recent event was of around 7m (or slightly more), then the penultimate event must have had a larger displacement (i.e. in the order of 9 – 10m which is still
conceivable). However if the 22m offset is considered then the offset increases to 15m which is almost certainly two events. In addition the subsurface evidence in the Toaroha 2 trench supports the surface evidence that three events formed the trace on Surface C.

*Toaroha 1 Trench*

The trench log for the Toaroha 1 trench is presented with the key and summary box as Figure 3.3. The Toaroha 1 trench shows that the surface expression of the fault trace coincides with subsurface evidence of faulting and an associated stratigraphic contrast across the fault trace. On the relatively up-thrown southeastern side there are crudely imbricated coarse fluvial schist derived gravels, typical of the modern river bed load, which extend virtually to the surface (unit 1). On the down-thrown northwestern side unit 1 is at a greater depth and overlain by an interlayered sequence of finer grained sands, silts and sandy gravels, some of which have organic material associated with them (units 2 - 7). Note that in the various trench logs organic soil horizons are designated by the suffix “A”, attached to the parent unit on which they have developed (e.g. unit 6/7A).

Fault planes are clearly apparent. There is evidence of displacement and truncation of the finer grained units at the fault planes. This is created by the different sedimentary environment on the down-thrown side, and the likely cycle of local and catchment wide deposition immediately following each fault rupture event.

There are multiple fault strands which cause the fault zone to splay and widen towards the surface. These strands are irregular but generally the fault zone is steeply dipping, with a significant number of west dipping strands. The surface morphology mapped directly west of the Toaroha 1 trench (Fig. 3.2) suggests there is some small scale slip partitioning. The most western fault plane (F1, note that for convenience all faults and fault zones have been numbered from
Toaroha 1 Alpine Fault Trench

Stratigraphic key for Toaroha 1 trench

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Interpretation</th>
<th>Event Horizons</th>
</tr>
</thead>
<tbody>
<tr>
<td>6A/7A</td>
<td>Soft brown organic clayey silt.</td>
<td>Most recent topsoil</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Grey, silty sandy, fine - medium gravel with no fluvial structure</td>
<td>Regrading from forest clearing (and scarp derived colluvium?)</td>
<td>Earthquake event?</td>
</tr>
<tr>
<td>5</td>
<td>Grey silty sand and sandy silt with fine-med. gravel and no fluvial structure</td>
<td>Scarp derived colluvium</td>
<td></td>
</tr>
<tr>
<td>4A</td>
<td>Grey slightly gravelly sandy silt with no fluvial structure</td>
<td>Finer scarp derived colluvium.</td>
<td>Earthquake event</td>
</tr>
<tr>
<td>4</td>
<td>Soft black brown charcoal and sandy peat</td>
<td>Buried soil and local swamp accumulations</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Grey blue gleyed slightly sandy silt with some clay. Charcoal, wood and fine gravel locally abundant. Minor stratification.</td>
<td>Overbank fluvial silts, aged around 210 ± 50yr BP from sapwood of transported branch</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Grey sandy medium to coarse, subangular to rounded, schist gravels</td>
<td>Fluvial gravels</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Grey coarse sand and fine schist gravel with local lenses of silty sand. Stratification common</td>
<td>Flood channel fluvial sands and fine gravels. Age &gt; 250 ±50yr BP date on an old root</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Grey sandy medium to coarse, subangular to rounded, schist gravels with clast alignment and imbrication shown diagrammatically</td>
<td>Fluvial gravels of surface C showing various degrees of disruption by faulting</td>
<td></td>
</tr>
</tbody>
</table>

Summary

- Predominantly steeply dipping strike slip faults (068° strike)
- Earthquake associated rupture (not aseismic deformation)
- At least two ruptures confirmed, but earlier ruptures possible
- Last event apparently post 1700 AD

Figure 3.3 - Log, key and summary of the main features of the Toaroha 1 trench
west to east regardless of their apparent age) appears to be accommodating the oblique thrust component while the fault strands on the relatively eastern side form a dominantly strike slip fault zone. The surface trace clearly bifurcates directly northeast of the trench into these two east-west and more northerly striking component traces. This may indicate that on a detailed scale a half flower structure is developing locally in this area (Sylvester, 1988).

The subsurface evidence reveals two definite faulting events that can be recognised in this trench although more are possible. F2 displaces units 1-2 but is in turn cut by F1 (a strand of FZ3). Other strands of FZ 3 extend virtually to the surface and displace units 1-7.

Pantosti et al. 1993 define the term “event horizon” in relation to paleo-earthquakes. An event horizon is stratigraphically defined by either scarp-derived colluvium that buries the pre-faulting surface, and/or by unconformities that develop as a result of warping and subsequent deposition. The unconformities typically develop on the downthrown block in fluvial and lacustrine sediments (McCalpin, 1996).

In the Toaroha 1 trench the clearest event horizon is between units 4a and 5. Here a soil horizon (4A) is buried by scarp derived colluvium (unit 5). Unfortunately this buried soil 4A does not contain reliable datable material. Although charcoal is present in the soil charcoal can have considerable self-age and the fine peat in a buried soil is rendered unreliable by the mixing processes of worm action and eluviation (Tonkin, pers. comm. 1997). Thus this paleo-earthquake can not be dated from this trench.

The subsequent warping of unit 5, the faulting of scarp colluvium (unit 6), and the apparent burial of the soil from which sample 5510 was taken, all define a more recent event horizon. Unit 6 contains buried wood from the outside of a branch (probably kahikitea, Dacrycarpus dacrydiodes). This returned a "modern" date which indicates a calendric age anywhere between c.1700 AD
and 1950 AD (but not post 1950 AD). The faulting of this unit implies an earthquake event post 1700 AD.

However, based on this trench alone there is still doubt about this conclusion. The difficulty arises from the date returned from the apparently buried part of unit 6A. A twig sample from in this unit (Wk 5510) returned a date of post 1950 (i.e. post the first influence of atomic bomb testing on wood growing in the New Zealand area). Unfortunately, like many of the river flats in Westland this area was first logged around 1880, and then fully cleared later in 1940 and 1950 with grazing and some minor cropping continuing to the present day. At least some disturbance of the original ground surface is inevitable in these clearing and farming operations and the date from this shallow sample shows how much mixing can occur. It is possible the date from unit 6 could have a similar origin.

The final date from this trench is provided by Wk 5511 from a root in apparent growth position in the coarse sand unit. Finer roots can be traced from this large root which are all confined to the sand unit and do not extend down from the gravel or overbank silts above it. I infer these roots belong to vegetation killed by the subsequent overlying gravel inundation but the possibility of root intrusion from the surface cannot be entirely discounted. This sample returned a date of 250 ± 50 yrs BP which indicates the faulted sand unit is at least older than this (it could be considerably older) but unfortunately this does not constrain any particular event timing.

The Toaroha 1 trench is therefore inconclusive. There is a suggestion of a faulting event between 1700 and 1950 AD (WK 5512) but contamination associated with logging cannot be ruled out as the alternative explanation for this date.
Toaroha 2 Trench

Fortunately the Toaroha 2 trench, shown in Figure 3.4, is more definitive. It is located only 15m from the first trench, and similar units appear to be represented. However these units are not necessarily lateral stratigraphic equivalents and trenching the intervening 15m would be required to demonstrate this.

Once again the surface scarp corresponds with clear subsurface fault planes which truncate, and in part offset, finer grained fluvial units which are only present on the northwest down-thrown side. The fault strands are relatively steep and a significant number dip to the northwest. Clearly at least one fault rupture has occurred that extends virtually to the ground surface (F3/F4). An earlier faulting episode can be inferred from F1 that offsets units 1 – 3 but is abruptly truncated by erosion and deposition of unit 4. Others earlier events may predate unit 2 and have been responsible for creating the local depositional basin in which this sand accumulated. However the subsurface record does not conclusively show this.

The most recent event horizon occurs between unit 4 and unit 5. Unit 4 is a grey blue overbank silt with minor buried wood and has been clearly deformed and faulted. Unit 5 is scarp derived colluvium that has apparently infilled a synclinal depression between F1 and F2. A soil horizon has formed over this (5A) and in part over unit 4 (4A). The dates available for unit 4A/5A and the date for the deposition of unit 4 effectively bracket the age of the earthquake responsible. The youngest dates on wood fragments in the soils (WK 5514 and WK 5515) both indicate a post 1700 AD age.

The most definitive paleoseismic date is provided by Wk 5513 from the overbank silt (unit 4). This has been obtained by dating the outer 20 rings of a small, slightly abraded, fragment of silver pine branch (*Lagarostrobos colensoi*) which was found completely encased in undisturbed overbank silt.
Summary

- Predominantly steeply dipping strike slip (068° strike)
- Earthquake associated rupture (not aseismic deformation)
- At least two ruptures confirmed, and earlier ruptures possible
- Last event post 210 ± 50 yr BP

Figure 3.4 - Log, key and summary of the main features of the Toaroha 2 trench
Silt has been deposited in the broken ends of the branch and there are signs of minor fluvial abrasion that has removed the bark and at least some sapwood. The conventional radiocarbon age for the outer twenty rings of wood is $210 \pm 50$ yrs BP and this represents the oldest possible radiocarbon age for the faulted overbank silt in which it is found. The calendric age derived from this radiocarbon age is some time post 1660 AD. Prior to burial the branch had to break and fragment before entering the river system and an unknown period of time has then elapsed in the process of deposition and the period before fault rupture. Thus it is possible the faulting post-dates the maximum unit age of 1660 AD by some tens of years.

The Toaroha 2 trench provides the first unequivocal evidence of a rupture of the central Alpine Fault some time after 1660 AD. In this thesis the earthquake responsible for this rupture is referred to as the Toaroha River event. The adoption of a name for this event, as opposed to a date, recognises the likelihood of further improvements in the estimates of event timing. This name does not imply an epicentre at or near the Toaroha River, or suggest the event was restricted to only this area.

Possible indirect evidence for an earthquake post 1660 AD is provided by a dated rockfall within 300m of this trench. There is obvious debris near the Alpine Fault scarp in this area (GR J33/564100) derived from a moderately large granite mylonite rock fall which has come from the eastern slope of Mount Harry. The scar remaining on the hill side is still obvious in the forest regeneration pattern. An excavation was made in this rock fall debris and a buried punga trunk was located at the base of the deposit. The sapwood from this yields a radiocarbon date of $200 \pm 50$ yrs BP (Wk 4919, refer Table 4.1). Mount Harry is not particularly high or steep, having been glaciated in the late Pleistocene, and rainfall triggered rockfalls in the relatively hard and resistant granite mylonites of this area are not common. It is possible this rockfall was triggered by the Toaroha River event, however there is no way to confirm this.
3.2.2 Kokatahi River Site

On the true right bank of the Kokatahi river there are Holocene fault traces with a more north-south strike than the Toaroha River traces, and which frequently exhibit a wide zone of deformation with warping and back tilt to the east on the relatively upthrown side. The wider zone of ground deformation, including the tilting, warping and folding, is consistent with the hanging wall deformation of a low-angle oblique thrust fault.

The earlier Figure 3.1 shows the area in relation to the fault zone structure and the Toaroha trenches. Figure 3.5 is a detailed map of the trench sites showing the multiple traces and associated folding. Two adjacent scarps were selected for trenching.

*Kokatahi 1 Trench*

The southeastern trace is around 1m high. An abandoned channel crosses the scarp 50m southeast of the trench, and there is no surface trace of the fault across this, suggesting flow and sedimentation in this channel since the last rupture. After crossing the fault this channel swings north along the base of the scarp. A trench (Kokatahi 1, designated K1 in Fig.3.5) was excavated perpendicular to the scarp with a drain below the scarp base at right angles to this within the young channel, extending along the scarp base to the creek approximately 50m away. The trench log is presented in Figure 3.6 and also includes the logged section of the drain approximately 30m away from the main trench.
includes the logged section of the drain approximately 30m away from the main trench.

Figure 3.5 - The Alpine fault and the location of the two trenches on the true right of the Kokatahi River (GR J33/580111). Refer also Figure 3.1 for proximity to the Toaroha trenches and Figures 1.3, 2.1 and 2.2 for location information.
Kokatahi 1 Alpine Fault Trench

Stratigraphic key for Kokatahi 1 trench

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Interpretation</th>
<th>Event Horizons</th>
</tr>
</thead>
<tbody>
<tr>
<td>4-6A</td>
<td>Soft brown organic clayey silt</td>
<td>Topsill</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Grey uniform coarse sand burying stump (Wk 5469)</td>
<td>Young flood channel sand fill post 1700 AD</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Grey blue clayey silt with angular coarse sand and fine gravel clasts. No stratification. Clasts aligned parallel to slope.</td>
<td>Scarp derived colluvium.</td>
<td>Earthquake event</td>
</tr>
<tr>
<td>4</td>
<td>Grey blue gleyed slightly sandy silt with minor stratification.</td>
<td>Overbank fluvial silts.</td>
<td>Earthquake event</td>
</tr>
<tr>
<td>3</td>
<td>Grey brown silty fine sand. Minor stratification.</td>
<td>Overbank fluvial silty sands.</td>
<td></td>
</tr>
<tr>
<td>2a</td>
<td>Grey uniform coarse sand capped by a discontinuous lens of iron stained pebbly sand (2a). Transported and abraded rata branches near base (Wk 5470) and some broken roots close to fault zone (Wk 5471)</td>
<td>Flood channel fill from around 490 ± 40 yr BP. Hiatus following deposition marked by iron staining and growth of stump (Wk 5469)</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Grey and brown medium to coarse subangular to rounded sandy schist and greywacke gravel.</td>
<td>Main Kokatahi River bed load.</td>
<td></td>
</tr>
</tbody>
</table>

Summary

- Predominantly oblique thrust faulting (045° strike)
- Two events, post 490 ± 40 yr deposition of the sand unit

Figure 3.6 - Log, key and summary of the main features of the Kokatahi 1 trench
Figure 3.7 - The excavator at work beginning the Kokatahi 1 trench. The bucket is just starting to reach the actual fault rupture and blue grey coarse sand channel fill is apparent immediately downhill of the bucket.

Figure 3.8 - Photograph and the relevant section of the face log showing evidence for two rupture events since deposition of the sand containing transported wood from post 1470 AD.
Figures 3.7 and 3.8 are photos taken during the trenching. The first shows the scarp and the digger beginning the excavation while the second is a detail of the trench log showing two successive east dipping shears discussed further below.

The surface trace corresponds with subsurface evidence of faulting. It is very clear in the trench relationships that this is a low-angle east dipping fault in marked contrast to the sub-vertical and west dipping faults in Toaroha trenches which are located on more east-west aligned strike-slip segments of the trace. Once again finer grained units dominate on the relatively down-thrown side but in this case some tentative correlation of units is possible across the fault zone.

The oldest unit on both sides (unit 1) is crudely imbricated coarse gravel that is typical of the modern Kokatahi River bed load. This is a larger river than the Toaroha River and has a catchment extending further southeast. As a result the gravel contains proportionally more greywacke clasts, although schist is still common, and all the clasts are more rounded.

Following the gravel deposition, the sand of unit 2 has been deposited on both sides of the fault zone. More of this sand has accumulated on the relatively down-thrown northwest side suggesting the scarp may have acted as a channel edge following the earliest fault rupture.

There has then been deposition of overbank silts on both sides of the fault (unit 3). This silt has subsequently been ruptured on the down-thrown side in the western most of the fault strands (F1). It is in turn overlain by a scarp derived colluvial unit, a silty sand containing fine schist gravel clasts with a crude alignment parallel to the scarp slope. This unit has been clearly faulted at the southeast end by F1 and this subsequent rupture of a scarp derived colluvium implies at least two episodes of faulting in this trench (refer trench face log detail and photograph in Fig. 3.8). The northwest continuation of this colluvial unit has been eroded away at the northwest end, apparently in conjunction with
deposition the most recent sand (unit 6). The geomorphic surface that is created by this youngest sand unit crosses the fault trace 50m southwest of the trench and there is no sign of a surface scarp. This relationship indicates that this youngest sand has been deposited after the most recent fault rupture.

Thus event horizons can be recognised between units 4 and 5 and between unit 5 and 6. Unfortunately the radiocarbon dates available to bracket these events are limited. Wk 5470 was collected in the oldest sand unit (unit 2) from a section of a large rata branch (*Metrosideros umbellata*) from which the sapwood has been abraded away in virtually all areas and it is definitely fluvially transported.

To get sufficient sample volume to allow an age determination wood was collected from nearer the branch heart, and between 50 and 60 rings from the youngest remaining ring. This older wood yielded a radiocarbon age of 490 ± 40 yrs BP, and when the adjustment for sample age is made, this represents a maximum calendric age for the basal sand unit of post 1470 AD. This maximum age does not allow for the lost abraded sapwood, which would normally have a width of at least 20 –30 years in a slow growing species such as rata. Allowing for this the most likely age of the deepest sand unit is therefore around 1500 AD.

A disconformity in the sand unit on the west side of the trench is marked by a discontinuous thin pebbly horizon (unit 2a) approximately 400 mm below the surface. Iron staining is abundant along this horizon and in the drain from the trench there is a buried stump of an unidentified softwood species in growth position. A sample from the sapwood was collected (Wk 5469) which yields a post 1700 AD and pre 1950 AD date. As is noted above, this sand post-dates the most recent faulting event and the inundation of the stump reflects the late reoccupation of the channel and removal of the surface trace where the channel crosses the fault.
Near the fault itself, and very close to the transported rata branch, several small kahikitea (*Dacrycarpus dacrydiodes*) roots were found in apparent growth position that have also returned a post 1700 AD and pre 1950 AD date (Wk 5471). These appear to have had their upward continuation torn out, possibly by shearing but also conceivably by flood scour. However, the roots do not extend all the way to the fault plane, appearing to stop at the silt unit, and inferences regards fault rupture based on roots are tenuous.

The most important feature of the Kokatahi 1 trench is the evidence for two rupture events in the pattern of shears and associated colluvial material in relation to unit 2, the oldest sand unit (refer photo and face log detail in Figure 3.8). Given that the maximum age of the basal sand unit is post 1470 AD (and probably post 1500 AD) *this indicates two separate episodes of shearing since 1470 AD*. This is consistent with an initial rupture sometime between 1470 and 1660 AD and renewed movement in the subsequent Toaroha River event.

*Kokatahi 2 Trench*

The Kokatahi 2 trench (K 2 in Fig. 3.5) was excavated 150m northwest of the first trench and across the highest scarp in the area, which is approximately 2.5m high. The trench face log is presented in Figure 3.9.

Coarse gravels form the underlying units on both sides of the fault zone (unit 2) but on the southeast side of the fault these gravels and adjacent unit 4 contain intrusions of sand (unit 1) that are inferred to be the result of the liquefaction of deeper sand horizons (below trench level) which predate all the other exposed units. On the southeast side of F2, unit 2 extends virtually to the surface with a thin cover of colluvium and topsoil (units 7 and 8A).
**Kokatahi 2, Alpine Fault Trench**

**Stratigraphic key for Kokatahi 2 trench**

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Interpretation</th>
<th>Event Horizons</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>Soft brown peat and silty peat with abundant wood including twigs and young branches.</td>
<td>Local swamp accumulation in fault depression around 220 ± 40 yr BP predating the last rupture and incorporated in &quot;breccia&quot;</td>
<td>Earthquake event</td>
</tr>
<tr>
<td>5</td>
<td>Grey blue gleyed slightly sandy silt with minor stratification.</td>
<td>Overbank fluvial silt.</td>
<td>Earthquake event</td>
</tr>
<tr>
<td>4</td>
<td>Grey brown silty sand with minor stratification.</td>
<td>Overbank fluvial silty sand.</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Grey uniform coarse schist and greywacke sand. No stratification.</td>
<td>Flood channel fill.</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Grey and brown medium to coarse subangular to rounded sandy schist and greywacke gravel.</td>
<td>Main Kokatahi River bed load.</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Grey brown medium - coarse sand in Unit 2 gravels</td>
<td>Liquefaction injection feature</td>
<td></td>
</tr>
</tbody>
</table>

**Summary**

- Predominantly oblique thrust faulting (045° strike)
- Earthquake associated rupture and liquefaction (not aseismic deformation)
- At least two ruptures represented
- Last event post 220 ± 40 yr BP

**Figure 3.9** - Log, key and summary of the main features of the Kokatahi 2 trench
Northwest of F2 the underlying gravel unit (unit 2) is overlain by a sequence of coarse sands, silty sands, and silts (units 3–5) and in the area adjacent to FZ1 there are younger organic units (unit 6), scarp derived colluvium (unit 7) and a fault “soil breccia” (unit 8).

Once again the fault planes within the scarp dip to the east. More than one event can be recognised from the detailed face log. The evidence for more than one event is provided by the subsequent faulting of the scarp derived colluvium and the creation of a local swampy area (furrow) at the scarp base, in which peat has accumulated, and then subsequently been faulted.

The apparent relationship between the coarse and silty sands is complex and likely to be influenced at least locally by liquefaction. Figure 3.10 is a montage of photographs from this trench. The peaty material and wood in unit 6 are sheared and “intruded” with a loose assemblage of angular pieces of soil from all the other fine grained soil units contained within a silty sand matrix (unit 8). The pieces of exotic soil within the silty sand are angular and broken, resembling a “soil breccia”. This soil unit appears to have formed by liquefaction in the last earthquake.

The close-up photograph of the fault plane (Detail B) shows similar evidence of liquefaction, with pieces of blue grey soil within coarse sand, and the sand of unit 1 shows evidence of flow along the shear. This evidence of liquefaction confirms that the observed rupture of the soil units in these trenches is associated with seismic shaking. In principal, offset soil units alone do not categorically indicate earthquake rupture. Arguably this offset could also result from slow progressive aseismic creep, despite this never having been observed along the Alpine Fault, and it being a rare phenomena internationally. However, this trench provides definitive evidence of paleo-liquefaction associated with surface ground rupture, and confirms that the Alpine Fault is a source of moderate to large earthquakes.
Figure 3.10 - Logging the first section of the Kokatahi 2 trench. The beginning of the main faulting commences at the large boulder immediately uphill of the orange string level line. Disrupted pieces of grey blue silt are visible in the peaty soils 300mm above this boulder. Details A and B are from the subsequent extension of the trench further into the scarp.

**Detail A (left):** The main oblique thrust of sandy gravels sheared over coarse sand which contains fragments of other soils and laminar "flow" bands (both visible in Detail B).

**Detail B (right):** Close up showing faint laminar bands parallel to the shear and fragments of blue grey gleyed silts, some of which at the top and bottom have been emphasised for the photo by scratching around their outlines. Both features suggest liquefaction of the sand during shear.
Although only one radiocarbon date could be obtained in this trench the date indicates a relatively recent rupture and is consistent with the Toaroha 2 trench. Sample Wk 5473 was obtained from a piece of partly rotted rata branch (*Metrosideros umbellata*) with only the heartwood preserved within unit 6, the peaty soil which appears to have accumulated in a paleo-depression at the bottom of the scarp.

The radiocarbon date for the sample was $220 \pm 40$ yrs BP but this came from heartwood with at least an estimated 15 rings missing (and possibly more). This estimate is based on the average ring density of the heartwood and the typical minimum thickness of the sap wood expected on a branch of this diameter. The corresponding oldest possible calendric age for this wood is 1665 AD. This once again indicates a young episode of Alpine Fault rupture post 1665 AD.

### 3.2.3 Haupiri River Site

The Haupiri River is 25 kilometres northeast along fault strike from the Hope Fault junction at Inchbonnie (Figure 1.3 and Figure 2.1 [map pocket]). This location has been described earlier in Chapter 2 (section 2.5.2) and a horizontal fault parallel slip rate of $6.5 \pm 2.5$ mm/yr has been derived from faulted terrace risers on the northeast bank on the river. This section outlines the paleoseismic investigations in the area.

Figure 3.11 shows the general fault trace details at the Haupiri River. Initial investigations provided several young radiocarbon dates in the Haupiri River area near the fault. An aggradation terrace approximately 3 m above modern average river level yields dates of $260 \pm 50$ yrs BP (Wk 4874, Table 4.1) for
sapwood on a red beech log (*Nothofagus fusca*), and 270 ± 50 yrs BP (Wk 4873) for some rata heartwood (*Metrosideros umbellata*).

An old debris flow is also apparent in an unnamed small stream near the Haupiri River which drains a section of the range front near the fault zone (GR K32/060435, referred to here informally as Moss Creek, Table 4.1). This debris flow had buried a once large area of trees in growth position. The trees have subsequently died and the projecting trunks fallen so that the currently grazed paddock has several short trunk sections broken at grass level. Moss creek has cut back through the debris and exhumed several of the stumps re-exposing the root plates and the paleosol.

![Figure 3.11 - The Alpine Fault at the Haupiri River (GR K32/056437). Initial trenches on the true left (southwest) bank did not provide any material for dating to constrain rupture timing. A hand-dug pit at the location shown on the true right bank has been more successful.](image-url)
Sapwood was collected from a totara stump exposed in Moss Creek (Podocarpus cunninghamii) and submitted for dating. Wk 4876 returned a radiocarbon age of 210 ± 50 yrs BP which matches the date from the Toaroha 2 trench and is very similar to the Kokatahi 2 trench and the Mt Harry rockfall dates (220 ± 50 yrs BP and 200 ± 50 yrs BP respectively).

This indirect evidence suggested that this catchment may also have been subjected to a young earthquake event matching the Toaroha River event, with associated landslides, debris flows and river aggradation. For this reason two separate attempts were made at trenching the fault at Haupiri. The first was a large machine dug trench on the true left bank where the trace forms a low scarp across cleared paddocks. Despite careful logging this trench only yielded wood from a young channel fill deposit on the down-thrown side which was not present at the shear itself. The relationship between the channel fill and the age of last rupture was therefore ambiguous and the wood was not submitted for dating. There was also abundant evidence of surface disturbance of the upper soil horizons from timber clearing, river protection works, and farm improvement.

A second site on the true right bank was selected in an area of remaining bush cover (Figure 3.11). Because of potential damage to this area of remnant forest, and the small scale of the selected feature, the decision was made to carry out the excavation by hand.

Figure 3.12 is a more detailed map of the trench area. The scarp at this location is approximately 5m high and has developed over a period of approximately 2000 years (refer Chapter 2, section 2.5.2). Figure 3.13 shows the detail of a displaced landslide which has been dextrally offset approximately 6 m by the fault thereby creating a small furrow at the base of the scarp. The geomorphology indicates that the most recent fault shear must be in this furrow and so this area was selected for the excavation.
Figure 3.12 - A more detailed map of the Alpine Fault trace in the boxed area shown in Figure 2.11 (GR K32/056437), located on the true right bank of the Haupiri River.
Figure 3.13 – Site plan of the pit area and the displaced landslide with the associated fault furrow at the base of the scarp shown in Figure 3.12. A small hand dug pit has been excavated at the location indicated.
**Figure 3.14** - Log of the hand dug pit excavated across the shear displacing the landslide shown in Figure 3.13. A date of 210 ± 50 yrs BP was obtained from twigs and leaves within peat incorporated into the shear zone.
Figure 3.14 is the log of the small pit. The top 150mm of material (unit 7) is a saturated soft brown peat containing twigs, leaves, and occasional larger branches. This changes downwards to a loose horizontally imbricated fine to medium schist and greywacke derived sandy gravel (unit 6). Minor twigs and wood are scattered through out unit 6 and twig material was collected for possible dating from near the base of the unit. This material appears to have accumulated as a locally derived alluvium and colluvium in response to spring sapping and slope wash off the scarp. There is no sign of shearing in this unit.

This horizontally bedded unit passes down abruptly into a very tight and hard to excavate silty medium gravel (unit 4). This has no particular bedding or imbrication on the eastern side of the pit but the more elongate of the schist clasts present become progressively more rotated and imbricate towards the western side. They change abruptly laterally into a dense peat of small twigs and leaves pressed in a fine organic matrix (unit 5). This appears to be the zone of maximum shearing and organics have been trapped during local dilation associated with coseismic displacement between the tight silty gravel and the large boulders of the distal end of the displaced landslide. With depth in the pit the organics of unit 5 become progressively more gritty and gravel rich and finally disappear.

A sample of fine twigs and leaves was extracted from the peat matrix of unit 5 near the ground surface in the inferred shear zone, directly below the horizontally imbricated upper unit which truncates it. This sample (Wk 5529) returned a radiocarbon age of 210 ± 50 yrs BP.

This matches the dates for the Toaroha River event, as indicated in both the Toaroha 2 and Kokatahi 2 trenches (these dates were 210 ± 50 yrs BP and 220 ± 40 yrs BP respectively). This Haupiri River date comes from organic material with virtually no self-age (i.e. the fine twigs and leaves were very young at the time of death). This young organic material has then been trapped.
within the fault plane by the most recent fault movement, whereas the Toaroha 2 and Kokatahi 2 radiocarbon dates come from older units that predate the faulting and therefore provide maximum ages. The very similar radiocarbon date for the actual shear episode in this Haupiri River trench suggests that the Toaroha River earthquake event occurred very close to the maximum ages provided by the other two trenches.

The Haupiri shear zone date of 210 ± 50 yrs BP is very close to the youngest limit of radiocarbon dating resolution at around 200 yrs BP. It corresponds to a most likely general time period for the Toaroha River event at around 1700 - 1750 AD. Organic material of this age will sometimes return a finite conventional age of around 200 - 250 yrs BP but will also sometimes appear as "modern" (A. Hogg, pers. comm. 1999). We suggest this is the explanation of the apparently anomalous "modern" dates returned for some of the samples in the Toaroha 1 and Toaroha 2 trenches, which had otherwise been attributed to forest logging disturbance.

The Haupiri pit dates are also a good fit with possible earthquake triggered mass movement and sedimentation features near the Alpine Fault such as the Moss Creek debris flow near the Haupiri River, the Haupiri River aggradation terrace, and the Mt Harry rock fall near the Toaroha River.

In relation to Alpine Fault segmentation models the Haupiri pit is particularly important because it confirms that ground rupture in the Toaroha River event extended past the Hope Fault junction at the Taramakau River. Previously this has been considered by others to be a possible rupture segment boundary (Berryman et al. 1992; Bull 1996).
3.3 TRENCH EVIDENCE FOR A PENULTIMATE EARTHQUAKE RUPTURE AT AROUND 1600 AD – THE CRANE CREEK EVENT

As part of this thesis the Alpine Fault has also been trenched at two localities in the north Westland section of the fault, both of which are further north of the Haupiri River site described in the previous section (refer Figure 1.3 and 2.1 [map pocket] for location details).

3.3.1 Crane Creek Site

Crane Creek is a small river draining the Mt Rochfort basin in the Haupiri Range (GR K32/098465). It is located 5 km northeast along fault strike from the Haupiri River trench site. Previous investigation of the area was carried out in the late 1940’s and the conclusions are summarised by Munden (1952).

The Alpine Fault crosses Crane Creek about 1 km upstream from the Haupiri Road Bridge, and the creek has formed a bend along fault strike (Figure 3.15). The scarp at this location is 22m high with a 20 - 30° scarp slope angle. The scarp slope is covered in a dense forest of mature red beech (Nothofagus fusca) and the occasional member of the podocarp family. There is no bare or eroding ground above the trench, and a thick forest floor litter covers the entire scarp slope. The trench, which was dug by hand and is of limited depth and extent, was located across a small swampy furrow at the base of the scarp marking the most recent rupture at this location.

The trench log is presented in Figure 3.16 with the soil descriptions and the interpretation. The stratigraphic relationships are relatively simple in the trench. The scarp has formed in fluvial silty gravels overlying moraine (unit 1). A shear zone can be defined that contains green - grey silty fault gouge mixed with dense clayey peat (unit 2). Overlying this is a white gravelly silt with minor (unit 3). This in turn passes up abruptly into soft brown peat and twigs which extend without variation to the swamp surface (unit 4).
Figure 3.15 - General location of the trench site elevated approximately 60 m above Crane Creek. The location of a 4 m high aggradation terrace further down stream in the current river channel, which contains buried logs which have been radiocarbon dated at 390 ± 50 yrs BP (Wk 4343), is also shown. The section A – A shows the Alpine Fault scarp, which has a vertical offset at this location of 22 m, and the location of the trench excavated at the base of the scarp across a small linear furrow.
Figure 3.16 - Log of the trench across the fault furrow shown in Figure 3.15. Inferred coseismic and directly post seismic sediment (unit 3) yields dates of 360 ± 50 yrs BP (Wk 4489) and 380 ± 25 yrs BP (Wk 5263). The peat above this (unit 4) contains no sign of a more recent coseismic sediment pulse.
I infer the gravelly silt of unit 3 formed as a coseismic and immediately post-seismic scarp erosion sediment, which has rapidly filled the furrow in the first few rains following the earthquake. The scarp has then settled back down to a stable and well-vegetated state, and subsequently purely organic detritus filled the furrow to the present day. Material was collected for dating from the unit 3 and two locations in the gouge (unit 2). Conventional radiocarbon dates were obtained of 360 ± 50 yrs BP (Wk 4489) for unit 3 and 5430 ± 60 yrs BP (Wk 4490) and 5030 ± 60 yrs BP (Wk 4492) for the peaty gouge (unit 2). The gouge dates may reflect a major trapping of organics in the fault plane during a strongly dilatant rupture event around 5000 - 6000 years ago however these peaty gouge dates do not help to constrain the timing of more recent events.

A second twig sample from unit 3 was submitted for high precision dating of extracted cellulose to refine the age estimate and has yielded a radiocarbon date of 380 ± 25 yrs BP (Wk 5263). The calendric calibration of this high resolution date adopts the southern hemisphere offset of Vogel et al., 1993 and two sigma error limits. Using the calibration curve of Stuiver & Reimer, 1993 the possible calendric age for the inferred earthquake at this location is between 1480 and 1645 AD. This cannot be reconciled with the Toaroha River event which was definitely post 1660 AD and probably post 1700 AD.

This older earthquake is named here as the Crane Creek event. As noted earlier the adoption of a name for this event, as opposed to some preferred date, recognises the likelihood of further improvements in estimates of the exact timing of this earthquake. This name does not imply an epicentre at or near Crane Creek, or suggest that this event was restricted to only this area.
Figure 3.17 - Detailed plan of the fault scarp in the vicinity of the Crane Creek fault trench. The furrow through which the trench has been excavated has been covered over by a landslide that post-dates the last rupture. Red beech trees growing on the landslide (B1 and B2) have estimated ages of 236 and 295 years respectively, indicating no rupture at this location since approximately 1700 AD. A 540 year old tilted rimu (R) predates the shallow landslide, which may have been earthquake triggered.
There is no evidence at this location for the younger *Toaroha River event*. In the trench itself there is no evidence of renewed detrital sedimentation within the top forest litter, or any sign of subsequent deformation or shearing of the dated scarp erosion sediment. Detailed mapping of the scarp at this location (Figure 3.17) demonstrates there are no other locations on the scarp where a more recent rupture may have occurred. The cumulative dextral offset of the 60m terrace riser is only present on the strike projection of the trenched fault furrow which also implies this is the only zone of shearing.

A small landslide, which may (or may not) have been earthquake triggered, has filled the furrow 30m further northeast along strike. An old rimu tree (*Dacrydium cupressinum*) has been partially buried by this landslide and two red beech trees (*Nothofagus fusca*) grow on the upper landslide surface. This has been cored with the assistance of Andrew Wells who has subsequently carried out tree ring counts on the thin "straws" of wood the corer extracts. This work shows the rimu tree which was inundated by the slide commenced growing around 1450 AD.

The younger beech trees growing on the surface of the landslide reached corer height at least as long ago as 1700 AD (Fig. 3.17). This makes no allowance for any time for establishment, germination or seedling growth to corer height, which is typically around 50 years but may range from 20 - 100 years (Wells et al., 1998). The absence of a more recent fault offset of the landslide debris indicates there has definitely been no rupture at this location since at least 1700 AD, and if establishment time is included, probably none since around 1650 AD.

*Matching aggradation terrace*

In Crane Creek itself, 600m downstream from the trench site, there is a low aggradation terrace 4m above the modern river. The coarse gravel forming this
is loose, sandy and poorly imbricated. In one location several large red beech trunks protrude and the sapwood of one of these has a conventional radiocarbon age of $390 \pm 50$ yrs BP (WK 4343, refer later Table 4.1 for details). A photograph of this locality is presented as Figure 4.1 in the next chapter.

Growing on the surface of the aggradation terrace is a dense stand of red beech (*Nothofagus fusca*) which has an even age structure typical of a colonising stand. Unfortunately many of the largest beech trees are now rotten and have fallen relatively recently with a new generation of younger beech replacing them. Coring of the largest of the remaining trees indicates the oldest tree still surviving is 295 years. To this age must be added a period to establish and grow to the height of the corer. This is normally estimated to be 30 to 50 years. This data alone suggests a minimum age of formation of the aggradation surface of around 330 years before present. When some allowance is made for the probable loss of the oldest trees, and also the time needed for the river to cut back down and regularly abandon the surface after forming it, a range of 350 - 400 years is indicated.

The excellent match between the dated aggradation terrace and the fault trench date at this location provides support for the use of dates from aggradation terraces in this type of indirect paleoseismic analysis in other areas of Westland. This method is discussed further and applied in Chapter 4.

### 3.3.2 Ahaura River Site

Approximately 6 km farther northeast along strike from Crane Creek, the Alpine Fault crosses the Ahaura River (Fig. 1.3; GR L32/138488). This is a large river system that coincides with the postulated westward extension of the Clarence Fault to join the Alpine Fault. Little geological investigation out in this area and nothing has been published. Earth Deformation Section of the former New Zealand Geological Survey briefly inspected the trace in 1975 and 1985.
(Officers of the Geological Survey 1975; 1985) noting the sites potential for paleoseismic investigation.

The Alpine Fault at this location has formed a double surface fault rupture trace and the geomorphic relationship between the truncated terraces and the fault traces indicate that the northwest trace, which has the sharpest and steepest profile in the field, is the youngest of the pair (Figure 3.18). A site was selected for trenching along this northwest trace at a location where there is a swampy furrow directly below the scarp.

The trench face log for the trench is presented in Figure 3.19. The scarp has formed in Ahaura River derived gravels (unit 1) and an associated overbank silt terrace cap (units 2 and 3), with local schist derived fan gravels (unit 4) and swamp sediments (unit 5) on the relatively down-thrown northwestern side.

The Ahaura River drains a dominantly greywacke catchment and the high energy river also preferentially removes the softer schist clasts. As a result post glacial Ahaura River sediment is sandy gravel containing coarse well rounded greywacke cobbles and unit 1 in the trench is a typical example. The critical feature in the trench log is the sudden influx of a locally derived, poorly sorted, schistose, fine angular gravel (unit 4) onto a fissured silt unit below the scarp. It appears that the most recent earthquake at this location has down-faulted and fissured the soft silt. It has also triggered landslides and debris flows in the steep schist catchment of the small creek near the trench (Fig. 3.18). This has created aggradation in the creek and the schist sediment of unit 4 has swept southwest around the scarp base and rapidly infilled the down-faulted area before weathering could degrade or obliterate the fissures.

A small branch in the base of unit 4 has yielded a radiocarbon date of $380 \pm 60$ yrs BP matching the timing of the Crane Creek event. A "modern" date (indicating a post 1700 AD age) has been obtained from twigs near the base of the organic unit 5 that has subsequently accumulated over the schist gravel.
Figure 3.18 - The Alpine Fault surface rupture traces at the Ahaura River (GR L32/138488) and the location of a trench across a shallow depression at the base of the youngest scarp (denoted by the black bar and section symbol). Note the close proximity of the trench site to Coates Creek and its location within the associated fan. Coates creek is a small creek draining part of the steep schist slopes surrounding the wide greywacke dominated post glacial floodplain of the Ahaura River.
Figure 3.19 – Face log of the trench at the location shown in Figure 3.17 (GR L32/138488). The last event is marked by the influx of schist derived aggradation material which has infilled fissures in the underlying deformed silts. A radiocarbon date of 380 ± 60 yrs BP has been obtained from the gravel which closely matches both of the dates from the Crane Creek trench.
Once again there is no evidence of a more recent aggradational pulse in the uniform organic top unit, and no evidence of rupture or folding of the 380 ± 60 yrs BP lower unit. The absence of both features is consistent with the evidence at Crane Creek and suggests that this event which occurred between 1480 and 1645 AD was the most recent earthquake at both these northern locations.

The crane Creek event can, therefore, be recognised in Alpine Fault trenches at two localities 6km apart in the northern section of the fault and in both these cases it appears to be the most recent earthquake. However, only another 5km southwest along fault strike from Crane Creek, at the Haupiri River, there is the first evidence of the younger Toaroha River event which then appears consistently as the most recent event in all the other Central Alpine Fault trench sites. This indicates the northern limit of surface rupture for this youngest earthquake was somewhere between the Haupiri River and Crane Creek. It is possible the rupture splayed off to the east of Crane Creek along one of the numerous shear zones which strike east from the Alpine Fault zone in this area. Alternatively the rupture at the ground surface along the Alpine Fault trace may have simply decreased to zero somewhere over this 5 km of dense forest.

The only trench evidence in the central section of the fault for the penultimate Crane Creek event is the evidence of two events in the last 500 years in the Kokatahi 2 trench. Based on the forest disturbance evidence presented later in Chapter 5, which indicates a single period of synchronous forest disturbance within the Crane Creek trench age range extending from the Rahu Saddle to at least the Paringa River, it can be inferred that the penultimate rupture recorded in the Kokatahi 2 trench was the Crane Creek event. The Crane Creek event is also represented in the central section of the fault in the radiocarbon ages of landslides, debris flows and aggradation terraces extending at least as far south as the Karangarua River. This indirect evidence is discussed in more detail in the next Chapter.
3.4 SUMMARY AND CONCLUSIONS

Direct investigation of Alpine Fault paleoearthquake rupture events has been undertaken utilising paleoseismic trenches and pits along the main fault rupture trace. A total of seven trenches and pits have been excavated at four localities along approximately 75 kilometres of the fault. From these excavations a total of 16 radiocarbon dates provide age constraints on the timing of the most recent two earthquakes.

This data shows:

- there have been two earthquake events in the last 500 years along the central section of the Alpine Fault with associated ground rupture and liquefaction at the fault trace.

- there is consistent evidence that the most recent earthquake rupture, which is referred to here as the Toaroha River event, occurred post 1660 AD and most probably between 1700 and 1750 AD.

- the trench evidence indicates that the surface rupture for this event ended in the north section of the fault at the Haupiri River, which is 25 km northeast of the Hope Fault junction with the Alpine Fault.

- prior to 1660 AD, and at sometime between 1480 and 1645 AD, an earlier earthquake designated the Crane Creek event occurred. This is inferred to be the penultimate rupture event recorded in the Kokatahi 2 trench in the central section of the fault.

- the Crane Creek event is the most recent rupture at the two trench sites which are northeast of the Haupiri River.
Chapter 4

INDIRECT EVIDENCE OF PAST EARTHQUAKES FROM TERRACE AGES AND LANDSLIDES

4.1 INTRODUCTION

The paleoseismic trench evidence presented in the previous chapter demonstrates there have been two Alpine Fault ruptures (and associated earthquakes) in the last 500 years. The earliest occurred between 1480 and 1645 AD, and the most recent was post 1660 AD (and most likely post 1700 AD), but definitely before European settlement in 1840.

Historical earthquakes, both in New Zealand and overseas, demonstrate that one of the main secondary impacts of earthquake shaking is extensive landsliding and associated river aggradation. Triggering of landslides occurs over very wide areas and at distances of up to 300km from the epicentre in a large M 8 earthquakes (Hancox et al., 1997) and because of this large earthquakes generally have a much greater regional impact than extreme climatic events. Is there indirect evidence preserved in the record of landslides and aggradation terrace development in Westland to support the evidence for Alpine Fault ruptures at the times indicated by the trenching?

method of analysis to ten landslides spanning an 8000 year period in the Porters Pass- Amberley fault zone in Canterbury.

Unfortunately landslides have many triggers in addition to earthquakes and can often be a local phenomena triggered by some other process. Crozier (1992) reviews the criteria which can be used to distinguish an earthquake origin and notes clusters of earthquakes of a common age as the principle criteria. However, very large landslides like rock avalanches or debris avalanches are most commonly earthquake triggered (Hancox et al., 1997) and there relative abundance can also be an indicator of a past earthquake event.

While landslides often have an origin other than earthquake triggering, and can frequently be local phenomena, significant aggradation in a large river catchment is generally a more profound geomorphic event. This frequently occurs on a regional scale following a large earthquake as opposed to the more localised aggradation events caused by local baselevel changes (e.g. raised baselevels upstream of a landslide dam), vegetation changes including fire damage, and extreme rainfall or storms.

Adams (1978) considered the mass balance between sediment yield in montane regions and the average denudation rate in the Southern Alps, concluding that although they are infrequent, large earthquakes are a very important component in the erosion and sediment transport cycle. More recently Hovius et al. (1997) assessed the sediment flux in the western Southern Alps and noted landslide derived material dominates the sediment discharge.

This chapter examines the available regional record of radiocarbon dates for landslides and aggradation terraces in Westland and presents 14 new radiocarbon dates obtained as part of this thesis. These new dates come from landslides and aggradation terraces at 8 sites located along 190 km of the Central and North Alpine Fault, ranging from the Karangarua River valley in the
southwest, to Crane Creek in the northeast. These new dates are combined with the existing data and compared with the date ranges for the most recent two Alpine Fault ruptures defined by the paleoseismic trenching evidence presented earlier in Chapter 3.

4.2 HISTORICAL EXAMPLES OF EARTHQUAKE TRIGGERED LANDSLIDES AND AGGRADATION

Landslides

Landslides are very commonly triggered in the epicentral areas of earthquakes. The largest earthquake since European settlement in New Zealand, the 1855 Wairarapa earthquake (Mw 8.0 – 8.4; Darby & Beanland, 1992) is estimated to have triggered landslides over an area of 20,000 km² at distances up to 300km from the epicentre (Hicks & Campbell, 1998; Hancox et al., 1997). The smaller 1929 Buller earthquake (M = 7.8) triggered landslides over an area of more than 5,000 km² (Adams, 1981; Hancox et al., 1997).

Hancox et al., 1997 review historical examples of earthquake triggered landslides in 22 historical earthquakes in New Zealand. They show that landslides become significant at earthquakes shaking intensities of MM 6 or greater. The most common landslides during earthquakes are rock and soil falls on very steep cliffs, escarpments, gorges, gravel banks, and high unsupported man-made cuts. Large dip slope failures of Tertiary sandstone and mudstone often occur on gentle to steep slopes (10⁰ – 40⁰). Very large rock avalanches are caused by earthquakes of M 6.5 or greater, on slopes steeper than 25⁰ – 30⁰ and more than 100 – 200m high, especially on strongly shaken high narrow ridges.
Aggradation

The consequence of massive aggradation following a historical earthquake was noted as early as 1770 by Captain J. Cook in his visit to Jakarta (Hough, 1994). These impacts in steep forested tropical regions are well described in the international literature (Wright & Mella 1963; Pain 1972; Pain & Bowler 1973; Tutton & Browne 1994; Schuster et al 1996). In New Zealand Adams (1978) highlighted the importance of the process in his modelling of Southern Alps erosion. A subsequent more detailed study of the profound geomorphic and botanical impact of the 1929 Buller Earthquake is outlined by Pearce & O'Loughlin (1985) and Pearce & Watson (1986).

The terrain affected by landslides and aggradation in the Buller earthquake is steep, densely forested, and is subject to high rainfall. In these respects it closely resembles the Southern Alps range front. In the 1929 Buller event the short term impacts of the sediment supplied by landslides completely overwhelmed the river system. Aggradation raised the bed levels on average 1 - 3.5 m, and as a result inundated standing forest and buried large amounts of wood. Although in some locations the rivers have subsequently cut channels into the aggradation surfaces only a negligible proportion of the sediment delivered by the landslides has been removed since the earthquake (Pearce & O'Loughlin 1985).

Past Alpine Fault earthquakes will have had impacts at least as severe as the Buller earthquake and logically these should be reflected in the $^{14}$C ages of aggradation terraces and landslide deposits. In addition Alpine Fault rupture events change the regional base level along the range front, uplifting surfaces and thereby creating flights of terraces on the relatively uplifted southeast side. This increases the chances of finding wood in the modern river banks which are created by the subsequent river downcutting and helps in the recognition of individual events by physically separating the past aggradation surfaces.
Figure 4.1 - Photo from Pearce & Watson (1986) showing the extent of aggradation in the Matiri River catchment in the epicentral area following the Buller earthquake of 1929. In many cases such as this the aggradation is most apparent immediately upstream of large landslides which have partially dammed the river. This photo is taken 60 years after the earthquake but aggradation sediment at this location is still yet to be dissected and terraced.

Figure 4.2 – The epicentral area of the relatively small Avoca (Arthurs Pass) earthquake of 1994 ($M_w 6.7$) showing wood from landslides in the process of being buried by aggradation. Photo courtesy of Chris Chamberlain.
4.3 EXAMPLES OF AGGRADATION TERRACES NEAR THE ALPINE FAULT

Adams (1980) assumes the terrace record of rivers draining the Southern Alps range front reflects aggradation following massive sediment pulses created by Alpine Fault earthquakes and appears to discount the possibility of other factors. In reality the terrace history of the rivers draining the rangefront is more complex with local aggradational episodes in individual catchments overprinting the inferred regional earthquake signature.

*Terraces which are likely to be earthquake related aggradation*

Figure 4.3 is a photograph of the aggradational terrace from Crane Creek where in this study radiocarbon dates for this terrace are established. In this example there is excellent correlation between the age of an aggradational terrace and the date from the nearby trench (refer section 3.3.1). This provides support for Adams assumption.

Figure 4.4 shows another typical aggradation terrace near the Alpine Fault containing buried logs in the Poerua River. This type of terrace is now often discontinuous up and down the river valley, due to the lateral river erosion following down-cutting. In both examples the gravel making up the terrace is generally loose, sandy and poorly imbricated, reflecting the inability of the river to sort and transport the sediment influx. Wood collected for radiocarbon dating at locations such as this provides a method of estimating the approximate time of any earthquake which may have caused the aggradation.
Figure 4.3 - The aggradation terrace in Crane Creek 400m downstream of the Alpine Fault discussed in section 3.2.1. Red beech logs are visible projecting from the loose sandy gravel of the eroding terrace edge which were dated at 390 ± 60 yrs BP comparing well with the trench dates of 380 ± 25 yrs BP and 360 ± 60 yrs BP. For further discussion refer section 3.3.1.

Figure 4.4 - A rimu log projecting from an aggradation terrace near the Alpine Fault in the Poerua River near Hari Hari (refer Fig. 1.3). Sapwood from this log returned a radiocarbon age of 520 ± 60 yrs BP, similar to Adams (1980) inferred event of c. 550 yrs BP.
In many terraces an aggradational upper horizon can be recognised capping an older degradation erosion surface. Presumably this occurs as an aggradational pulse fills the existing channel, and the alluvial fill then spills out, to grade over the top of older low level degradational terrace treads further downstream. In other cases exposures indicate a particular terrace tread is almost entirely degradational but the same terrace tread further downstream is underlain by younger aggradational material. This presumably results from of a new length of river bed cutting associated with channel avulsion (i.e. diversion and cutting of a degradational tread) during a dominantly aggradational episode. Thus it appears many terrace treads, in both plan and crosssectional view, are a complex mix between the two basic terrace types.

Finally it should also be noted that in some of the very large river catchments, where stream power is relatively high with respect to sediment load, the response to periodic uplift at the Alpine Fault is the cutting of flights of entirely degradational terraces. This is described and discussed in the next chapter (section 5.3) with an example from the Karangarua Valley. In this type of setting radiocarbon dates from wood etc within the gravels into which the flights of terraces are cut are unrelated to the age and mechanism of subsequent terrace formation.

Up till now I have described the type of aggradation feature which can be found close to the Alpine Fault, and indicated in some cases the likely link between these features and past Alpine Fault earthquakes. However, it is important to also note that very similar geomorphic features occur in these same areas which are known to be triggered by processes other than earthquake, and an example of this is described in the next section.
Figure 4.5 – The remains of buried standing trees in a creek fan surface draining the Alpine Fault zone at Moss Creek, Haupiri River. This is a small scale paleo-equivalent of the type of aggradational surface which can be seen in the earlier Figure 4.1 burying standing trees after the 1929 Buller earthquake. Note the absence of a diameter increase near ground level in this buried tree remnant. The root plate was found by excavation at a depth of 1.4 m below the fan surface. Sapwood immediately adjacent to the bark was radiocarbon dated at 210 ± 40 yrs BP (Wk 4876), and is a good match with the nearby trench date of 210 ± 50 yrs BP (WK 5529) obtained from organics trapped in the shear during the last rupture (refer section 3.2.3). An aggradation terrace in the nearby Haupiri River contains buried logs of a similar age (260 ± 50 yrs BP, WK 4874).
Aggradation in the Alpine Fault zone from triggers other than earthquake

An excellent historical example of local aggradation is “Robinson Slip” on the true right of the Waitaha River in central Westland (GR 134/265875). This is very close to the paleoseismic trench site of Wright (1998), and this landslide has occurred within weakened Alpine Fault zone rocks, the same general location where coseismic landslides are normally the most abundant.

This old landslide feature is now marked by a prominent fan which has built out of Macgregor Creek, a small second order tributary to the Waitaha River, in which a large landslide occurred in 1903. This was after a period of sustained heavy rain and extreme local flooding (Jim Ferguson, pers. comm. 1996). The tributary drains the Alpine Fault zone and the fan is composed of mylonites, protomylonite, rare gouge (cataclasite) and high grade schist.

The debris flow associated with the landslide formed most of the large fan shape, with the fines extending the greatest distance and flowing onto the upper Waitaha Valley road. Accounts at the time describe horses getting stuck in the “quick sand”, and it took many weeks to clear the road. A large area of mature forest was inundated and completely covered by the debris. The fan continued to grow for a period after the landslide, as other smaller failures occurred on the margins of the main landslide, and the upper catchment disgorged the remaining debris.

Figures 4.6 and 4.7 are photos taken in 1996 in Macgregor Creek, which has now incised into the aggradation surface of the original fan. This shows trees buried in growth position which are being exhumed by the subsequent downcutting. The height of the trunks conforms to the original aggradation surface where most of the tops of the dying trees have subsequently broken off. This process of trunk trimming at the new ground level is apparently very rapid (often just 1 - 2 years) and assisted by massive infestations of pinhole borer which occur amongst so many dead trees. Not all trees have fallen, and in
Figure 4.6 - Macgregor Creek on the true right of the Waitaha River near the Alpine Fault (GR 134/265875). The creek has re-exposed buried forest from a large rainfall triggered landslide which occurred in 1903.

Figure 4.7 - A close up view of the current creek edge where standing trees are being exhumed from the active terrace riser.
some areas of the fan surface, the heart wood of dead trunks still protrudes amongst the young forest colonising the surface.

Older examples of this type of non-seismic local burial event no doubt exist in many other places along the Southern Alps range front. The critical difference is that these type of features are seldom synchronous over a very large area. Apparently the rain responsible for Robinson Slip was severe in the Waitaha catchment and areas immediately to the south and north, but was not a widespread regional extreme event. Similarly, although the sediment overload was profound for Macgregor Creek there is no obvious equivalent new terrace in the much larger Waitaha River, which was in flood at the time, and this also indicates the essentially local impact.

Robinson's Slip is a good example of the type of sediment overload which would happen synchronously on a much larger scale during an Alpine Fault earthquake. This type of seismically triggered aggradation would be enough to extend into the large rivers and well forward of the range front, and in some areas the impacts of this aggradation would reach the coastline.

Robinson's slip indicates it is inevitable that any group of prehistoric landslides and aggradation terrace radiocarbon dates collected from the area adjacent to the Alpine Fault will include at least a few examples which are non seismic in origin. However, other historical earthquakes examples indicate that in this type of location Alpine Fault earthquakes are likely to numerically dominate the landslide and aggradation terrace record. Trenching indicates Alpine Fault earthquakes occurred within defined time ranges so the record of the ages of these secondary geomorphic features can now be compared to the trench date ranges to check for any possible agreement.
4.4 RADIOCARBON DATING OF TERRACES AND LANDSLIDES -
RESULTS FROM AN ENLARGED DATA-SET

As part of this study a further fourteen $^{14}$C dates have been obtained from aggradational gravels, fans, and landslide deposits for the last 750 year period. These new dates have been combined with three collected by others (T. Chinn pers. comm. 1996, Basher, 1986) and two dates originally used by Adams in this youngest time period. These dates have a geographic spread from Crane Creek to the Karangarua River (more than 200 km, refer Figure 1.3 for location details) and Table 4.1 summarises the site information details.

Ten $^{14}$C dates come from terraces and nine from mass movement deposits including rock avalanches, rock falls, landslides, debris avalanches and debris flows. To try to help focus on potential earthquake triggered failures most of the landslides in Figure 4.8 are from either large-scale deep-seated catastrophic failures, for example rock or debris avalanche deposits within 10 km of the Alpine Fault, or smaller rockfalls and debris flows right on the fault scarp. The common association between the larger scale catastrophic features and earthquakes is well documented, for example Keefer (1984a, 1984b, 1994).

Figure 4.8 presents the two sigma calendric conversions for these dates adopting the methods of Stuiver & Reimer (1993) and the 40 year southern hemisphere offset of Vogel et al (1993). All the available dates for this time period are included in Figure 4.8, none have been excluded because of a poor match. Despite this the repetition of common date ranges is immediately apparent, particularly in the middle group. Detailed fluctuations in the radiocarbon calibration curve (refer Appendix 1) for some dates in the middle group (for example the Kokatahi 5m terrace) produce several discrete possible calendric time ranges for a single dated sample. As a result three of the dates in the middle cluster could belong the youngest cluster, but they have a much higher probability of belonging to the middle group.
<table>
<thead>
<tr>
<th>Feature</th>
<th>Description and location with respect to the Alpine Fault</th>
<th>$^{14}$C age + error</th>
<th>Lab. code</th>
<th>Collector</th>
<th>Grid ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moss Creek Debris Flow</td>
<td>Youngest sapwood from a totora stump, buried by the distal end of debris flow from a creek 300m east of the fault.</td>
<td>210 ± 50 yrs BP</td>
<td>WK 4876</td>
<td>M. Yetton</td>
<td>K32/060435</td>
</tr>
<tr>
<td>Haupiri River 3m Terrace</td>
<td>Youngest sapwood from a red beech log in a terrace on the true left of the Haupiri River, 200m upstream of the fault.</td>
<td>260 ± 50 yrs BP</td>
<td>WK 4874</td>
<td>M. Yetton</td>
<td>K32/056437</td>
</tr>
<tr>
<td>Mt Harry Rockfall</td>
<td>Punga fragment in granitic rockfall debris from the east side of Mt Harry near the Toaroha River, 200m west of the fault.</td>
<td>200 ± 50 yrs BP</td>
<td>WK 4919</td>
<td>M. Yetton</td>
<td>J33/567100</td>
</tr>
<tr>
<td>Crane Creek 4m Terrace</td>
<td>Youngest sapwood from a red beech log in a terrace on the true left of Crane Creek, 100m west of the fault.</td>
<td>390 ± 50 yrs BP</td>
<td>WK 4343</td>
<td>M. Yetton</td>
<td>K32/098465</td>
</tr>
<tr>
<td>Styx Rockfall</td>
<td>Peat in voids of protomylonite rockfall debris which has fallen into a peat swamp at the base of the 30m high fault scarp.</td>
<td>430 ± 60 yrs BP</td>
<td>WK 4918</td>
<td>M. Yetton</td>
<td>J33/602126</td>
</tr>
<tr>
<td>Kokatahi River 5m Terrace</td>
<td>Twigs from near the top of the terrace on the true right of the Kokatahi River, 30m downstream of the fault.</td>
<td>330 ± 60 yrs BP</td>
<td>WK 4009</td>
<td>M. Yetton</td>
<td>J33/579103</td>
</tr>
<tr>
<td>Toaroha River 1m Terrace</td>
<td>Small rata branch near the top of a terrace in a sandy unit. The terrace is down - warped and is 25m west of the fault.</td>
<td>320 ± 60 yrs BP</td>
<td>WK 4014</td>
<td>M. Yetton</td>
<td>J33/572102</td>
</tr>
<tr>
<td>Poerua River 6m Terrace</td>
<td>Sapwood from a rata branch in a loose sandy gravel terrace on the true left of the river, 1 km downstream of the fault.</td>
<td>370 ± 60 yrs BP</td>
<td>WK 4340</td>
<td>M. Yetton</td>
<td>J34/083747</td>
</tr>
<tr>
<td>McTaggart Debris Avalanche</td>
<td>10 year old sapwood from an exhumed miro stump under a kahikatea branch near the top of a terrace in a sandy unit. The terrace is down - warped and is 25m west of the fault.</td>
<td>420 ± 25 yrs BP</td>
<td>WK 5264</td>
<td>M. Yetton</td>
<td>H36/534259</td>
</tr>
<tr>
<td>Cropp River Rock Avalanche</td>
<td>Twigs from the base of a schist debris avalanche deposit on the true right of the Cropp River, 9 km east of the fault.</td>
<td>360 ± 50 yrs BP</td>
<td>NZ 5252</td>
<td>L. Basher</td>
<td>J34/478907</td>
</tr>
<tr>
<td>Parker Creek Debris Flow</td>
<td>Branch from the base of a schist debris flow and fan deposit overlying Whataroa River alluvium, 300m west of the fault.</td>
<td>430 ± 50 yrs BP</td>
<td>NZ 6684c</td>
<td>T. Chinn</td>
<td>J35/003673</td>
</tr>
<tr>
<td>Konini Creek Debris Flow</td>
<td>Twigs and peat from the base of a schist derived debris flow from a side creek of the Haupiri River, 700m west of the fault.</td>
<td>510 ± 60 yrs BP</td>
<td>WK 4875</td>
<td>M. Yetton</td>
<td>K32064454</td>
</tr>
<tr>
<td>Styx River 2m Terrace</td>
<td>Heartwood of a small rata branch in a low terrace on the true right of the Styx River, 300m upstream of the fault.</td>
<td>680 ± 50 yrs BP</td>
<td>WK 4011</td>
<td>M. Yetton</td>
<td>J33/595122</td>
</tr>
<tr>
<td>Toaroha River 3m Terrace</td>
<td>Sapwood of a large kamahi log in terrace base, true left of Toaroha River, 200m upstream of fault.</td>
<td>650 ± 60 yrs BP</td>
<td>WK 4019</td>
<td>M. Yetton</td>
<td>J33/572099</td>
</tr>
<tr>
<td>Toaroha River 2.7m Terrace</td>
<td>Sapwood fragment of rata log in same terrace as above but on the true right of the river and 100m upstream of fault.</td>
<td>580 ± 60 yrs BP</td>
<td>WK 4013</td>
<td>M. Yetton</td>
<td>J33/573102</td>
</tr>
<tr>
<td>Muriel Creek 4m Terrace</td>
<td>Sapwood for ribbonwood branch in fine schist gravel terrace on true left of Muriel Creek, 400m upstream of the fault.</td>
<td>540 ± 60 yrs BP</td>
<td>WK 4439</td>
<td>M. Yetton</td>
<td>J33/502032</td>
</tr>
<tr>
<td>Waitaha Fan</td>
<td>Wood of unknown species and self age buried in fan gravels on true left of Waitaha River, 500m downstream of the fault.</td>
<td>680 ± 40 yrs BP</td>
<td>NZ 4629c</td>
<td>J. Adams</td>
<td>J34/254866</td>
</tr>
<tr>
<td>Wanganui River 10m Terrace</td>
<td>Heartwood sample with estimated 80 years allowed for missing rings from base of terrace 6 km east of fault.</td>
<td>610 ± 70 BP</td>
<td>NZ 4628c</td>
<td>J. Adams</td>
<td>J34/210763</td>
</tr>
<tr>
<td>Geologists Creek Rock Avalanche</td>
<td>Heartwood from small branch in 15m thick mylonite and gouge rock avalanche deposit 300m from fault scarp.</td>
<td>550 ± 50 BP</td>
<td>NZ 6471</td>
<td>T. Chinn</td>
<td>J33/607151</td>
</tr>
</tbody>
</table>

Table 4.1: Sample details for dated aggradation terrace and mass movement deposits less than 750 yrs old within 10 km of the central and north Alpine Fault
Figure 4.8 - Terrace and landslide age ranges for all dated examples within 10 km of the Alpine Fault in north and central Westland. Refer Table 4.1 for the site information. Also shown for comparison are the possible date ranges for the most recent two Alpine Fault earthquakes defined by the trenching work presented in Chapter 3.

The considerable time span for each date is the inevitable result of the radiocarbon dating method and this is particularly the case with dates for the last 500 years where the calibration curve has many deviations from a simple equation. In this age range increased laboratory analysis precision (for example ± 20 or 25 years) offers only a minor increase in resolution. The most
example ± 20 or 25 years) offers only a minor increase in resolution. The most important feature in Figure 4.8 is the apparent coincidence of the date ranges, rather than their precise calendric resolution. Although it should be noted that the stair-step shape of the calibration curve has a clustering effect on raw radiocarbon dates, this is largely offset by the desegregation back into a full calendric date range (refer discussion in Appendix 1). Some artificial enhancement does still remain, particularly for dates close to 1700 AD (McFadgen et al., 1994). Equally however the normal statistical variations in the dating method also mean that a series of radiocarbon dates from the same age piece of wood will look just like the middle cluster in Figure 4.8.

4.5 DISCUSSION

Figure 4.8 shows that there is a reasonable agreement between the trench age ranges and the ages of landslides and aggradation terraces so the data is at least consistent with episodes of landsliding and regional aggradation triggered by Alpine Fault earthquakes.

Not withstanding the limited dataset with respect to statistical analysis some support for the separation of these dates into three clustered groups comes from the sequence of terrace dates on the true right of the Toaroha River. Here a date of 320 ± 60 yrs BP (Wk 4014) comes from a down warped terrace at the fault scarp that is now only 1 m above river level. This date fits into the middle 1600 AD cluster. The down warping of this terrace tread implies a younger earthquake, possibly the 1700 AD event. The next terrace above the surface with the 1600 AD date yields a date of 580 ± 60 yrs BP, which fits in the oldest 1400 AD group, and suggests separation by uplift during the 1600 AD event. The earthquake triggered aggradation process responsible for burying and preserving wood in a region of rugged and forested catchments such as central Westland is summarised in Figure 4.9.
Figure 4.9 - A summary of the aggradation process responsible for burying wood following a large earthquake in a steep forested catchment in Westland.

What about other alternative explanations for this apparent clustering in the dates of landslides and aggradation terraces? Heavy rainfall and flooding is a short-term climatic impact which can trigger landslides, and in some cases cause local aggradation. However, there are some important differences between the sedimentary products of earthquake triggered aggradation, and those of high intensity rainfall induced floods.
Floods are the main method by which the rivers clear themselves of relatively buoyant material such as wood. Apart from the wood temporarily stored on high areas by receding floodwaters, it is extremely rare for wood to be trapped in gravel bed load which is being moved during a flood. Floods are characterised by an abnormal increase in stream power, with an associated exponential increase in bed load and suspended sediment carrying capacity, and a corresponding increase in sorting efficiency.

As a result, in any major river only the very largest of rain triggered landslides will leave a large volume of organic rich gravel in the main river bed. In cases such as this the landslide aggradation material will generally be confined to second order catchments (for example Macgregor Creek). The existence of wood in a major river valley, trapped within the poorly sorted and loose bedload of the main river and now exposed in a cut terrace, is a reasonable indication that a flood has not been responsible for its deposition.

In contrast an earthquake results in a massive sediment overload relative to stream power and carrying capacity, even for the largest rivers. The resulting aggradational sediment is typically less sorted and more sandy than normal, with coarse gravel clasts which are often more angular, and the sediment is less imbricated and more loosely packed. Most importantly from the perspective of future dating, earthquake derived aggradation sediment is often locally rich in buoyant material such as wood and plant matter, which normally the river would have winnowed out. It follows that for this reason alone the post-earthquake environment of excess sediment favours the preservation of a datable record. However, in addition there is the abnormally high amount of organic material which is entering the river at such a time from the simultaneous impact of the earthquake on the tree and plants. The wood and vegetation from earthquake induced forest damage enters the river system at the same time that an excess of sediment is available to bury and preserve it.
The earlier Figure 4.1, which shows the burial of both fallen and standing forest by aggradation in the Matiri River after the 1929 Buller earthquake, is a classic example of sediment overload relative to stream power and carrying capacity. More than fifty years later this section of the Matiri River has still not recovered. Presumably the river at this location will ultimately down-cut again, and convert this alluvial fill surface to a corresponding terrace tread and riser. When this occurs the terrace riser (i.e. the new river bank) will expose a loose sandy alluvium which is rich in wood and plant material, the dating of which will reflect the event age. In contrast to the Matiri River, for rivers which cross the Alpine Fault, the subsequent process of terrace creation is made much simpler and faster by the component of co-seismic uplift in each successive earthquake.

There are also some possible implications for age interpretations based on the dates from wood in earthquake triggered aggradation gravels. Some of the dated wood and plant material would have been killed synchronously by landsliding associated with the earthquake. The $^{14}$C date from the youngest sapwood of such material will then reflect the true event age. However, there are also two mechanisms by which the radiocarbon age can overestimate the time of the event. The first is where the wood is fragmented or abraded and only the older stronger heartwood is preserved. This is an example of an "in-built" wood age.

In addition to this in-built age a second process can conspire to bias the dates obtained. At any one moment of time the highest points in the active river beds of Westland are normally littered with logs and wood left by receding flood waters. This is old wood from trees that died years ago, and sometimes tens of years ago. Although most of this wood is being progressively floated away in successive floods, some of the more durable wood can survive on surfaces like this temporarily for along time. It is likely some of this type of material will also be buried during an earthquake triggered aggradation pulse, and the subsequent reliance on such dates to estimate the earthquake timing could distort the event age.
For both these reasons it is likely the $^{14}$C dates of wood samples in aggradation gravels, when averaged, will tend to have an age mode older than the aggradational event responsible for their burial. However, by wherever possible dating bedload wood which has the sapwood still intact, and preferably wood which appears relatively fresh and undecayed, it is possible to minimise both of these potential errors.

Fires can also cause landslides and local aggradation by vegetation disturbance, but these landslides tend to be triggered after the fire in times of heavy rain, and once again when stream power is relatively high. In addition fires are not common in the humid and wet range front of Westland.

The best criteria for assuming the record of aggradation reflects earthquakes in Westland as opposed to the alternative explanations outlined above, is the regional extent of any pattern. A check of available records indicates there is no historical precedent for fires or storm induced regional river aggradation in Westland over the past 160 years (Benn, 1990). Westland weather and floods tend to be characterised by localised extremes of rainfall occurring within a particular regional wet cycle. The landscape has generally adapted to the high annual rainfall and the regular wet cycles so widespread damage is not common.

**4.6 MEAN AGES FOR LANDSLIDES AND TERRACES**

For the reasons outlined above it is reasonable to expect that the two earthquakes which are indicated by the trench data will dominate the record of landslide and terrace ages for the last five hundred year period. However, it is also likely that at least one or two of these dates may be affected by wood in-built age or could be of a non-seismic origin. Fortunately one or two incorrect dates will make little difference to a mean age, at least where the number of dates is significant.
If the assumption is made that the eight dates which appear to be a good match to the Crane Creek event in Figure 4.8 were all triggered by an earthquake at around that time, then the mean can be calculated using the method of Wilson & Ward (1978, 1981). The mean for this Crane Creek group is 350 ± 16 yrs BP which has an associated calendric range of 1480 – 1640 AD.

Applying this same principle to the inferred Toaroha River event, which is the smallest group and therefore the most sensitive to any incorrect dates, a mean is obtained of 207 ± 30 yrs BP with a corresponding calendric range of post 1660 AD, but prior to European settlement in 1840. The oldest group of 8 dates in Figure 4.8, which was the earthquake originally inferred by Adams (1980), averages to 514 ± 20 yrs BP and has a much narrower calendric range (by virtue of the steeper calibration curve in this period) of 1420 –1450 AD.

4.7 AN ATTEMPT AT PRECISE DATING OF A LARGE LANDSLIDE – THE MCTAGGERT DEBRIS AVALANCHE

An attempt was made to refine the age estimate of a typical representative of the largest middle Crane Creek event group of apparently clustered landslide and aggradation terrace ages in Figure 4.8. It was hoped a comparison between this refined age and the age range from the Crane Creek trench would provide a better test for any possible synchronicity.

The Karangarua River, south of Franz Josef, is a landslide site 200km southwest of Crane Creek which has yielded radiocarbon dates in this age range. Peter Wardle obtained a date of 370 ± 60 yrs BP on buried stumps exposed in the main river near the McTaggert Creek tributary (NZ 1292; reported in Adams, 1980). This location was revisited, in conjunction with Andrew Wells who has been working on forest age patterns in this catchment, and samples were collected for high resolution dating.
Figure 4.10 – A sketch of the materials in the toe of the McTaggart Creek debris avalanche, as exposed in the Karangarua River, south of Fox and Franz Josef townships (GR H36/534259). Wood is present both in the angular debris and as broken tilted stumps in growth position. Radiocarbon dates on the stumps have yielded ages of 360 ± 60 yrs BP and 420 ± 25 yrs BP, while wood in the debris has provided a date of 320 ± 60 yrs BP. Trees on top of the fan began establishing around 1630 AD and best fit to all dates is a triggering earthquake between c.1610 and 1630 AD. The location of the debris avalanche with respect to the Alpine Fault is shown in the map below the sketch.
Figure 4.10 shows the McTaggert Creek location and sample details. The site is 6 km southeast from the Alpine Fault, and the distal end of the debris avalanche is approximately 2 km from the steep valley slopes from which it has been derived (Figure 4.10). Mass movement features on this scale are commonly triggered by a large earthquake. Tilted dead standing trees, the stumps of which now protrude from the modern river, have been buried by a debris avalanche of angular schist debris which has swept into the original forest area. This debris avalanche material has been subsequently covered over by fan sediments.

An extensive alluvial fan surface has developed over this debris avalanche. There are sands and fine sediments immediately above the avalanche debris deposit and below the typical alluvial fan gravels. This suggests there may have been a temporary landslide dam behind the avalanche deposit which then filled with sands and fine gravels prior to dam burst and the resumption of normal coarse gravel fan sedimentation from McTaggert Creek.

The Karangarua River has subsequently cut back down through all this to expose the basal stumps in the main river at the McTaggert Creek confluence. Meanwhile up on the fan surface a new forest has re-established and Wells cored trees from this surface to define the minimum age of the surface and the time of forest colonisation (Wells, pers. comm. 1987; Wells, 1998).

Figures 4.11 shows one of the stumps in the modern river, and Figure 4.12 shows the typical angular schist exposed in the river trimmed end of the debris avalanche. Broken wood can also be found mixed with the debris in many locations. Sapwood averaging 10 years of age was collected from a different buried stump to that dated by Wardle. The date obtained was $430 \pm 25$ yrs BP (effectively $420 \pm 25$ years with the sap wood adjustment and reconcilable with Wardle's earlier $360 \pm 60$ yrs BP date). Sapwood from a broken branch 10m further up in the debris avalanche gave a date of $320 \pm 60$ yrs BP.
120

Figure 4.11 (above) - A rimu stump protruding from the modern Karangarua River at the same location. Note the large moraine boulders further upstream on to which the debris avalanche has fallen.

Figure 4.12 - Typical angular chaotic schist debris at the same location which forms the main basal section of the debris avalanche.
The oldest trees growing on the fan surface are conifers, which increment coring indicates are up to 330 years old. Recent work on conifer species in this catchment has refined the establishment, germination and seedling growth time to 36 years with a range of 15 – 60 years (Wells, pers. comm. 1998). This indicates fan formation was complete by c.1630 AD (range c.1650 - 1610 AD).

![Graph showing tree dating and age distribution](image)

**Figure 4.13** - Summary and comparison of the date information from Crane Creek and the debris avalanche in the Karangarua River.
The radiocarbon date information for the underlying debris avalanche is plotted along with the Crane Creek trench age range and the surface tree age in Figure 4.13. The high resolution date for the buried stumps has a minimum age of 1450 AD and maximum age of 1635 AD and this is inferred to be the age range for the debris avalanche itself. Unfortunately there is an indefinable period of time associated with capping the debris avalanche with sands and the upper most coarse gravel.

If the inference of temporary blockage of the main river is correct, then the high bedload of the creeks and rivers would fill the area upstream of the debris avalanche with sand and fan gravel very rapidly (i.e. in years not decades). In addition the Karangarua River is capable of downcutting and removing even a large barrier to its flow relatively quickly. Once this process of downcutting was underway the fan surface would have been abandoned and tree growth was able to initiate. Given that the trees indicate the fan was being colonised by 1630 AD, a process which commences virtually immediately in this area of the humid range front, a date for the debris avalanche of around 1600 AD is suggested.

Unfortunately none of this definitely precludes the alternative of slow sand and gravel capping of the debris so the minimum date of formation at 1450 AD cannot be categorically ruled out. Thus even in this case where there are three radiocarbon dates from the same location, including one with an especially high resolution, and additional tree coring work on the landslide surface, the age resolution for the initiation event remains relatively poor.
4.8 SUMMARY AND CONCLUSIONS

In this thesis I have significantly enlarged the original data-set of Adams (1980) which collates the available radiocarbon ages of aggradation terraces and landslides in Westland. In particular, for the most important period over the last 750 years, the three original radiocarbon dates have now been increased to a current total of nineteen.

There are apparent clusters of dates which suggest periods of regional aggradation and landslides at around 1400 AD and 1600 AD (and to a lesser extent around 1700 AD). These may have been triggered by rupture on the Alpine Fault. The timing of the last two earthquakes, as defined in the fault trenches, match well with the landslide and terrace ages and are consistent with the hypothesis that many of these landscape features were earthquake triggered. The limited geological evidence for terrace uplift and deformation of some of these terraces is also consistent with three earthquake events in the last 750 years.

However, the very large calendric age range associated with radiocarbon dating means it is impossible to conclusively demonstrate regional synchronicity within this data set or refine estimates of earthquake timing. It is also likely that the paleoseismic signal within this data is being muted by both the in-built age of some of the wood, and the inadvertent inclusion of some dates of a non-seismic origin.

Unfortunately, even if regional synchronicity could be proven, this alone does not demonstrate an earthquake origin. This is the inevitable limitation of any paleoseismic analyses based only on an indirect indicator. But in this thesis, where consistent paleoseismic trench data is available to define the time ranges of the past Alpine Fault rupture events, the terrace and landslide record provides important additional evidence that large regional earthquakes did occur during the date ranges indicated.
PALEOSEISMIC EVIDENCE FROM FOREST AGE PATTERNS AND TREE RING CHRONOLOGIES

5.1 INTRODUCTION

Trenching has indicated two earthquakes in the last 500 years, one before 1645 AD, and the other after 1660 AD but before European colonisation. Where is the evidence in the forest and tree record which can logically be expected for these events?

Tree shaking and damage to trees are regularly mentioned in observations of earthquakes e.g. Lawson (1908); Fuller (1912); Louderback (1947); Gu (1983). Some of the original criteria for assigning Mercalli intensity are based on the degree of tree disturbance during the earthquake (Wood & Neumann, 1931).

More recent international examples of seismic effects on trees are the many dead trees around Cook Inlet, Alaska, killed by the 1964 earthquake (Jacoby, 1997). Analysis of forest age and tree growth ring widths to determine the impact of past earthquakes has recently been applied to forested areas overseas in seismically active locations. For example Jacoby (1997); Veblen et al. (1992); Kitzberger et al. (1995).
In New Zealand Berryman (1980) noted red beech trees growing on the White Creek fault scarp with bends in their trunks which tree ring analysis indicates occurred in the 1929 Buller earthquake. Cooper & Norris (1990) noted silver beech trees which had lost their crowns near the Alpine Fault scarp and inferred earthquake shaking was responsible. This evidence was used to better constrain their sag pond radiocarbon dates. In this case the trees were not cored to count the rings but their age was inferred from tree diameter, which is not always a reliable guide (Jacoby, 1997).

Downes (1995) notes beech trees in the epicentral area of the relatively small Lake Tennyson earthquake of 1990 snapped off at a surprisingly constant 1.5 m above the ground. Allen et al. (in press) document damage to beech forest from earthquake shaking and landslides triggered during the moderate size (Mw 6.7) Arthurs Pass earthquake of 1995. Most recently Grapes and Downes (1997) noted that newspaper reports from the time describe the 1855 Wairarapa earthquake (Mw 8.0 – 8.4) triggering landslides that removed one third of the forest on the western Rimutaka range as viewed from Wellington. All these examples indicate past Alpine Fault earthquakes would have had enormous impacts on the forests of Westland.

New Zealand conifer species such as rimu (Dacrydium cupressinum); matai (Prumnopitys taxifolia); miro (Prumnopitys ferruginea) and kahikitea (Dacrycarpus dacrydioides) regenerate prolifically only on recently created or bared surfaces. A large earthquake creates these surfaces by triggering landslides and the associated aggradation and flood damage clears forest from existing surfaces. In addition earthquake shaking causes extensive tree fall in forests not directly affected by deposition, floods or landslides. Tree fall can also result from liquefaction of the underlying substrate. In this manner a large earthquake can cause disturbance over a wider range of landforms than a climatic extreme event such as a wind storm or flood.
This type of damage and the associated impact on the overall age structure of a forest is referred to as forest disturbance. The more specific impact on trees which are able to live through the earthquake is frequently recorded in their growth rings and will generally show up in cross-matched tree ring chronologies.

Dr Andrew Wells (Lincoln University) has been working on forest disturbance in the westland conifer forests during his doctoral studies, continuing the earlier work of Duncan (Duncan, 1981; Duncan & Stewart, 1991) and Stewart (Veblen & Stewart, 1982a; Stewart & Veblen, 1982b; Stewart & Rose, 1989). As far back as the diaries of explorer Charlie Douglas in the 1890s there was speculation that the even-sized stands of trees in the valleys of Westland were the result of synchronous tree regeneration on surfaces exposed by earthquake triggered landslides (Douglas, cited in Hollaway, 1957).

Wardle (1980) noted that numerous prominent and apparently even-aged stands of rata/kamahi (Metrosideros umbellata/Weinmannia racemosa) can be found along the Alpine Fault and suggested the possibility that a major earthquake had triggered massive landslides and allowed simultaneous regeneration. He proposed a date for this event at around 1730 - 40 AD. More recently Wells et al. (1998) followed this up with suggestions that the forest disturbance pattern appeared to coincide with the estimates of Bull (1996) for Alpine Fault earthquakes.

The author first met Andrew Wells after the Wells et al. (1998) paper (which is co-authored with Stewart and Duncan) was in press and discussed our data. New refinements in the estimates of disturbance times which were becoming available at this time were not such a good fit with the Bull (1996) dates for large earthquakes on the Alpine Fault. We (Yetton & Wells) quickly realised a combined geology – plant science approach is required to improve on the earlier estimates of Alpine Fault rupture events. The majority of the field, laboratory preparation and analysis work outlined in this Chapter has been
carried out by Andrew Wells. However we have been frequently working together in the field in several locations (Crane Creek, Toaroha River, Karangarua River valley) and regularly discussing and combining our results as our mutual investigations have proceeded.

Figure 5.1 - Extensive mixed red beech forest and podocarp forest typical of the Alpine Fault zone immediately north of Crane Creek (photo taken from GR K32/096465).
Figure 5.2 - Damage to beech forest in the Harper River after the 1994 Arthurs Pass earthquake (GR K34/980850). Note also the large amount of wood in the aggrading channel. Photo courtesy of Chris Chamberlain.
5.2 REGIONAL FOREST AGE PATTERNS

Figure 5.1 shows the dense forest cover typical of most of the Alpine Fault in Westland. In contrast Figure 5.2 is another photograph taken in the Harper Stream following the 1994 M_w 6.7 Arthurs Pass earthquake. This area was also densely forested below the snowline, but the extensive landsliding triggered by this earthquake, stripped trees from many areas and deposited them in the rivers.

Rupture of the much larger Alpine Fault will cause much greater and geographically extensive damage. It logically follows that this type of massive disturbance event will be reflected in the distribution of forest age, with common modes of tree age corresponding to the simultaneous re-establishment of forest following such a major earthquake event, ultimately resulting in even age forest stands.

Even age forest stands can be recognised visually on many slopes in Westland. Figure 5.3 shows a relatively young example from Goat Creek, near Otira. It is possible to obtain the age of trees in this type of location by extracting a thin core of wood revealing the annual rings. Figure 5.4 shows the process in action. This coring does not kill the tree and it is possible to sample up to 30 trees or more by this method in a day of field work. The information from individual trees can then be combined into forest stand age for the combined population.
Figure 5.3 - An even age forest stand following forest re-establishment on a relatively young (post 1800 AD) landslide near Goat Creek, Otira (GR K33/920180). An earthquake triggers synchronous landslides so large areas of new forest regenerate simultaneously. Photograph courtesy of Glen Stewart.

Figure 5.4 - Coring a large rimu tree (age greater than 500 years) on Surface C on the true left bank of the Toaroha River (refer Fig 3.2) at GR J33/573089.
Figure 5.5 - Forest locations in Westland and Southern Alps for which forest stand age information is available for analysis (Wells et al., 1998; Wells, 1998). Abbreviations as follows: AF = Alpine Fault; CF = Clarence Fault; HF = Hope Fault; MDFZ = Main Divide Fault Zone (Cox & Findlay, 1995).

Wells et al. (1998) compiled all the information available up until early 1997 on forest stand ages throughout Westland. Figure 5.5 shows the locations of the forests for which the data has been compiled and these extend along approximately 250 km of fault strike. The most southern forest for which data are available is Ohinemaka, near the Paringa River, while the most northern forest is the Rahu forest, near Reefton.
Figure 5.6 - Modes of forest stand establishment from the forests shown in Figure 5.5 from Wells et al. (1998). The date of stand establishment has been estimated with an arbitrary forty-year allowance for colonisation and growth to shoulder (corer) height.

Figure 5.6 shows the data with an arbitrary 40 years added to the tree ages to allow for establishment of the first colonisers. This shows two main modes of forest initiation following massive regional disturbance at around 1400 - 1450 AD and 1600 - 1700 AD. The later peak is much broader than the 1400 - 1450 peak and the authors suggest this may reflect two disturbances relatively close together, basing this on concurrent more detailed work on larger tree populations at the Karangarua River valley. These modes of common forest age can be recognised across a wide range of landforms from steep uplands to much flatter alluvial and moraine deposits and as noted above the authors suggest Alpine Fault earthquakes are a possible explanation of such patterns.

The forest disturbance and regrowth pattern fits the date clusters in the aggradation and landslide records presented in Chapter 4 well. The disturbance initiated between 1600 and 1650 AD matches the Crane Creek event and connects this forest disturbance with the Alpine Fault. The disturbance record for the Crane Creek event has also been further analysed.
by Wells to check for any systematic north-south variation in the timing of disturbance.

Within the limitations of the 25 year age classes the timing appears synchronous across the region, suggesting the Crane Creek event was a single earthquake responsible for forest damage between Ohinemaka Forest and the Rahu Forest (approximately 250 km). However, it is also possible that a progressive series of smaller ruptures over a relatively short time period occurred.

Something similar occurred historically in Turkey with six events of Magnitude 7–8, which migrated westward over 750 km of the North Anatolian Fault in the 28 years between 1939 and 1967 (Barka & Kadinsky-Cade, 1988). This type of fault rupture pattern may not be recognised in this forest age analysis based as it is on 25 year age classes. It is likely cross-matched tree ring chronologies (i.e. chronologies based on communal tree ring width indices which have been adjusted and standardised using the common climatic pattern to reduce counting errors) will be the only method with sufficient precision to check for this possibility.

To investigate the suggestion that the broad forest age peak between 1600 and 1750 is actually two peaks close together, Wells (1998) has recently selected only the regional data able to be plotted in 10 year age classes, and then combined this with age information obtained from his new work in the Karangarua River valley. The forests selected were also restricted to those clearly colonising new young surfaces created either by landsliding or terrace formation. Ages for many of the even aged rata/kamahi stands on the range front noted by Wardle (1980) were also refined for inclusion. For this analysis forest on older surfaces such as moraines and high level terraces was excluded.
Figure 5.7- Ages of new landslide and terrace surfaces from the Karangarua Valley (37 surfaces) and central and south Westland (24 surfaces) in comparison to the date bands for the two most recent ruptures of the Alpine Fault based on the paleoseismic trench data in Chapter 3. Ages of landslide and terraces surfaces have been derived from the age of the oldest tree on each surface plus a species dependant addition for colonisation time (angiosperms: 20 yrs; conifers 36 yrs). Surface age data from Wells, 1998.

In addition the arbitrary 40-year delay added for colonisation time and growth to corer height was refined. New surfaces known to have formed in historical times were investigated and aged separately for the angiosperm species (kamahi and rata) and the five conifer species. The mean for the angiosperms was 20 years (range 14 - 27) and for the conifers was 36 years (range 15 - 60). Figure 5.7 presents the new information.

The improved data quality and delay time estimates allow for clear delineation of two modes of surface formation post 1600 AD. The peak around 1450 AD remains as a single mode. The error on the peaks is much reduced and the
disturbance events responsible for the creation of the new surfaces can be more accurately estimated at 1425 ± 15 yr AD, 1625 ± 15 yr AD, 1715 ± 15 yr AD.

The paleoseismic trenching work outlined in Chapter 3 shows that Alpine Fault earthquakes occurred in the date ranges 1480 – 1645 AD and 1660 – 1840 AD. These date ranges are also shown in Figure 5.7 and it is apparent there is only one age mode within each of the earthquake date bands. Historical examples of earthquakes demonstrate the impact of earthquakes in this type of terrain on forests. I propose these modes effectively date the two most recent Alpine Fault earthquakes and provide a much more accurate guide to the event timing.

What about the alternative of regional climatic disturbance as the explanation of these age modes? In section 4.4 I have already noted the absence of any equivalent regional climatic disturbance in the historical record. However, ignoring for a moment this historical record, on the basis that it is too short, if this type of climatic event had occurred these would supplement the earthquake disturbance record. Any climatic disturbance modes would be additional to the earthquake disturbances. However, as Figure 5.7 clearly shows, there is only one significant mode within each of the trench date ranges.

5.3 FOREST AGE AND THE TERRACE SEQUENCE IN THE KARANGARUA RIVER VALLEY

Figure 5.8 summarises the terrace sequence in the Karangarua River valley upstream of the Alpine Fault. The terrace surfaces, where they are exposed in section by river erosion, appear to be thinly capped straths cut into older gravels, as opposed to the alternative of aggradational terraces. However the exposure is limited and more work clearing moss and overhanging vegetation is required to confirm this is a consistent pattern. I suggest the dominance of degradational terraces as opposed to aggradational terraces in this catchment
reflects the very high "water power" available to remove landslide material and sediments.

The modern river occupies a broad bed cut within the lowest terrace. The river occasionally floods over this terrace and so strictly speaking this is part of the active flood plain. Vegetation growth on the lowest tread is patchy and flood influenced (approximately 10% forest cover and the rest grass).

![Diagram](image)

**Figure 5.8** - Successive terraces in the Karangarua River valley which have been abandoned at time ranges matching the modes of forest disturbance presented in Figure 5.7. An average uplift rate of 11.7 ± 2 mm/year can be derived from an age - height graph of the terraces where the three terraces plot as a straight line.
The few trees that are present on this lowest surface indicate river abandonment of some of the more sheltered parts of the terrace between 1820 and 1875 AD but this may not be reliable given the sparseness of tree cover.

The important observation is that forest conforming to the three regional forest age modes, which we propose are the result of disturbance during Alpine Fault earthquakes, occupies the three successive terraces above modern river flood level (Wells & Yetton, in prep.). The first terrace tread which is consistently above modern river flood level is 80% covered in well-developed forest. The oldest trees on this surface indicate river abandonment in the period 1710 - 1720 AD.

The next highest terrace tread has trees established in the period 1610 - 1620 AD, exactly matching those growing on the nearby McTaggart debris avalanche and fan. In section 4.8 I attribute this to earthquake triggering in the Crane Creek event. Amongst this forest, on both the terrace tread and McTaggart fan, there is also evidence of a younger forest disturbance event with areas of trees established at the same time as those on the youngest terrace (ie 1710 - 1720 AD).

The next highest tread has trees of an age indicating abandonment between 1405 - 1445 AD. Once again patches within this forest show signs of subsequent disturbance and representatives of both the 1620 and 1715 disturbances can also be recognised.

Unfortunately river erosion has removed most areas of higher terrace, and the trees on these are so old that many are no longer the first colonisers. In addition coring of large trees becomes less and less accurate because even the largest corer cannot reach the heart and a less reliable estimate for lost rings must be made. However the oldest trees on the next terrace are around 700 years old and the best estimate for the time of abandonment is 1215 - 1245
AD. Because it is less reliable we have not included this terrace in Figure 5.8 but the height (12m) and age does plot on the expected projection of the straight line of the age - height graph in Figure 5.8.

It is generally accepted that the flights of terraces upstream of the Alpine Fault are the result of episodic uplift by earthquakes along the fault (Adams, 1980). The successive occupation of the new surfaces of uplifted terrace treads by forest conforming to the common forest age modes demonstrates the regional disturbance events are synchronous with new terrace formation. We suggest this new terrace formation is the result of the component of vertical uplift associated with the last three Alpine Fault earthquakes and as such the ages of abandonment closely match the earthquake dates.

By plotting the average terrace height against the time of abandonment an average uplift rate can be obtained (Figure 5.8). The three successive terraces plot as a straight line with a gradient of $11.7 \pm 2$ mm/year. This is higher rate than most estimates of uplift at the fault itself which are of the order of 6 - 7 mm/yr (refer section 1.2.2). However, the terrace heights have been obtained from averages taken over an 8 km stretch of the river upstream of the fault and some additional broader scale uplift east of the fault is likely. This estimate of 12 mm/yr is still well below the estimated rates for uplift at the main divide of 17 mm/yr and 22 mm/yr (Wellman, 1979; Walcott, 1979).

There is no possible alternative climatic explanation for a flight of successively younger terraces in this very young age range. A massive storm, or a series of massive storms, can produce aggradation and floods thereby creating or clearing alluvial surfaces. However, this will not initiate the uplift required to preserve a successive record so each climatic event will work through approximately the same flood and aggradation level range thereby cannibalising the earlier record. Invoking uplift by earthquakes between postulated climatic events to allow preservation is an unnecessary complication.
of the pattern and the abandonment of the surface by the river still dates to the earthquake.

I have noted that the timing of abandonment is the same as the regional forest disturbance modes (refer Figure 5.7). The successive terrace sequence in the Karangarua River valley is further very strong evidence that the regional forest age modes are the result of earthquakes, and have not been caused by some sort of regional climatic aberration.

5.4 EVIDENCE OF EARTHQUAKES FROM TREE RING ANALYSIS

5.4.1 Tree ring analysis of podocarps on the Alpine Fault scarp at Crane Creek

Disturbance of trees which live through an earthquake is preserved in the pattern of tree ring growth and this can be used to precisely date an prehistoric earthquake. Tree ring dating as a method of dating fault movement in New Zealand was tested by Berryman (1980) on the White Creek fault scarp after the 1929 Buller earthquake. Berryman found red beech (*Nothofagus fusca*) growing on the scarp in 1979 with distinct bends in the trunk. The corresponding asymmetry in growth rings as the trees grew reaction wood precisely dated the tilt to the 1929 event. Tree ring methods have been applied to earthquake dating on the San Andreas fault (for example Jacoby *et al.*, 1988) and other faults in wooded areas of the American west coast (Sheppard and Jacoby, 1989; Jacoby, 1997). Recently another historical earthquake in California (the 1978 Stephens Pass earthquake) also demonstrated the reliability of the method (Sheppard and White, 1995).

In general the most profound impacts of earthquakes on trees occur right on the fault scarp. Here the earthquake will frequently result in a period of growth suppression in response to branch or root damage from extreme shaking and/or rupture and mass movement on the scarp. In a few cases tree growth
will actually accelerate following the earthquake where neighbouring trees which previously competed for light or nutrients have been killed. Unfortunately the Alpine fault scarp in Westland is mainly covered in beech and podocarp forest and both these tree are not always reliable in their pattern of tree ring growth. They can sometimes grow “false rings” (ie two per season) and in other cases rings can be so narrow they are missed or sometimes completely absent. However it is generally possible to be accurate in tree ring counts for individual podocarp trees to ± 10 years, which is still much more accurate than radiocarbon methods.

Several old rimu and miro trees (*Dacrydium cupressinum* and *Prumnopitys ferruginea* respectively) grow amongst the younger red beech forest along the Alpine Fault fault scarp at Crane Creek. We have cored seven within 1km of the fault trench site and analysis of the growth rings has been carried out by Wells (pers.comm.1997).

Figure 5.9 presents the ring width count for several of the trees cored. These ring counts are not cross-matched (refer the discussion of cross matching at the start of section 5.5) but in each tree the rings have been counted in from the outermost ring. Some error is inevitable without cross matching because of some individual trees having a few hidden and/or double rings and Wells estimates errors up to ± 10 years with this method.

All trees show a growth suppression (or in two cases an acceleration) at around 1620 AD ± 10 years. This suppression lies within the fault trench date range for the *Crane Creek event* from this site of 1480 – 1645 AD and is a good fit to the likely age of the aggradation surface of 1600 – 1650 AD which buries the dated logs just downstream of the paleoseismic trench at this site (section 3.2.1). It also matches the regional forest disturbance evidence, the terrace age abandonment age in the Karangarua, and the most likely time of the McTaggert debris avalanche.
Figure 5.9 – Examples of tree ring width counts for three trees growing on the fault scarp at Crane Creek which show particularly clear suppressions at around 1620 AD ± 10 years. The y axis is ring width in microns, the x axis is years before 1997. The arrow notes the suppression, while the dashed line is approximately 1620 AD (i.e. 377 years before 1997). A total of seven old podocarp trees growing on the scarp were cored and all show ring width suppression or acceleration at this time however five also show suppressions at around 1580 AD.
However, there is also a change in growth patterns in 5 of the Crane Creek podocarp trees at 1580 AD ± 10 years and more trees must be cored to better define the pattern. The 1620 AD ± 10 years date remains the best estimate because it is present in all trees cored to date and it is the best fit to the regional pattern.

5.4.2 Tree ring analysis of New Zealand cedar at the Karangarua River valley

Wells (1998) has collected tree ring data in the Karangarua River valley from 40 old NZ cedars (*Libocedrus bidwillii*) growing on steep slopes at four sites within 12 km of the Alpine Fault in the Karangarua Valley (Wells et al., in prep.). As noted above seismic shaking commonly causes a marked reduction in the radial growth of trees in the fault scarp area, however, at vulnerable locations in steep landscapes these impacts can also extend to trees as far as 200 km from the fault (Jacoby et al., 1988; Sheppard & Jacoby, 1989; Kitzberger et al., 1995).

Cedars are particularly well suited to detailed tree ring analysis and are much more reliable than podocarps. Cross matching of several hundred *Libocedrus bidwillii* ring width series shows that this species does not produce false rings, and has less than 1% absent rings (Norton, 1986). Unfortunately the cedars do not normally grow at the low altitudes typical of the majority of the fault scarp, so sites some distance away were selected. These sites are on steep slopes (25 – 40 degrees) adjacent to landslide scars with re-established vegetation of an age matching the 1620 and 1720 establishment periods. The trees selected are therefore at sensitive sites and are likely to have been directly affected by seismic shaking but not enough to destroy them.

Wells (1998) defines an abrupt suppression in radial growth as a > 100% decrease in mean ring width between consecutive 10 year means (i.e the mean ring width over one ten year period is less than half the size of the preceding
10 year mean ring width), and identified the starting date of all such suppressions in ring width series by counting in from the outermost ring. The suppression dates are combined into 10-year age classes. Figure 5.10 shows the results of the *Libocedrus bidwillii* suppression analysis.

**Figure 5.10** - Dates of significant growth suppressions in 40 cedars (*Libocedrus bidwillii*) in the Karangarua Valley. The criteria for suppression was a greater than 100% difference in mean growth rate between consecutive 10 year means.

Two periods of unusually common growth suppression stand out over the last 450 years where more than 20% of the trees are affected. These periods (1710 - 1720 AD and 1610 - 1620 AD) match the suggested dates of the Crane Creek and Toaroha River earthquake events deduced from the landslide, trench and forest age evidence presented earlier.
5.5 PRECISE DATING OF THE TOAROHA RIVER EVENT

The method of cross matching tree rings utilises the common climatic record to correlate and adjust ring counts taken from individual trees. Climate pervades stands, forests and regions and often on a 100km or greater scale (Jacoby, 1997). This communal signal allows the tree ring series from individual trees to be pattern matched and generally the most stressful years in a trees life match the best.

Figure 5.11 - Cross-matched chronologies for cedar (Libocedrus Bidwillii) from Norton (1981) for Tarkus Knob and Cream Creek in the Cropp River. Also shown are the plots for five representative individual trees. Note this location is approximately 8 km from the Alpine Fault trace in the expected epicentral area of maximum shaking. The nearest trench sites in this study are also shown.
This pattern matching corrects for missing or false rings in individual trees and thereby reduces possible errors allowing the cross-matched growth rings to be assigned to the correct and exact calendar year.

Cross-matched cedar tree ring chronologies already exist for the Cropp Valley near the Hokitika River (Norton, 1981). Originally intended for dendroclimatic purposes these chronologies are each based on 19 trees. Wells (Wells et al., in press) has re-examined these in light of the trenching results in this investigation to look for unusual fluctuations in growth for the period between 1660 and 1840, and in particular the period around 1720 AD as suggested by the forest age and Karangarua data.

The Cropp River sites are 9 km from the Alpine Fault and 25 km from the Toaroha River trench locations. Figure 5.11 shows the location details with respect to the Toaroha trench sites. Separate chronologies have been constructed by Norton (1981) for Tarkus knob, a steep south facing slope underlain by schist at shallow depths, and Cream Creek which is a much more gently sloping alluvial surface. The dip of the fault and the likely depth of rupture suggest this area would have been within the zone of maximum earthquake shaking in the Toaroha River event.

The cross-matched chronologies and a selection of ring width counts from individual trees is presented in Figure 5.11, along with the general location details. Fortunately these chronologies commence in 1660 AD, the earliest possible date for the Toaroha River event based on the paleoseismic trenches, and extend all the way through to 1980, the date of Norton's fieldwork. The paleoseismic trench evidence outlined in Chapter 2 consistently demonstrates that within the period between 1660 AD and 1840 AD (the earliest explorers) an earthquake rupture occurred on the Alpine Fault. The trees in this location must have experienced the extremely strong earthquake shaking and the response to this must be reflected somewhere in their growth patterns.
Figure 5.11 shows that abrupt marked growth declines began in both chronologies at 1717 AD and were followed by several years of below average growth. As is expected with earthquake shaking, the steeper Tarkus Knob site shows the greatest impact and this event was the largest growth interruption in 400 years. In four trees the impact began in 1717 with a small suppression but was also followed by a sharp suppression the following year. This pattern suggests that growth was disrupted during the 1717 growing season.

This suppression coincides with the suppression shown in Figure 5.10 for the Karangarua Valley and most importantly the regional forest age mode. It also matches the terrace abandonment in the Karangarua River valley. Six other existing cross-matched chronologies (Norton, 1983a & 1983b) were also examined, extending from the Paringa River to the Takahe River near Lake Te Anau, all between 25 and 75 km from the Alpine Fault (Figure 5.12). These chronologies are mainly from silver beech (Nothofagus menziesii) with one from mountain beech (Nothofagus solandri) and also commence in 1660 AD or earlier and extend to at least 1980. All chronologies show a marked growth anomaly at this same time.

The paleoseismic trenching presented in Chapter 2 indicates the northern limit of the Toaroha River event was between Haupiri and Crane Creek. Two cedar chronologies within 10 km of the fault are available for the area further north from sites near the Ahaura River and Rahu Saddle (Limin, 1996). Neither show a growth anomaly at this time, and in fact the Ahaura chronology indicates a period of above average growth. Five other chronologies are available from the central eastern South Island (Limin, 1996 [four L. bidwillii chronologies]; Norton, 1983b [one N. solandri]). None of these show a suppression at this time.
Figure 5.12- All the available cross-matched tree chronologies along the strike of the Alpine Fault from Lake Te Anau to the Rahu Saddle. The abrupt growth suppression in 1717 AD coincides with the northern limit of the post 1660 AD rupture at the Haupiri River determined by trenching the fault. The y-axis on all the graphs is the ring-width index and x-axis is calendar years AD.
Exceptional climatic events such as heavy snowfall, or unseasonal frost, could cause long-lasting growth suppression in trees at a single site. Historically severe storms in the region have only had a local impact (Benn, 1990). The impacts may extend over many tens of kilometres, but for this type of climatic event to extend for over 300 km from Lake Te Anau to the Cropp River, and yet be suddenly absent in chronologies from north of the Haupiri River, is extremely unlikely. There is no topographic or other barrier to constrain storms in this area. The absence from eastern South Island chronologies also indicates this was not a national climatic or atmospheric phenomena.

This evidence indicates the best estimate of the timing of the most recent Alpine Fault earthquake is the growing season of 1717 AD

5.6 SUMMARY AND CONCLUSIONS

The forest and tree record in Westland has been examined to better estimate the most likely timing of the last two Alpine Fault earthquakes. The occurrence of these earthquakes is proven by the paleoseismic trenching work presented in Chapter 3 and is supported by evidence in the landslide and aggradation terrace record (Chapter 4).

Regional forest age modes can be identified at 1425 ± 15 yrs AD, 1625 ± 15 yrs AD and 1715 ± 15 yrs AD which are consistent with the landslide and terrace record and are a good fit to the paleoseismic trench data. Only one forest disturbance mode occurs within each trenching date range which is strong evidence that climatic impacts are not responsible for these regional disturbances. Further evidence that earthquakes are responsible for the regional disturbances comes from the match between the timing of these regional disturbance events and the ages of a flight of terraces created by coseismic uplift upstream of the Alpine Fault in the Karangarua Valley.
Potentially trees which have survived these disturbances preserve tree ring fluctuations which more accurately constrain the timing. The evidence suggests the Crane Creek event occurred at 1620 ± 10 yr AD and extended from at least the Paringa River to the Rahu Saddle. The more recent Toaroha River event appears to have occurred in 1717 AD and can be traced from Lake Te Anau to the northern limit near the Haupiri River previously identified by trenching.

The assignment of the most recent Alpine Fault earthquake (the Toaroha River event) to the year 1717 AD:

- is a very good fit to the consistent evidence from the Kokatahi, Toaroha and Haupiri paleoseismic trenches that the most recent Alpine Fault rupture occurred after 1660 AD and before 1840 AD, and most likely in the period 1700 - 1750 AD. Wood, leaves and twigs from 1717 AD will sometimes return a conventional radiocarbon age of around 200 yrs BP but frequently will appear as “modern”. This explains some of the “modern” dates obtained in the Toaroha trenches.

- coincides with the timing estimate of between 1650 AD and 1725 AD (Cooper & Norris, 1990) for the last event in Fiordland, based on pits excavated in sag ponds near the fault and broken beech trees.

- is the most logical explanation of terrace abandonment at this same time in the Karangarua River valley.

- is the best explanation and an excellent fit to the most recent regional forest age mode (1720 ± 10 years AD).

- and the most compelling independent evidence confirming that this is the earthquake signature is the match between the northern limit of this growth
suppression with the rupture limit at the Haupiri River independently
determined by the paleoseismic trenching.

All the evidence is consistent, and indicates that the most recent Alpine Fault
earthquake (the *Toaroha River event*) occurred in the growing season of 1717
AD and the shaking impacts were synchronous from Lake Te Anau to the
Haupiri River, a distance of approximately 400 km.
Chapter 6

MAGNITUDE ESTIMATES AND A SYNTHESIS OF THE ALPINE FAULT EARTHQUAKE CHRONOLOGY

6.1 INTRODUCTION

This chapter commences with a review of the methods most commonly used for assessing the magnitude of paleoearthquakes. In general the most reliable assessment comes from the correlation between historical earthquake magnitude and rupture length, and this method is applied here to estimate the magnitude of the most recent three Alpine Fault earthquakes for which data is available. An overall chronology of Alpine Fault earthquakes over the past 2000 years is proposed and a comparison is made between this earthquake chronology and various earlier inferences.

6.2 ASSESSING THE MAGNITUDE OF PALEOEARTHQUAKES

Earthquake magnitude is based on measurement of the maximum amplitude of earthquake vibration on a seismograph 100 km from the epicentre and is in principle a measure of the earthquake energy release. However each type of seismograph has a characteristic period and in effect records the maximum amplitude at a particular frequency. The earliest estimates of magnitude in New Zealand and USA came from the Wood-Anderson seismometer (period 0.8 sec) which is useful for earthquakes within 1000 km of the instrument. Magnitudes originally measured in this way are referred to as local magnitude or “Richter magnitude” (M_L). Subsequently longer period instruments came into wide use allowing the effective measurement of magnitudes at greater distances from the epicentre, however the magnitudes calculated from these
instruments resulted in small differences. To distinguish the scales these magnitudes are designated "surface wave magnitude" (M_s). The difference is not significant for large earthquakes, but below approximately M_s 5 the values obtained from distant records may be as much as half a magnitude unit too low.

Unfortunately none of these scales satisfactorily measures the largest possible earthquakes because M_s remains a measure of the seismic wave amplitude at a specific period (typically 18 – 22 sec). In large earthquakes these periods become saturated and no longer reliably record large scale faulting characteristics (Hanks & Kanamori, 1979). Furthermore significant variations in M_s may occur between distant stations related to azimuth, station distance, instrument sensitivity, and crustal structure (Panza et al., 1989). In this way traditional magnitude scales are limited by both the frequency response of the earth and the response of the recording seismograph.

A more direct link between earthquake size and fault rupture parameters is seismic moment, M_o = µΔA, where µ = the shear modulus (usually taken as 3 x 10^{11} dyne/cm^2 for crustal faults [Hanks & Kanamori, 1979]; Δ is the average displacement across the fault surface; and A is the fault surface area that ruptures. In turn M_o is directly related to moment magnitude (designated M or M_w) as follows M = (2/3)(log M_o – 16), where M_o is in units of dyne-cm (Hanks & Kanamori, 1979).

Moment magnitude (M) is the most reliable measure of energy release in an earthquake and has generally been adopted in the most recent analyses of the relationships between magnitude, surface rupture length, and single event slip. In paleoseismology these empirical regression relationships provide the only means to estimate earthquake magnitude.

The length of surface rupture has been used to estimate the magnitude of the earthquake responsible based on various regression relationships developed over the last 30 years (Bonilla & Buchanan, 1970; Mark & Bonilla, 1977;
Slemmons, 1982; Bonilla et al., 1984; Khromovskikh, 1989; Wells & Coppersmith, 1994; Anderson et al., 1996). These models have improved progressively as more and more data becomes available on the magnitude and rupture length of historic earthquakes from around the world.

In general rupture length has been shown to provide the most reliable correlation with magnitude and the standard deviation in this relationship is considerably less than the correlation between magnitude and either single event slip (displacement) or magnitude and average single event slip (average displacement). For example Wells & Coppersmith 1994 obtained a standard deviation for the rupture length correlation of 0.22M, in contrast to standard deviations of 0.4M and 0.33M respectively for the other two correlation's.

The most recent refinement of the rupture length correlation method involves the inclusion of average long-term slip rate in selecting the appropriate regression line (Anderson et al., 1996). The rationale behind this modification is the observation by the authors that the highest moment magnitudes for a given rupture length tend to be produced by faults with relatively low long term slip rates. This may reflect the more irregular nature of the fault plane for faults with longer return times (effectively a higher overall residual friction) but also the concept that as the time from the last earthquake increases, geological processes strengthen the fault (Kanamori and Allen, 1986; Scholz et al., 1986). Greater energy is therefore required to rupture a given length of fault plane and a larger stress drop and higher magnitude earthquake is the result. The converse is likely for the Alpine Fault, given relatively short return times and a very large cumulative geologic offset (refer section 1.2.4).

For the Alpine Fault estimates below both Anderson et al. (1996) and Wells and Coppersmith (1994) give very similar estimates and both are used in this study.
One reservation in adopting any of these international correlation’s is the small number of New Zealand examples in the data base (only one in Anderson et al., 1996; and four in Wells and Coppersmith, 1994). Where sufficient information is known the New Zealand examples all plot well outside the 95 % confidence levels, and generally on the high side (i.e. greater magnitudes have resulted from a given rupture length). This fact, in conjunction with the use of only minimum rupture lengths for the Alpine Fault, suggests any magnitude estimates should be applied with caution and may if anything be underestimates of the real magnitudes.

It is interesting to note that in the 244 historic earthquakes in the Wells & Coppersmith (1994) database only one, the San Francisco earthquake of 1906, has a rupture length longer than the 375 km minimum suggested below for the Toaroha River event. Table 6.1 summarises the seven historic earthquakes for which magnitudes and rupture lengths are known. In the oldest of these, the 1857 Fort Tejon example, the magnitude should be viewed as an estimate only.

<table>
<thead>
<tr>
<th>Magnitude (Ms)</th>
<th>Rupture length</th>
<th>Country</th>
<th>Earthquake Name</th>
<th>Year</th>
</tr>
</thead>
<tbody>
<tr>
<td>8.5</td>
<td>220 km</td>
<td>China</td>
<td>Kansu</td>
<td>1920</td>
</tr>
<tr>
<td>8.3</td>
<td>297 km</td>
<td>USA</td>
<td>Fort Tejon</td>
<td>1857</td>
</tr>
<tr>
<td>7.9</td>
<td>236 km</td>
<td>Mongolia</td>
<td>Gobi – Altai</td>
<td>1957</td>
</tr>
<tr>
<td>7.9</td>
<td>432 km</td>
<td>USA</td>
<td>San Francisco</td>
<td>1906</td>
</tr>
<tr>
<td>7.8</td>
<td>360 km</td>
<td>Turkey</td>
<td>Erzihcan</td>
<td>1939</td>
</tr>
<tr>
<td>7.5</td>
<td>280 km</td>
<td>Turkey</td>
<td>Kastamonu</td>
<td>1943</td>
</tr>
<tr>
<td>7.5</td>
<td>235 km</td>
<td>Guatemala</td>
<td>Motagua</td>
<td>1976</td>
</tr>
</tbody>
</table>

Table 6.1: The magnitudes of the few historic earthquakes for which surface rupture lengths are known and which exceed 200 km. Data from Wells & Coppersmith (1994).
The problems with correlating rupture length and magnitude are immediately obvious in this table which shows that the highest historic magnitude was associated with the shortest rupture length (i.e. the 1920 Kansu earthquake in China). Similarly, by far the longest rupture (the 1906 San Francisco earthquake in the USA) resulted in a magnitude which is only in the middle range of the case studies presented in Table 6.1. Regardless of this lack of precision, rupture length is a better indicator of magnitude than single event slip for the reasons outlined below.

*Estimates based on single event slip*

Single event slip is expressed by the horizontal and vertical components of movement (or offset) at the surface fault trace in a particular earthquake. There are major problems in determining this parameter with any reliability. The difficulties derive in two quite different ways, the first relating the nature of the process and the other to the limitations in field determination.

In historical earthquakes many observers have noted it is very rare for a surface offset to remain constant along a fault trace. Most earthquakes seem to have sub-regions that may have greater slip (Lay et al., 1982; Beck & Ruff, 1987; Segall & Du, 1993; Wald & Heaton, 1994; McCalpin, 1996). Horizontal or vertical slip variations of 5 or even 10 to 1 are not uncommon in these historical observations. This phenomena can be replicated in physical models (Das & Aki, 1977; Mikumo & Miyatake, 1978) and appears to be a natural consequence of structural barriers along the fault plane.

A good example of slip variation in a large strike slip earthquake is provided by the 1906 San Francisco earthquake. Thatcher et al. (1997) reviewed the variation in single event slip as measured at the fault trace soon after the earthquake and compared this to the slip variation as determined by the reoccupation of preexisting geodetic survey points. Figure 6.1 summarises their findings.
Figure 6.1: Variation in single event slip along the 470 km of surface rupture trace in the 1906 San Francisco earthquake after Thatcher et al. 1997. The lines and circles show variations in actual surface horizontal offsets while the black histogram shows variations in slip calculated from resurvey of the geodetic network.

Surface offsets at the fault scarp varied from 0.5 – 6 metres with rapid variations over relatively short distances. The survey networks also demonstrated substantial variations, between 2.3m and 8.6m and demonstrate the variations measured at the fault scarp are real differences in surface slip on the fault plane, and not the result of short scarp profiles which may not measure subtle far field components of deformation.

The 1906 earthquake occurred on a relatively mature strike slip fault (c.f. Wesnousky, 1989). Like the Alpine Fault this section of the San Andreas Fault has a high slip rate (approximately 30mm/yr) and a relatively large total cumulative offset (approximately 250 km). Despite this the single event slip varied unpredictably by approximately 300%. This type of variation is also likely to have occurred in past Alpine Fault earthquakes.
The concept of average single event slip has been introduced in an attempt to better utilise slip measurements for magnitude estimates (Kanamori, 1977; Thatcher & Bonilla, 1989; Wells & Coppersmith, 1994). This has been found to be on average 32% lower than the maximum single event slip where sufficient data is available to make a meaningful comparison (Wells & Coppersmith, 1994). However there are no criteria to establish the length of fault which must be averaged to obtain a reliable value. In the case of the Alpine Fault, where at best there may be on average 2 locations per 100 km of fault length on which to assess single event slip, the limitations drastically restrict the use of this method.

The Alpine Fault may have another unique source of intrinsic single event slip variation and this relates to variations in trace strike and complex in detail segmentation. Strike varies by as much as 20 degrees along traces and as a consequence trigonometry indicates horizontal fault parallel slip will vary by approximately 15-25% between the most extreme oblique slip traces (where it is a minimum) and the most extreme strike slip traces (maximum).

There are also very real field difficulties in measuring single event slip in the central area of the Alpine Fault. At three of the five trench locations there are multiple traces (Toaroha River, Kokatahi River, Ahaura River). Where two adjacent traces were both trenched (at the Kokatahi River) the results imply both traces moved simultaneously in the last event. There is no way of knowing how the slip was apportioned between them. At Crane Creek and Haupiri River the fault is represented by a single trace, but there have been multiple rupture events, and no clear single event offsets can be recognised.
6.3 ALPINE FAULT EARTHQUAKES IN THE LAST 750 YEARS

6.3.1 The Toaroha River Event

This is the best defined of the recent earthquakes due primarily to the excellent tree ring chronologies available for the last 300 years. We summarise the details in Table 6.2 below.

<table>
<thead>
<tr>
<th>Evidence</th>
<th>Trenching</th>
<th>Landslides and terrace age</th>
<th>Forest Age</th>
<th>Tree ring chronologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Estimate of Timing</td>
<td>Post 1665 AD, probably 1700 - 1750 AD</td>
<td>Post 1660 AD</td>
<td>1715 AD ± 15 yr</td>
<td>1717 AD</td>
</tr>
<tr>
<td>Rupture Length</td>
<td>A minimum length of 375 km is suggested by trenching and the landslide record but a possible length of 450 km is suggested by the tree ring chronologies.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Moment Magnitude Estimate</td>
<td>For 375 km: $M = 8.05 \pm 0.15$ (Wells &amp; Coppersmith, 1994) $M = 8.0 \pm 0.26$ (Anderson et al., 1996)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>For 450 km: $M = 8.15 \pm 0.2$ (Wells &amp; Coppersmith, 1994). $M = 8.05 \pm 0.26$ (Anderson et al., 1996)</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 6.2: Summary of key features of the Toaroha River event
**Rupture length**

A rupture length of 375 km is the minimum which the direct fault data indicates. It is based on the northern limit as defined by the trenching at the Haupiri River and Crane Creek. The southern limit is taken as the John O'Groats site of Cooper & Norris (1990) in Fiordland. However the tree ring impacts, which extend further south to Lake Te Anau, suggest the rupture may have extended along the section of fault offshore of the Fiordland coast to at least opposite the lake. This is a total length of approximately 450 km.

Barnes (pers. comm. 1999) considers the offshore continuation of the Alpine Fault southwest of Milford continues without any apparent change in fault character for approximately 70-90 km. At this point, designated the Hawley transfer by Barnes, the fault overlaps with fault strands approximately 5km to the west forming a stepover. He infers the Alpine Fault between Haast and the Hawley transfer is a single rupture segment. However until more detailed forest work is undertaken in the Fiordland area, which looks in more detail at the geographic variation in the impacts of 1717 Toaroha River event, it is most appropriate to assume a southern rupture limit at around Milford. In Table 6.2 we note that the extra rupture length could result in a magnitude increase from around M 8.05 ± 0.15 to M 8.15 ± 0.2.

Similarly, there may ultimately be a case for extending the northern rupture length to include a section of the Hope Fault. Cowan & McGlone (1991) estimate the timing of the penultimate earthquake on the Hope River segment at 1728 AD but with an error of ± 53 - 93 yrs (sic.). The Hope Fault may be triggered by an Alpine Fault earthquake after a delay of many years but it is also possible the Alpine Fault and Hope Faults rupture together in a pattern similar to the Imperial Valley earthquake in California of 1979 (Archuleta, 1982). Figure 6.1 shows the surface rupture pattern for this event.
Figure 6.2: Surface rupture pattern in the Imperial Valley earthquake of 1979 in California (Archuleta, 1982). The rupture splay into the adjacent Brawley fault zone may be similar to the pattern of rupture at the various Marlborough fault junctions with the Alpine Fault.
The continuation of the rupture past the branch to the Brawley Fault zone may be similar to the north section of the Alpine Fault rupturing past the Hope Fault and on up to the Haupiri River.

The last earthquake rupture crossed the Alpine Fault "segment" boundaries proposed by Bull (1996) and Berryman et al. (1992). These boundaries had been tentatively proposed on geomorphic and structural grounds. While the division into geographic sections adopted by Berryman et al., 1992 may be still be useful for location description, it appears there may not be persistent segmentation of the Alpine Fault in the normal sense.

6.3.2 The Crane Creek event

Table 6.3 summarises the known information regards the Crane Creek event. The length of rupture is very much a minimum and may seriously underestimate the magnitude. The northern minimum is the Rahu Saddle, based on the presence of the Crane Creek event in the forest age record in this area of forest. The Alpine Fault extends another 85 km to Lake Rotoiti and the Tophouse Saddle and nothing is currently known of the earthquake history in this area of the north section.

The southern limit is arbitrarily assumed to be just north of the Haast River because the currently available forest and tree ring data does not extend any further south than the Paringa River. This limit north of Haast is also based on the observations of Berryman et al.(1992) who suggest Haast may be a fault segment boundary, and who note evidence for just one earthquake since a radiocarbon date of 860 ± 55 yrs BP at the Pyke River in Fiordland. Unfortunately the constraints and details regards this date are not presented in the paper.
Table 6.3: Summary of key features of the Crane Creek event

<table>
<thead>
<tr>
<th></th>
<th>Trenching</th>
<th>Landslide and terrace ages</th>
<th>Forest Age</th>
<th>Tree ring chronologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Estimate of Timing</td>
<td>1480 - 1645 AD</td>
<td>1488 - 1640 AD</td>
<td>1625 AD</td>
<td>1620 AD ± 10 years</td>
</tr>
<tr>
<td>Rupture Length</td>
<td>A minimum length of 200 km is suggested by trenching and the landslide record but a minimum of 250 km is indicated by forest age</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Moment Magnitude Estimate</td>
<td>M &gt; 7.8 (both Wells &amp; Coppersmith, 1994 and Anderson et al., 1996)</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

It is not easy to compare the magnitudes of the Toaroha River and Crane Creek events, but the impression from both the forest age record (refer Figure 5.7) and the landslide and aggradation terrace dates (refer Figure 4.8), is that the Crane Creek event was the larger of the two. This is also suggested by the tree ring chronologies in the Karangarua (Figure 5.9). However this apparent difference could also be the result of the time of the year in which each earthquake occurred. While the tree ring chronologies suggest the Toaroha River event occurred during the growing season, and most likely in late summer, the Crane Creek event may have coincided with a much wetter period. This could make a very large difference to the frequency of landslides and rock avalanches triggered by an earthquake in this terrain and affect the corresponding impact on both the forests and rivers.
It is possible that the two events were separately biased to the south and north with overlap in the central section. The *Toaroha River event* may have been triggered in the Fiordland segment and then extended with decreasing severity up to Lake Haupiri. The *Crane Creek event* may have been the slightly earlier, but equivalent large earthquake for the north section, but which also extended down in to the central section. In relation to this it is interesting to note that a series of landslide dams with an age around 1650 AD have been recognised in the area north of Springs Junction (Perrin and Hancox, 1992). These may also be related to the *Crane Creek event*, but numerous other earthquake sources also occur in this region, and more paleoseismic investigation is required to test this hypothesis.

### 6.3.3 The Geologists Creek Event

The *Geologists Creek event* has not been recognised in trenches to date and is the least definite and most poorly defined of the three most recent events. Evidence for this event comes from the landslide and aggradation terrace record and the forest age patterns. The name “Geologists Creek” derives from a large rock avalanche deposit from the fault scarp in Geologists Creek, at the eastern end of Lake Kaniere, first identified by Trevor Chinn (pers. comm, 1996). This deposit has been radiocarbon dated by Chinn at 550 ± 50 yrs BP (NZ 6471) and this is the best radiocarbon estimate currently available of the precise timing of this inferred earthquake event. Fortunately the radiocarbon time period from 380 - 575 yrs BP corresponds to a relatively steep and unambiguous section of the calendric calibration curve so radiocarbon dates provide much narrower calendric time constraints for this event (see Table 6.4).

Currently work is continuing by others on a series of recently felled tilted rimu trees (*Dacrydium cupressinum*) growing on the fault scarp at the Waitaha River (Wright, pers. comm., 1999; Wright, 1996). Other trees further from the scarp have remained straight. The logs have been removed by Timberlands for timber milling but discs cut from the tilted stumps are being analysed to
determine the exact time of the tilt episode. This approach is similar to that adopted by Berryman (1980) who was able to demonstrate consistent tilting in *Nothofagus fusca* trees growing on the White Creek fault scarp in the 1929 Buller earthquake.

<table>
<thead>
<tr>
<th>Estimate of Timing</th>
<th>Trenching</th>
<th>Landslide and terrace record</th>
<th>Forest Age</th>
<th>Tree ring chronologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Not yet recognised</td>
<td>1420 - 1450 AD</td>
<td>1425 AD ± 15 years</td>
<td>Current chronologies not old enough</td>
<td></td>
</tr>
</tbody>
</table>

Rupture Length: A minimum length of 250 km is suggested from the forest age data.

Moment Magnitude Estimate: \( M > 7.8 \) (both Wells & Coppersmith, 1994 and Anderson et al., 1996)

**Table 6.4: Summary of key features of the Geologists Creek event**

To date the best estimate by Wright for the age of tilt is 1450 AD, which is a good match to the *Geologists Creek event* as defined in the landslide and forest disturbance record. This preliminary date estimate of 1450 AD may yet extend back as more rings are recognised, and although it is close to the true age, it is effectively a minimum.

The estimate of rupture length is based on the extent of the current forest age record described in Chapter 4 which has been collated from forests between the Ohinemaka Forest, near the Paringa River, up and including the Rahu Forest near Reefton (Rahu Saddle). This is once again a minimum estimate.
All the available constraints on timing and rupture length are combined below in a space-time diagram of the most recent three events (Figure 6.3).

Figure 6.3: Space-time diagram for the most recent three Alpine Fault earthquakes. Note that only the Toaroha event of 1717 AD has a clearly defined rupture limit at the north end. All the other limits are effectively minimums.
6.4 POSSIBLE EARLIER ALPINE FAULT EARTHQUAKES

The evidence for older events becomes less and less conclusive but some patterns do appear to emerge. Most dates come from landslide and aggradation terrace ages, although the Muriel Creek event has some support from forest age in the Karangarua River. In addition new work at the Waitaha River site provides some dates for the inferred coseismic creation of sag ponds along the trace at this location (Wright 1997; Wright et al. 1998; Wright 1998). Table 6.5 below summarises the supporting evidence for each postulated earthquake event.

<table>
<thead>
<tr>
<th>Possible event</th>
<th>Supporting evidence</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Muriel Creek Event</strong></td>
<td>A 6m high terrace in Muriel Creek (GR J33/502032) is the next highest terrace to a 4m terrace dated at 540 ± 60 yrs BP (Wk 4439) which is the inferred Geologists Creek terrace. The 6m terrace contains young wood dated at 810 ± 60 yrs BP (Wk 4339, this thesis). The next most abundant forest age mode prior 1440 AD in the Copland and Karangarua Rivers is also around 1200 AD (Wells, pers. comm. 1997; Wells 1998). Wright (1998) has a date for a possible coseismic sediment pulse in the southern sag pond at the Waitaha River site of 910 ± 50 yrs BP (Wk 5424) which matches this time period. Bull (1996) also proposes an Alpine Fault event at around this time (1226 ± 10 yrs) based on lichenometry.</td>
</tr>
<tr>
<td><strong>Roundtop event</strong></td>
<td>The large Roundtop debris avalanche near the Toaroha River (Wright, submitted; Wright, 1996) has been dated in this thesis from radiocarbon dates on two samples of wood 95 ± 5 rings apart. These were taken from the same 200 mm diameter kahikatea (Dacrycarpus dacrydioides) exposed in the base of the avalanche debris in Cunningham Creek (GR J33/535095). The dates are Wk 4877, 1100 ± 45 yrs BP and Wk 4914, 1300 ± 40 yrs BP. This method reduces the error by defining the overlap in ages which corresponds to a</td>
</tr>
</tbody>
</table>
Continued from previous page

calendric age of 940 AD ± 50 yrs. In addition Wright (1998) has obtained a radiocarbon date of 1160 ± 50 yrs BP (Wk 5426) for the inferred coseismic formation of the central sag pond at the Waitaha River site which is compatible with this date.

<table>
<thead>
<tr>
<th>Roundtop event</th>
<th>940 AD ± 50 yrs</th>
</tr>
</thead>
<tbody>
<tr>
<td>A large debris flow on the right of the Karangarua River (GR H36/530311) has been dated at 1210 ± 50 yrs BP (Wk 5268, this thesis) from the sapwood of a buried ribbonwood branch (<em>Hoheria lyalli</em>). Adams (1980) quotes an early radiocarbon date collected by P. Suggate of 960 ± 150 yrs BP (NZ 9) from a log in a terrace in the Wanganui River which may also be related to this inferred event, but this is a very large error. In addition Adams quotes a date of 1020 ± 50 yrs BP (NZ 1155) collected by P. Wardle from a terrace near the Waiho River.</td>
<td></td>
</tr>
</tbody>
</table>

Bull (1996) notes the large number of rock avalanche deposits in the Southern Alps dated by weathering rinds which approximate to this date (a total of 8, Whitehouse & Griffiths, 1983). He presents an updated and recalibrated data set which places the highest probability for the combined 8 events at 963 ± 1 AD.

Bull (1996) also proposes an Alpine Fault event at 967 ± 10 AD based on lichenometry.

<table>
<thead>
<tr>
<th>Waitaha River Event</th>
<th>595 AD ± 60 yrs</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wright (1998) has a radiocarbon date from the base of the southern sag pond which he attributes to coseismic formation of 1460 ± 60 yrs BP (Wk 5425). Adams (1980) quotes a radiocarbon age of 1560 ± 55 yrs BP (NZ 1293) collected by P. Wardle from buried stumps beneath a young aggradation terrace at Welcome Flat in the Copland. In the Cascade Valley, near Jackson Bay, a large rock slide from Mt Delta buries logs dated at 1540 ± 80 yrs BP (NZ 4626C) by P. Wardle and is quoted by Adams (1980).</td>
<td></td>
</tr>
</tbody>
</table>
Cooper & Norris (1990) obtained a date of 1980 ± 60 yrs BP (Wk 1478) from the base of a sag pond at John O’Groats in Fiordland. In Steep Creek (GR K33/ 719226), near the Wainihinihi River, a large debris flow at the fault scarp contains logs dated at 2100 ± 60 yrs BP (Wk 4004, this thesis). In the Karangarua River, at the junction with the Copland, a large rockfall yields a date of 2150 ± 60 yrs BP (Wk 5269, this thesis). These three dates can be reconciled within the 95% significance level and the averaged radiocarbon date yields the calendric range noted opposite. Adams (1980) quotes a date from the Okuru River, near Haast, of 2210 ± 90 yrs BP (NZ 1370A, Cooper & Bishop, 1979) which the Ward & Wilson test (Ward & Wilson, 1979) indicates is statistically different, despite the relatively large error associated with this date.

Table 6.5 - Summary of inferred Alpine Fault earthquakes prior to 1250 AD.

This may not be all the events over this period, but all of these events have more paleoseismic evidence than just the apparent coincidence of landslides and terrace ages, and three of the four have direct fault evidence (the exception to date is the *Muriel Creek event*).

In addition to the dates noted in Table 6.5, there are two more dates for this time period of which we are aware. These are 1820 ± 60 yrs BP (Wk 4003, this thesis) for a debris flow near the Steep Creek site and a date noted by Adams (1980) of 1740 ± 60 yrs BP (NZ 296) for wood from a terrace on the Waiho River. Although both these dates overlap, and both fall into the relatively large time gap between the inferred John O’Groats and Waitaha events, I consider this is insufficient evidence to postulate an earthquake at this time.
6.5 COMPARISON WITH EARLIER ESTIMATES OF THE DATES OF ALPINE FAULT EARTHQUAKES.

6.5.1 Comparison with Adams (1980)

Figure 1.8 (refer Chapter one) from Adams (1980) presents the ten dates over the last 2,000 years on which Adams based his inferences. Within this very limited dataset, two or more dates which coincide within the one sigma radiocarbon error are considered by Adams to be sufficient evidence to infer an earthquake. The series of inferred earthquakes appear to be at approximately 500 year intervals with the most recent event around 550 years ago (the Geologists Creek event).

However, Adams does recognise that the record may still be incomplete and that future dating may reveal intermediate age "earthquakes", and notes a single date from the Crane Creek event which he discarded at that time because there was no match (NZ 1292, 360 ± 60 yrs BP, collected by Peter Wardle from McTaggerts Creek).

With hindsight there were insufficient dates in the original data set for Adams to make any reliable inference regards earthquake timing. All of the inferences regarding earthquakes are based on indirect indicators (landslides and aggradation terraces) which can also result from processes other than earthquake. Furthermore, the youngest event at 550 years is supported by just two dates which coincide, and one minimum date (i.e. younger than 680 ± 40 yrs BP from a faulted fan at the Waitaha River). The event approximately one thousand years ago (the Roundtop event) is supported by one normal resolution date (NZ 1155; 1,020 ± 42 yrs BP) and another with a very large error (NZ 9; 960 ± 150 yrs BP).

Although there are three dates in support of the event 1550 years ago, only two coincide within the normal error (i.e. NZ 1293; 1560 ± 42 yrs BP and NZ
4626C; 1540 ± 80 yrs BP). An event 2200 years ago is postulated based on only one date and the previously inferred recurrence interval of 500 years.

However, as has been demonstrated in this thesis – albeit based on a much more extensive data - it appears that in many cases Adams (1980) is broadly correct in inferring earthquake events at around these times but some important recent and intermediate events have been omitted. Furthermore, the match between the aggradation terrace and landslide ages and the two most recent earthquakes recognised by paleoseismic trenching, does indicate the usefulness of the method as an indirect paleoseismic indicator.

6.5.2 Comparison with Cooper & Norris (1990)

There is good agreement between the date of the last earthquake in Fiordland as inferred by Cooper & Norris (1990) and the date for the Toaroha River event proposed in this study. We have also cited their much earlier date of 1980 ± 60 yrs BP from the second pit as direct evidence of an earthquake at around this time, and this is compatible with two landslide dates obtained in this work.

6.5.3 Comparison with Bull (1996)

Bull (1996) inferred events at 1748 AD, 1489 AD, 1226 AD, and 960 AD (all ± 10 yrs) with a remarkably constant 260 ± 15 year recurrence interval.

The dates of the most recent two earthquakes are not compatible with three of the four independent lines of evidence on which we base our earthquake chronology for the last 750 years. The broad radiocarbon date bands from the trenches do not highlight the discrepancy. Bull's inferred 1748 event could be the post 1660 Toaroha event. The Crane Creek event (1480 - 1645 AD) could be his 1489 AD earthquake. However, if this is the case, then there is no equivalent of the Geologists Creek event. The next two events coincide well i.e. Muriel Creek (his 1226 AD) and Roundtop (960 AD.).
The main miss-matches are in the comparison with the forest age record, the landslide and terrace date record, and the tree ring chronologies. The sequence of successive terraces in the Karangarua is also a poor fit. None of these can be reconciled with the Bull (1996) dates for the youngest two earthquake events.

This may in part be because the lichen method is another indirect method with no unique connection to a single fault. It is essentially a variation on the use of landslides as paleoseismic indicators and because of this it is open to the same criticism. Processes other than an earthquake regularly trigger rockfalls, and even if an earthquake is the trigger, it is very difficult to assign a particular fault as being the fault responsible.

In some of the more arid regions near the Hope Fault, Bull has been able to demonstrate the existence of a belt of lichen dated rock outcrop damage, in a relatively narrow line along the fault responsible for the earthquake (Bull, 1997). This is a much more reasonable method but unfortunately cannot be utilised for the Alpine Fault where the high humidity and forest cover combine to prevent the growth of suitable lichens.

This difficulty has lead to a considerable eastern bias in the sites that Bull is obliged to use (Figure 6.4). The closest site of Bull's sites to the Alpine Fault is the Otira rock slide at Arthurs Pass. This is still 18 kilometres from the Alpine Fault trace at Inchbonnie, and many other active faults such as the Kelly Fault, the Kakapo Fault, and the Hope Fault, are much closer.

Further south, other faults such as the Main Divide Fault zone of Cox and Findlay (1995), pass right through the main lichen dating areas used by Bull near Mount Cook. Therefore it is likely that earthquakes on faults other than the Alpine Fault have at least supplemented, if not locally dominated, the lichen record of Bull (1996).
Figure 6.4: Data sites for the lichen analysis of Bull (1996). Note the considerable distance from the Alpine Fault of many of the sites where lichen of an appropriate species for dating can be found.

While there are criticisms of lichenometry as a method (see for example McAlpin, 1996; Oelfke and Bulter, 1985) Bull is able to demonstrate a correlation between historical earthquakes and lichen modes. Furthermore, the
inferred lichen event dates appear to progressively coincide with the older earthquake events inferred in Table 6.4, particularly the Roundtop event.

However, the lichen method may be better suited to near-fault dating of outcrop damage on the relatively arid east coast, rather than attempting to extend the inferences to areas where forest surrounds the fault and can provide a more reliable record.

6.5.4 Recent work since completion and distribution of the Yetton et al. 1998 Alpine Fault earthquake chronology.

Since completion of the EQC (Earthquake Commission) funded report of Yetton et al. 1998, the results of two subsequent paleoseismic investigations have been published and the authors have made comparisons with the earthquake chronology outlined above. They are:

- Wright et al. (1998) and Wright (1998)

Most recently Wright et al., 1998 and Wright 1998 detail the results of the excavation of some subsidence features along the fault trace immediately north of the Waitaha River. Based on radiocarbon samples from sag ponds, and a fault dammed swamp, they conclude that the last ground rupturing earthquake occurred post 210 ± 50 yr B.P.. They consider the penultimate event occurred between 430 ± 50 yr B.P. and 420 ± 50 yr B.P. These results are strikingly similar to the radiocarbon dates for both the Toaroha event and the Crane Creek event previously obtained in Yetton et al. 1998 and this thesis (c.f. 210 ± 50 BP [WK 5529] for the Toaroha River event at Haupiri River and 380 ± 25 yrs BP [Wk 5263] for the Crane Creek event at Crane Creek).

Dendrochronology was also applied at the Waitaha River site. Discs were taken from 25 podocarp trees (mainly rimu) which had been recently felled along the fault scarp. These recorded a large number of suppressions and
growth accelerations, spread over an approximately 750 year period, some of which the authors relate to possible earthquake disturbance. There is a significant growth suppression at 1720 AD ± 10 yrs which the authors consider is the most recent earthquake and note the compatibility of this date with the date for the Toaroha River event proposed here.

There is some difference in their interpretation of the exact timing of the penultimate earthquake. Although 11 trees of the 25 trees at the Waitaha site had growth suppressions or accelerations at 1615 AD, a greater proportion (16) were affected in 1580 AD. On this basis the authors prefer the 1580 AD date for the penultimate event. Wright has also reexamined the lichen record of Bull (1996), and can recognise a period of increased rockfall in 1580 AD. He proposes a relatively small local Alpine Fault earthquake at that time with a rupture length in the order of 100 km.

As noted in Chapter 4, the trees on the Crane Creek site also show a suppression in 1580 AD, but more trees were affected in 1620 AD. However the reason for selecting 1620 AD in this thesis as the preferred date for the penultimate Alpine Fault earthquake in this thesis is the regional forest disturbance pattern (Fig 5.7). This shows a very clear regional disturbance and regeneration of forest in 1620 AD but there is no evidence of forest disturbance and regrowth in 1580 AD.

A moderate growth suppression is apparent in 1580 AD in the NZ cedar in the Karangarua River (Fig.5.10). Given that the 1580 AD growth suppression is indicated at several locations, but there is no corresponding increase in regional tree mortality and subsequent regeneration, I consider the suppression at 1580 AD is most likely to have a climatic origin. A very wet or cold period may also explain the increase in rockfalls that is suggested by the lichen record of Bull (1996) at this same time.
Wright et al. (1998) also provide evidence for possible earthquakes at around 1428 AD ± 10 yrs (Geologists Creek event), 1224 AD ± 10 yrs (Muriel Creek event), and 961 AD ± 10 yrs (Roundtop event). They note this generally matches the earthquake chronology proposed here for the older events.

- Berryman et al. (1998)

The authors summarise the results of trenching, and limited dendrochronology work, at Haast and Okuru in the South Westland section of the Alpine Fault. They outline evidence which indicates a most recent rupture in 1718 AD ± 5 yrs and note that this supports the Yetton et al. 1998 date for the Toaroha River event. They have not been able to date the earlier events, but have evidence of only three ruptures on the Alpine Fault in this area in the last 1000 years.

This contrasts with the chronology proposed here for the central Alpine Fault, which is based on evidence of four ruptures over an equivalent period. It is not possible to confidently assess which of the earthquake ruptures in the central section did not extend southwest to Haast and Okuru. However, there is good subsurface evidence in the trenches at Haast for three 8m single event displacements. This is the first reliable evidence for single event displacements on the Alpine Fault, and this implies that the lapsed time for strain to accumulate in this area may have been broadly similar. If this reasoning is correct then, given the good evidence for the Toaroha event in the south Westland area, it is most likely that it is the Crane Creek event that did not reach this far Haast. Instead, presumably the surface rupture terminated somewhere between Haast and the Paringa River, which is the current southwest extreme of the regional forest disturbance record. More work is required to extend this record further southwest and test this hypothesis.
6.6 SUMMARY AND CONCLUSIONS

On the basis of the evidence presented here, three earthquakes can be recognised during the last 750 years on the central section of the Alpine Fault in the area north of the Paringa River. Details of each event are summarised in Table 6.6 below.

<table>
<thead>
<tr>
<th>Evidence</th>
<th>Trenching</th>
<th>Landslide and terrace ages</th>
<th>Forest age</th>
<th>Tree ring chronologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Toaroha River event</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

| Estimate of Timing | Post 1665 AD, probably 1700 –1750 AD | Post 1660 AD | 1715 ± 15 yr | 1717 AD |
| Rupture Length | A minimum length of 375 km is suggested by trenching and the landslide record but a possible length of 450 km is suggested by the tree ring chronologies. |
| Moment Magnitude Estimate | M = 8.05 ± 0.15 for 375 km and M = 8.15 ± 0.2 for 450 km |

| Crane Creek event | | | | |

| Estimate of Timing | 1480 – 1645 AD | 1488 – 1640 AD | 1625 ± 15 yr AD | 1620 ± 10 yr AD |
| Rupture Length | A minimum length of 200 km is suggested by trenching and the landslide record but a minimum of 250 km is indicated by forest age |
| Moment Magnitude Estimate | M > 7.8 ± 0.1 |
**Table 6.6: Summary of the key features of the most recent three Alpine Fault earthquakes.** The best date estimate for the timing of each event is shown in bold text.

In addition to these three earthquakes, other earlier events can also be inferred, but with decreasing data reliability. These earthquakes are estimated to have occurred at around 1200 AD, 940 AD, 600 AD and 25 BC but it is also possible that some intermediate events within this sequence have not yet been recognised.

The paleoseismic investigations of Wright *et al.* 1998, Wright 1998 and Berryman *et al.* 1998 have all been completed since initial recognition of the earthquake chronology outlined in this thesis and in Yetton *et al.* 1998. These subsequent paleoseismic investigations have provided independent supporting evidence from other locations for this chronology, and in particular support the date estimate of the most recent Alpine Fault earthquake in 1717 AD.
Chapter 7

ESTIMATES OF PROBABILITY OF RUPTURE

7.1 INTRODUCTION

Probability estimates have traditionally been used in engineering to quantify risk. If the probability of a given event occurring can be estimated then the relative risk can be balanced against the likely consequences and appropriate action taken. In general a risk with serious consequences will be planned for even if the probability of the event is assessed as being low. For example river engineers and dam builders regularly consider floods with annual probabilities as low as 0.2% (the “500 year” flood). The advantage that rivers present is that they are always active, with a flow that can be monitored daily to track the variation over time with respect to the average. Furthermore the basic laws of hydraulics to which they conform are relatively simple and well understood.

Unfortunately earthquake behavior is episodic, with long periods of relative hiatus, and many of the critical mechanical controls of the physical process are still poorly understood. As a result the application of probability theory to seismic behavior is at a relatively basic stage. Uncertainty regards the pattern of earthquake recurrence over time leads to most of the difficulty. Nevertheless in this chapter I outline the application of several current methods to the assessment of the probability of an Alpine Fault earthquake as a function of elapsed time.
In all cases, the probability estimates apply to the large part of the central section of the fault between the Taramakau River and the Paringa River. This 190 kilometre length of the central section has the most data available, and in particular a detailed forest disturbance and tree ring record. The probabilities may prove to be broadly similar for the 40 kilometre extension of the central section further southwest to the Haast River, but an assessment of this requires further paleoseismic and forest age investigation in this region.

7.2 THE POISSON MODEL

The simplest model for the temporal distribution of earthquakes is the Poisson model. This assumes that earthquakes are independent events and because of this the elapsed time has no effect on probability estimate. With this method the probability of an earthquake occurring the day after the previous earthquake is assumed to be the same as the probability of an earthquake occurring after several hundred years of inactivity.

To calculate the Poisson probability of an earthquake occurring on the Alpine Fault within a given period of time (say 50 years) all that is required as input for the calculation is the average recurrence interval for the fault and the length of time of interest. The date of the last event is not relevant. If the event record from Chapter 5 is taken, and for this crude exercise the uncertainties in the exact date estimates for the events are ignored, then the probability can be calculated simply from:

\[ P_N = 1 - \exp \left( - \frac{t}{t_r} \right) \]

where \( P_N \) is the probability of rupture during a 50 year interval (i.e. \( t = 50 \) yrs in this example); and \( t_r \) = the average recurrence interval (in this case 211 years from the arithmetic mean of recurrence intervals of the last five events,
and for this purpose including the lapsed time since 1717 AD as one of the four recurrence intervals).

\[ P_N \text{ for } 50 \text{ yrs} = \text{approximately } 20\% \]

The Poisson model is the best method to apply if all that is known is the average recurrence interval. However, this method defies common sense considerations of stress accumulation and release which suggest that some period of time is required for stress to accumulate after an earthquake before the next earthquake can occur (the "seismic cycle"; refer Scholz, 1990). The method tends to overestimate the probability soon after an earthquake but for faults with considerable lapsed time (approximately 280 years for the Alpine Fault) it makes a serious underestimation. Therefore the Poisson probability for the Alpine Fault of around 20% for a 50 year return period gives a lower bound on the probability of an Alpine Fault earthquake but is likely to substantially underestimate the true 50-year probability.

### 7.3 THE METHOD OF NISHENKO & BULAND (1987)

Early developments which improved on the Poisson model assumed that the variation of recurrence intervals was Gaussian and could therefore be described by a mean and standard deviation only (Sykes & Nishenko, 1984; Bakun and Lindh, 1985). However, subsequent examination of sets of historic and prehistoric recurrence data show that, although the concept is supported, the postulated distribution shape is not the best one.

Unfortunately the small number of recurrence intervals available for each fault strongly limits inferences of the correct distribution because each fault would take thousands of years to exhibit a complete set of recurrence intervals. The largest reliable set of recurrence intervals for a single fault in the historic
earthquake record is only nine (from the Miyagi - Oki fault in Japan commencing in 1616 AD). Sieh (1984) infers a sequence of 12 prehistoric earthquakes on the Pallet Creek section of the San Andreas fault but only the most recent 5 have been considered reliable enough to use in deriving possible recurrence models (Nishenko & Buland, 1987).

To overcome the problem, Nishenko & Buland (1987) adopted an alternative approach. They suggest that if enough recurrence histories are combined from as many plate boundary faults as possible, the compilation should be large enough to make some valid deductions about the possible statistical distribution of recurrence intervals. This is the so called "ergodic substitution of space for time", as outlined by McAlpin (1996), which is widely used in geology, seismology, geomorphology and hydrology where the time ranges of events are often large. This assumes that under certain circumstances sampling in space can be equivalent to sampling through time at a single location. In effect, while it may not be possible to extend a recurrence history for an individual fault much past 10 previous earthquakes, by combining the past 2 - 4 earthquakes on 20 similar faults, a composite data set of 40 - 80 recurrence intervals becomes available.

Nishenko & Buland (1987) collected data from a number of plate boundary faults where three or more earthquakes have ruptured the same fault. They initially normalised each recurrence time (T) by the mean for that fault segment (T_{ave}), and displayed the result as a histogram of T/T_{ave} (Figure 7.1a.).

If the normalized recurrence intervals are constant, the histogram would have a single peak at T/T_{ave} = 1.0, but as the figure shows this is not the case. However, the behaviour is quasiperiodic (i.e. displays a pattern but with a statistical spread) with a mean recurrence time which is well defined. Nishenko & Buland showed that a log normal distribution is a good fit to the data (refer Figure 7.1b) and preferable to the alternative of a Weibull distribution. They then refined the analysis by recognizing that a better normalizing parameter for
a lognormal function is the logarithmic mean of the recurrence data for each fault segment \((T)\). Errors associated with the measurement of the recurrence intervals by paleoseismic methods were also recognised in their more refined analysis of the data. Therefore, in a second analysis, they fitted a lognormal distribution to the combined data set, this time normalised by the local geometric mean recurrence interval. This distribution contains more information than the simple Poisson distribution, and thus yields better estimates of the probability of an earthquake occurring on any of the fault segments in the data set during a future period of time. Full details of their method can be found in their publication and a brief outline included here in Appendix two.

Figure 7.1- Recurrence time data from historic earthquake sequences on plate boundaries with more than two recurrence intervals from Nishenko & Buland (1987). (a) Recurrence time \(T\) normalised to the average recurrence time \(T_{ave}\) for that segment. (b) Fits of various probability distributions to the data: dashed - lognormal function; dotted - Weibull function. Lognormal provides the closest fit, especially to the tails of the distribution.
The critical parameter in the Nishenko & Buland approach is the standard deviation of the logarithm of normalized data about the geometric mean ($\sigma_d$) which, in effect, is a shape factor. Their 1987 analysis yielded a value for $\sigma_d$ of 0.21. Their original data set included 53 recurrence intervals from both convergent and transform plate boundaries, which spanned a range of two orders of magnitude in recurrence time and six orders of magnitude in seismic moment (i.e. seismic energy). They include historic earthquakes and earthquake sequences estimated from paleoseismic investigation including, where necessary, an allowance for the error from $C_{14}$ dating for the estimation of prehistoric events.

The method of Nishenko & Buland (1987) was used for estimates of the probability of earthquake rupture in California (Working Group, 1988, 1990). Savage (1991) criticised the methods of the Working Group, including challenging the adoption of the lognormal model and the standard deviation based shape parameter ($\sigma_d = 0.21$) of Nishenko & Buland (1987). Savage (1991) considered a shape parameter of $\sigma_d = 0.21$ was likely to be too small and pointed out that individual standard deviations will apply for each fault depending on the geological setting, fault geometry and age. Logically this is true, but as noted earlier, it is simply not practical to wait thousands of years to collect statistically valid recurrence data for an individual fault. Society requires probability estimates for future planning and hazard mitigation, and currently the best solution is to adopt the ergodic approach. However it should be acknowledged that in averaging the collective data the precision with respect to a given fault will not be as good as if many thousands of years of fault specific data are available.

However, the data originally available to Nishenko and Buland (1987) can now be extended to include an improved Pallet Creek paleoearthquake history (Sieh et al., 1989) and data from more recent paleoseismic investigations of other fault segments such as Wrightwood (Fumal et al., 1993; Biasi & Weldon, 1994);
Phelan Creek (Sims et al., 1993) and Indio (Sieh, 1986), all of which do exhibit more variability in their recurrence intervals. Furthermore, the most reliable recent events in the new Alpine earthquake history (as found in this thesis) can also be included. As a result there is now a set of 68 recurrence intervals, as opposed to the original 54, which can be used to examine Savage's objections and obtain better estimates of the distribution parameters.

The original method used by Nishenko & Buland (1987) has been used here with the expanded data set to derive improved parameters for the general lognormal distribution. To answer the question of whether or not the enlarged data set is of sufficient size to produce statistically valid results, the data set was tested as follows. A random subset of 50% of the expanded recurrence data was selected and the Nishenko & Buland analysis repeated to yield lognormal distribution parameters from the reduced subset. This was repeated 20 times and the lognormal parameters derived from the 20 reduced data sets were compared with the result from the complete data set, and they were found to be not statistically different at the 5% significance level.

Updating and expanding the data set has the effect of widening the standard deviation based shape parameter to $\sigma_d = 0.34$, which reflects the greater scatter about the mean in the new recurrence data. This supports the assertions of Savage (1991) that the original $\sigma_d = 0.21$ is too low. In effect this wider standard deviation allows the possibility of a greater variation in the recurrence interval.

The statistically derived expected value of the recurrence interval for the central section of the Alpine Fault using the updated Nishenko and Buland method is 181 years. This is slightly less than a simple arithmetic mean of the last four recurrence intervals due to the shape of a lognormal distribution. With this statistically - derived fault - specific expected recurrence interval, and the global shape parameters of the fitted lognormal distribution, conditional probability calculations can be carried out which take into account the absence
of an earthquake event in the last 280 years on the Alpine Fault. Table 7.1 presents the conditional probabilities of an Alpine Fault rupture during various periods of time beyond 2000 AD.

<table>
<thead>
<tr>
<th>Years Hence from 2000</th>
<th>Probability of an earthquake event (%)</th>
<th></th>
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<tbody>
<tr>
<td></td>
<td>Average</td>
<td>Range</td>
</tr>
<tr>
<td>1</td>
<td>2</td>
<td>1-3</td>
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<td>5</td>
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<td>60-85</td>
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<td>70</td>
<td>80</td>
<td>65-90</td>
</tr>
<tr>
<td>100</td>
<td>90</td>
<td>80-95</td>
</tr>
</tbody>
</table>

Table 7.1: Approximate conditional probabilities of a future earthquake on the central section of the Alpine Fault (between the Taramakau and Paringa Rivers) for time intervals beyond 2000 AD using the updated data set of Nishenko & Buland (1987) and all available recurrence intervals from plate boundary faults around the world (i.e. with $\alpha_d = 0.34$). The probabilities for periods of 20 years and longer have been rounded off to the nearest 5%. The range represents the 90% confidence interval for each calculated conditional probability.
Sensitivity Analysis

Applying the Nishenko & Buland method to the updated data set of intervals yields the best current probability estimate of rupture on the Alpine Fault based on the recurrence behaviour of other plate boundary faults around the world. Nevertheless, it can be criticised because the recurrence pattern that the method assumes has been derived in large part from other faults, and it is possible that the Alpine Fault is somehow different.

<table>
<thead>
<tr>
<th>Years Hence from 2000</th>
<th>Probability of an earthquake event (%)</th>
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<tbody>
<tr>
<td></td>
<td>Average</td>
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<td>1</td>
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<td>60</td>
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<tr>
<td>100</td>
<td>70</td>
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</tbody>
</table>

Table 7.2: Approximate conditional probabilities of an earthquake on the central section of the Alpine Fault (between the Taramakau and Paringa Rivers) using the method of Nishenko & Buland (1987), but based only on recurrence data from other transform plate boundaries (i.e. with $\sigma_d = 0.46$). The probabilities for periods of 20 years and longer have been rounded off to the nearest 5%. The range represents the 90% confidence interval for each calculated conditional probability.
It is apparent in the most recent estimates of recurrence intervals around the world, which tend to have been from transform plate boundaries, that these may exhibit more variation in recurrence interval than subduction boundaries. To look at the effect of excluding the subduction events the probability calculations have been rerun using only the data from transform plate boundaries such as California, Alaska and the Alpine Fault. As a result the population of recurrence intervals drops significantly from 68 to 29, which makes the statistical completeness doubtful. However, the trends are interesting, and there is a marked increase in $\sigma_d$ from 0.34 to 0.46. Some of this simply reflects the influence of a smaller data set, but there is no doubt that the transform recurrence data currently available are more variable.

Using this new $\sigma_d$ the estimated conditional probability of an earthquake event occurring in any given period from 2000 AD, given the absence of an event since 1717 AD, is shown in Table 7.2. Note that these estimates are somewhat less than those of Table 7.1 which are based on $\sigma_d = 0.34$. For example the fifty-year probability of an event reduces from 65%, down to 50 %, and the range increases slightly from 55% - 80% to 35% - 65%.

How sensitive are the probabilities to assumptions regards the timing of the most recent earthquake? The trenches tell us conclusively that an Alpine Fault earthquake definitely occurred between 1660 AD and 1840 AD. I have then used the forest disturbance and tree ring data to refine the timing estimate to 1717, but what happens when the method is applied to the full possible date range for this event supposing that the forest disturbance and tree ring results were not available? Adopting the statistically valid full data set of 68 recurrence intervals from all plate boundaries I have rerun the conditional probabilities for a last event in 1660 AD and a last event in 1840 AD. This makes little change to the calculated conditional probabilities. For a last event in 1660 AD the 50 year probability increases from 65% to 70 % (60 – 85%); for a last event in 1840 AD it reduces to 60 % (40 – 75%). This relatively minor change in the result reflects the indisputable fact that even the shortest
possible lapsed time of 210 years from 1840 AD to 2050 AD is large in comparison to the expected recurrence interval for earthquakes on the Alpine Fault.

Could the statistically derived expected value of the recurrence interval of 181 years be incorrect? The relatively short 100 year interval between the most recent event in 1717 AD and the penultimate Crane Creek event in 1620 tends to reduce expected recurrence interval. What is the effect of excluding the 1620 AD event altogether, and thereby artificially extending the average recurrence interval? Rerunning the probabilities with all the plate boundary data, but without the Crane Creek event, increases the statistically derived expected value of the recurrence interval to 275 years (from 181 years). This results in a 50 year conditional probability of approximately 50% (range 25 – 70%) and a 100 year probability of 70 % (50 – 90%) compared with 65% and 90% respectively in Table 7.1. Once again this is a relatively small reduction in the conditional probability despite a substantial artificial change in the expected recurrence interval.

In summary, no matter what assumptions are made regarding the date of the last earthquake and the expected recurrence interval of earthquakes on the Alpine Fault, the calculations indicate high conditional probabilities of rupture in relatively short future time periods. The lowest conditional probabilities are in the order of 50% in 50 years and 70% in 100 years. If the most likely dates for the prehistoric Alpine Fault earthquakes are adopted, in conjunction with the most statistically robust data set, then the 50-year conditional probability exceeds 50%.

7.4 THE METHOD OF SAVAGE (1994)

It was noted earlier that Savage (1991) was critical of the Nishenko & Buland approach and suggested that individual faults are likely to have their own standard deviation in recurrence interval which will vary from that derived from
an ergodic grouping. Savage (1994) suggests a simple method of assessing conditional probability, which is based on Bayes theorem, and requires only the data from an individual fault.

For this method to work at all for the Alpine Fault a large data set is needed and I have had to apply the method to the full earthquake sequence suggested in Chapter 6, not just the more reliable recent events. This provides 6 recurrence intervals if all the possible earthquake dates in Table 6.4 are included, along with their uncertainties. While it is quite possible that some earthquake events are missing from this extended sequence, and others may be incorrectly inferred, the approximate result is still useful for comparison. Table 7.3 compares the results with the conditional probabilities calculated by the Nishenko & Buland (1987) method for the same return periods.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>From 2000</td>
<td>Average</td>
<td>Range</td>
</tr>
<tr>
<td>30 yr</td>
<td>33</td>
<td>16 - 50</td>
</tr>
<tr>
<td>50 yr</td>
<td>33</td>
<td>16 - 51</td>
</tr>
<tr>
<td>100 yr</td>
<td>67</td>
<td>49 - 85</td>
</tr>
</tbody>
</table>

Table 7.3: Probabilities using the method of Savage (1994), for all possible Alpine Fault earthquakes listed in Table 6.4, in comparison to probabilities for similar periods adopting the updated Nishenko & Buland (1987) approach with all plate boundary data (Table 7.1).

The method of Savage shows the limitation in the resolution of probabilities in the medium term (i.e. 30 - 50 years) which follows from a limited data set. This also casts doubt on the statistical validity of such an approach given the present data set. Intuitively one suspects the true 30 year value should be lower than the 50 year value, but with the Savage (1994) method, and the available data set, there is no way to differentiate this. The range about the
average is also greater (i.e. for 50 years, a 34 year range compared to 27 years for Nishenko & Buland). In general the probabilities are lower than those predicted by Nishenko & Buland (1987), in part due to the increased likelihood of missed events in the extended record. It is interesting to note, however, that they are still significantly higher than the 20% predicted for 50 years using the simple Poisson method (earlier section 7.2).

7.5 POSSIBLE CLUSTERING IN EARTHQUAKE RECURRENCE

Part of the difficulty in assessing fault probability is the possibility that some faults may produce clusters of earthquakes followed by a much longer intercluster time interval. This was first suggested by Sieh et al. (1989), when they developed improved precision in their prehistoric dates for the Pallet Creek section of the San Andreas Fault and is supported by the historical earthquake catalogue in some regions of the world (Allen 1975; Kagan & Jackson 1991; Marco et al. 1996). McAlpin (1996) has developed the idea of bimodal recurrence intervals in relation to possible clustering, one mode representing the intercluster recurrence, and the other the longer interval between clusters. Figure 7.2 shows the pattern of inferred earthquakes for Pallet Creek which to date is the best-defined example of possible clustering.

The Pallet Creek pattern is characterised by variations in the recurrence interval from 44 years to around 330 years with a mean of 132 years. The standard deviation is 99 years (i.e. 75% of the mean recurrence). Sieh et al. (1989) point to the longer intervals as separating clusters of earthquakes with much shorter recurrence intervals. This is one interpretation of the data but the difficulty is once again the time of sampling and hence the small number of data. Any particularly long recurrence interval in this type of short sequence will appear to break the pattern and define "clusters" of an apparently more typical shorter recurrence interval for the periods before and after it.
These long "intercluster" intervals at Pallet Creek vary in length from 200 years (events F - I) to 332 years (events V - X). The clusters themselves vary in having either 2 or 3 earthquakes in each one, and the "intracluster" recurrence intervals are not constant (variation from 44 years to 134 years, a standard deviation of 30 years or 44% with respect to the mean).

Figure 7.2: Dates from the last 10 earthquakes to have ruptured the San Andreas fault at Pallet Creek obtained by trenching studies. Error bars indicate uncertainties in the radiocarbon dates. Events are assigned alphabetical letters back from Z (the most recent) but with gaps for possible missed events (many of which have now been discounted). For comparison the most recent events for the Alpine Fault are also shown with appropriate error bars.
An alternative explanation of the pattern is that it is simply a random expression of the natural variation in earthquake recurrence as demonstrated by the global analysis of Nishenko & Buland (1987). If another 20 or 30 earthquakes were recorded there may be "clusters" of one earthquake i.e. two consecutive long intervals, just as there are now three consecutive shorter intervals. Similarly there may also be "clusters" of more than 3 events.

To test this hypothesis one thousand recurrence intervals were synthesised as follows. For each event a random number between zero and one was generated. This random number was set as the cumulative probability associated with that event. The shape parameters (global distribution mean and standard deviation) and the mean recurrence interval, previously derived for the updated real Nishenko and Buland data set, were then used in an inversion of the log normal distribution to yield the synthesised recurrence intervals.

The one thousand synthesised data points were subsequently analysed using the Nishenko and Buland method to demonstrate that these yielded similar distribution parameters to the ones derived from the real data points. This confirmed that the synthesised data conforms to the characteristics of the real data. The synthetic intervals were then converted into a series of synthetic "dates", and plots of the resulting sequence were examined for patterns of clustering.

Figure 7.3 presents the best example of apparent clustering selected from this synthetic data. The clustering is not pronounced enough to demonstrate that the Pallet Creek recurrence pattern is simply the result of random variations conforming to the Nishenko and Buland distribution parameters, but it does suggest that this one possibility. A better resolution of the true pattern for the San Andreas Fault behaviour will either require several hundred more years of lapsed time and earthquakes, or a significantly improved paleoseismic record.
Figure 7.3 Apparently clustered synthetic "earthquake dates", selected from 1000 recurrence intervals. These were generated using random numbers, including a random dating error, and conform to the distribution parameters of Nishenko & Buland (1987). The top graph shows the "earthquake" sequence which surrounds the more detailed plot in the lower graph.
The earlier Figure 7.2 includes the Alpine Fault record plotted on the same scale as the Pallet Creek data, to provide a comparison between the Pallet Creek data, the synthetic data, and the most recent Alpine Fault earthquake record. It should be noted that the inferred Alpine Fault earthquake events prior to 1440 AD are tentative (refer section 6.4).

In general the Alpine Fault recurrence pattern appears to be more regular than that from Pallet Creek, and more closely resembles similar patterns apparent through random variation in the synthetic data (Figure 7.3). The current Alpine Fault earthquake record does not suggest that clustering is a major factor in the Alpine Fault recurrence pattern.

There are some important geological differences between the Pallet Creek segment of the San Andreas Fault and the Alpine Fault. In general the San Andreas fault is strongly segmented and has a smaller total cumulative offset, suggesting a greater range of mechanical properties along that fault. Therefore we would expect a more erratic recurrence pattern for the San Andreas Fault.

In conclusion, while there is apparent evidence for clustering of earthquakes on some faults, this may only be the result of having short time samples of the recurrence intervals available for analysis. The variation in recurrence intervals exhibited in the best-defined example of the San Andreas Fault is similar to that predicted from the results of the ergodic analysis of Nishenko & Buland (1987). In the absence of clear evidence to the contrary, the best guide to future Alpine Fault behaviour remains the averaged pattern of recurrence which has been observed globally on other plate boundary faults, including the Pallet Creek segment of the San Andreas Fault, which forms part of the updated global data set.
7.6 MECHANICAL AND SLIP RATE CONSIDERATIONS

Some authors have considered the likely mechanical strength of the Alpine Fault in relation to the high strain rates indicated by geodetic networks and the long term slip rates. This has not been used directly to calculate a probabilities of rupture but the observations of high strain do highlight the finite limit of the elapsed time before rupture.

Walcott (1978) made several assumptions about crustal thickness, fault dip, and the rates of plate motion. By calculating the rate of accumulation of seismic moment from this data he estimated a return period of 255 years for Magnitude 8 earthquakes. Walcott went on to say that if earthquakes of this magnitude and regularity actually occurred "we would expect to see features a few thousand years in age or older offset across the Alpine Fault at rates comparable to plate motion". At that time (1978) such features had yet to be recognised (e.g. Berryman, 1978). However, since that time the evidence for very high rates of horizontal movement accounting for at least 75% of the plate motion has been well documented (the remainder of this movement being distributed further east of the Alpine Fault, particularly in the Porters Pass – Amberley Fault zone, refer section 1.3.4).

Scholz (1990) emphasises energy balance considerations with respect to the general question of earthquake recurrence. The seismic cycle of stress build up and earthquake release is an energy balance, and though it is aperiodic, it must be characterised by a mean recurrence time because the strength of the fault must be a well defined and relatively constant quantity (though spatially variable).

Berryman et al. (1992) noted that the strain rates are high on the Alpine Fault, going on to note that they "can only continue to accumulate for one or two hundred years before ultimate rock strain of c. 1 - 10 x 10^-6 is reached". They based this observation on general estimates of ultimate crustal rock strain from
Rikitake (1976), and the absence of an Alpine Fault earthquake since 1840 AD. In light of our conclusion that the last Alpine Fault earthquake was much earlier than 1840 (i.e. 1717 AD) it follows that the time available before rock strains of this order are exceeded is now less than 100 years.

Finally the other indication that the time of the next earthquake is approaching is the limited information on single event displacement. The largest single event displacements reported are in the range of 8 - 9 m (Berryman et al., 1992; Sutherland & Norris, 1995; Berryman, pers. comm. 1999), and all these are from the south Westland end of the fault. Definite single event displacements are harder to find in the central section, although it appears that single event slip may decrease in the central section (i.e. 6 – 7 m; Toaroha River, section 3.2.1, page 59; Wright, pers. comm. 1998). At a conservative 25 mm/yr of average horizontal fault slip, it would take only 300 years to produce 8 m of slip and 240 years to produce 6 m. This compares with a lapsed time since the last event of c. 280 years. If 30mm/yr is closer to the true horizontal fault slip rate (refer discussion section 2.5, page 53), then these times reduce to 267 and 200 years respectively.

7.7 SUMMARY AND CONCLUSIONS

Estimates of the probability of occurrence of the next earthquake in a given time interval from the present for the central section of the Alpine Fault between the Paringa and Taramakau Rivers vary depending on the analysis method adopted. The simplest method of probability assessment assumes a Poisson distribution of earthquake events through time and indicates a 50-year probability of around 20%. However, application of this method disregards the basic mechanical considerations of stress build-up and release on faults (the seismic cycle), and makes no allowance for the substantial lapsed time since the last earthquake. Clearly it represents a significant underestimation of the true probability.
A much better guide to the future behavior of the Alpine Fault is given by the averaged pattern of recurrence intervals that have been observed globally on other plate boundary faults. This approach assumes that the Alpine Fault behaves like other plate boundary faults around the world, which appears to be a reasonable assumption. Analysis has defined the extent to which previous historic and prehistoric earthquake recurrence intervals have varied about a mean, and how often this variation has occurred. This is the basis of a method of probability assessment first proposed by Nishenko & Buland (1987), and which has been updated here by including more recent data. Applying this method to the Alpine Fault indicates a conservative 50-year probability estimate of 50% or more, and a 100-year probability of more than 70%. However, both these estimates have substantial uncertainty ranges of around 25% and 30% respectively.

Other methods of probability assessment, which rely solely on the recurrence record of an individual fault, have also been attempted for the Alpine Fault but the resolution in prediction is severely restricted by the limited number of recurrence intervals for which reliable data are available. However, by adopting these methods, the probabilities appear to be less than the Nishenko & Buland (1987) estimates, but are still relatively high (i.e. 50-year, 33%; 100-year, 67%, uncertainty ranges for both of approximately 35%).

It is unlikely that the Alpine Fault can continue to accumulate stress at the current rate without rupture in the next 100 years. Mechanical considerations, as well as the limited single event slip information that is available, both suggest that there is a very high probability of an Alpine Fault earthquake occurring within the next 100-year period. This time period is less than the nominal design life of most structures, and the time window adopted in most planning and infrastructure development, and as a result the Alpine Fault presents a significant seismic hazard. Therefore, it would be prudent to consider an M 8 Alpine Fault earthquake in the design of structures, and in the planning and management of both the natural and built environment.
Chapter 8

LIKELY CHARACTERISTICS OF THE NEXT ALPINE FAULT EARTHQUAKE

8.1 INTRODUCTION

The assessment of the range of various seismic parameters for the next Alpine Fault earthquake, such as the magnitude, intensity, and the likely range of ground motion parameters, is important for effective hazard mitigation. There are inevitable limitations in accuracy in making this type of assessment, but by using relationships derived from past historical earthquakes as a guide, it is generally possible to at least bracket the expected range for the important seismic parameters, which is sufficient for most purposes. Hazard planning and civil defence preparation is not very sensitive to variation in these parameters, for example mitigation options are broadly similar for a M 7.7 earthquake as opposed to an earthquake of M 8.1. Variations such as these become more significant at the most distant sites, where the damage is normally much less anyway, and the consequences of an under or over assessment are of less importance.

The main unknown in making such an assessment is the length of the Alpine Fault that will rupture in the next earthquake. The previous two earthquakes on the Alpine Fault occurred pre-historically (i.e. more than 150 years ago
and prior to European settlement), but these provide important indications of
the likely future rupture pattern. As noted in chapter 6, the most recent
earthquake of 1717 AD (the Toaroha River event) ruptured all of the central
and south Westland sections of the fault (350 km), as well as a short length
of the north section (25 km, between the Taramakau and Haupiri Rivers). At
the assumed southern rupture limit near Milford Sound, the 1717 AD rupture
may also have included part of the offshore southern continuation of the
Alpine Fault, but this remains unproven.

The forest disturbance data outlined in chapter 5 indicates the penultimate
Crane Creek event of around 1620 AD ruptured the central section from at
least the Paringa River north, but also included a more significant length of
the north section (i.e. from at least the Taramakau River north to the Rahu
saddle, and probably further beyond). The rupture of central section in both
events within approximately 100 years of each other is consistent with this
section having the highest recorded long-term slip rates (estimated at 30 ± 5
mm/yr, refer discussion in section 1.1.2) and the highest current geodetic
strain rate (Beaven, pers. comm., 1999).

Therefore the most likely future earthquake scenario is for rupture of at least
the central section of the Alpine Fault, and the main question is how much of
the adjoining sections will also rupture. Will the south Westland section,
extending from Haast to Milford Sound and possibly offshore, be included
next time as it was in 1717 AD? The long-term slip rate on this section of the
fault is still comparable to the central section (i.e. 26 ± 7 mm/yr at Lake
McKerrow, South Westland [Sutherland and Norris, 1995] c.f. to 30 ± 5
mm/yr). And will the next event be once again limited at the north end to the
Haupiri River? Given that the area further north did not rupture in the last
event, could some (or all) of this section rupture next time?

The answer to these questions requires more paleoseismic investigation
work in these adjoining areas of the fault. In the interim the best guide to the
next event are the likely characteristics of the last two events, and in particular the Toaroha River event of 1717 AD, which has the best defined rupture limits.

8.2 LIKELY MAGNITUDE RANGE

The magnitude of the next event can be estimated by two methods. The crudest is to utilise the lapsed time (280 years) and the estimates of average slip rate (Section 1.2.2). Assuming rupture within the next decade, this gives an estimate of the average horizontal fault slip over the central section in the next event of approximately 8.4 m (range 7.0 – 9.8 m). If another 25 years were to elapse the expected slip would increase to 9.1 m (range 7.6 – 10.7 m). From Wells & Coppersmith (1994) regression relationships between average slip and magnitude, this equates to moment magnitudes ranging from $M = 7.5$ to $M = 8.4$.

The preferred approach, for the reasons outlined in section 6.1.1, is to use the more reliable rupture length estimates. This suggests the Toaroha River event of 1717 AD had a moment magnitude range of $M = 8.05 \pm 0.15$ (Wells & Coppersmith, 1994) to $M = 8.0 \pm 0.26$ (Anderson et al., 1996). This assumes the minimum case with no offshore rupture of the fault south of Milford.

For this assessment, which is intended primarily for general hazard planning and civil defence purposes, it is sufficient to work with $M = 8$ as a reasonable guide but to bracket the possible range as $7.75 < M > 8.25$.

8.3 TYPICAL INTENSITIES

The Modified Mercalli Intensity Scale is a commonly used measure of the effects of earthquake shaking and provides a reasonable guide to the impact of future earthquakes for most civil defence and general hazard planning
purposes. This scale can also be broadly correlated with ground accelerations and velocities if required, as outlined in the next section.

The original New Zealand Modified Mercalli Scale of Eiby (1966) defined twelve intensity levels ranging from MM 1 (essentially not felt) to MM 12 (damage virtually total). This scale was revised in 1992 to better define the MM 1 – 10 levels by adopting New Zealand building types and associated codes (Study Group, 1992). The highest two levels of MM 11 and 12 were not formerly redefined in this new scale because these levels of shaking intensity had not been reached in any of the relatively well documented New Zealand historical earthquakes since 1900. However, the possibility of levels of shaking higher than MM 10 occurring in future events was acknowledged.

An attenuation relationship for Modified Mercalli intensity was first proposed for New Zealand conditions by Smith (1978). This relationship was used with only minor variations in Smith & Berryman (1983,1986) and Elder et al. (1991). Based on the analysis of 30 New Zealand historical earthquakes Dowrick (1991,1994) proposed a revised intensity attenuation relationship which uniformly reduced the expected intensities at a given distance.

All these attenuation relationships take the general form:

\[ I = a + bM + cr + d \log r \]

Where \( a, b, c, \) and \( d \) are constants, \( M \) is the magnitude (moment magnitude where known), and \( r \) is the distance in kilometres to the earthquake epicentre.

Figure 8.1 provides a comparison of these various attenuation relationships for New Zealand conditions with some others from overseas. It is immediately apparent how conservative the original Smith (1978) model is in terms of the over estimation of shaking intensity at moderate to long
distances for a particular earthquake. This resulted in an over estimation of
the hazard associated with the Alpine Fault at distant locations such as
Christchurch by Elder et al. 1991.

For any given earthquake epicentre the relationships which take the general
form outlined above predict circular isoseismal lines. However, in nearly all
historical earthquakes in New Zealand the intensity isoseismals have been
elliptical in shape and this pattern varies regionally (Smith 1978; Elder et al.
1991; Smith 1995 a & b). In the South Island the major axes of the
isoseismals are generally aligned N40°E and this reflects the mean trend of
fault strike between N30°E and N50°E (Elder et al. 1991).

Smith (1995a & b) presents an updated attenuation model based on over
one hundred historical New Zealand earthquakes between 1855 and 1990
and introduces a more sophisticated modeling method. This incorporates the
observed regional variations in isoseismal ellipticity and allows the
construction of synthetic isoseismals for hazard planning purposes for
potential future earthquakes for which a magnitude and epicentral location
are known.

Based on the paleoseismic evidence of rupture length from this study and a
moment magnitude of 8, Smith (Seismological Observatory, pers. comm.,
1997) has provided synthetic isoseismals for the two most recent Alpine
Fault earthquakes (Figures 8.2 and 8.3). Figure 8.1 compares the new
attenuation relationship which is inherent in these two figures with the
various one dimensional attenuation relationships discussed earlier.

To include the Smith (1995a & b) relationship in this comparison in Figure
8.1, two attenuation curves have been derived from Figures 8.2 and 8.3
respectively. The first comes from a line normal to the Alpine Fault trace at a
bearing of 325 ° from Timaru, which intersects the fault at approximately
Franz Josef. The other curve comes from a second normal to the fault trace
Figure 8.1 - A comparison of various New Zealand and international intensity attenuation relationships at moderate epicentral distances for an M8 earthquake. Note the very conservative early relationship of Smith (1978) has been substantially revised (Smith 1995 a & b) and now compares well with that of Dowrick (1992, 1994). The curves from Smith (1995 a & b) are derived from Figure 8.2 and 8.3 for normals extending southeast from Franz Josef and the Ahaura River respectively.
at a bearing 325° from Christchurch, which intersects the fault at the Ahaura River. The slight difference between the curves reflects the expected widening of the isoseismals towards the south in the Smith model, which is apparent in Figures 8.2 and 8.3, and has been observed in other historical earthquakes in this part of the region.

Both examples fit closely to the Dowrick curve and effectively bracket his simplified relationship at epicentral distances of 90 – 200 km. At closer distances the Dowrick relationship predicts slightly higher shaking intensities. From the hazard modeling perspective the actual differences between the curves are small and it is clear from Figure 8.1 that there is now converging agreement between recent New Zealand intensity attenuation relationships.

The modeled isoseismals for the two most recent events in Figures 8.2 and 8.3 are based on the rupture lengths summarised in Chapter 6 and both assume a Moment Magnitude $M = 8$. Note that as well as being elliptical the isoseismals are slightly offset east with respect to the fault trace. The input required for the Smith model include both the fault dip and the focal depth. Based on the estimates of fault dip from the recent crustal studies, and the conclusions of Leitner et al. (submitted) regarding seismogenic depths along the Alpine Fault [refer discussion of both in Chapter 1, section 1.3.3], a fault dip of 45° and a focal depth of 10 km has been adopted.

This has the effect of shifting the expected isoseismal pattern east with respect to the surface trace by approximately 10 km. This offset also accords with observations from historical earthquakes overseas that the hanging wall of moderately dipping faults frequently experiences enhanced shaking effects (Archeletta, pers. comm. 1999).
Figure 8.2 - Synthetic isoseismals for the Toaroha River event of 1717 AD adopting the rupture limits summarised in section 6.6, a Moment Magnitude (M) of 8, and the intensity formula and computational methods of Smith (1995 a & b). Reproduced courtesy of Warwick Smith, Seismological Observatory (pers. comm., 1997).
Figure 8.3 - Synthetic isoseismals for the *Crane Creek event* of 1620 ± 10 AD adopting the rupture limits summarised in section 6.6, a Moment Magnitude \( (M) \) of 8, and the intensity formula and computational methods of Smith (1995 a & b). Reproduced courtesy of Warwick Smith, Seismological Observatory (pers. comm. 1997).
The most reliable portions of the isoseismals are in the middle section where assumptions regarding the length of the rupture are less relevant. These isoseismals are also for average ground conditions with no allowance for amplification, although Dowrick (1994) argues that some amplification is already built in to the data because most observed intensities in historical earthquakes are from alluvial locations prone to amplification. Stirling et al. 1999 also notes this potential data bias but expect deep soft sediment sequences such as that underlying Christchurch to generate amplification at levels significantly above this average (for further discussion refer below and section 9.3.2).

The isoseismal "lines" are drawn as 20 km fuzzy bands to better emphasise the limited accuracy possible in this type of regional construction. It is also likely that in both the previous earthquakes there was an area of MM 10 or greater shaking closest to the fault trace, but this is the area most sensitive to local fault effects and has not been modeled. There are also no directivity effects in the modeling and the epicentre and rupture patterns are assumed to be symmetrical. Given the current level of knowledge of these past events, and the regional perspective of this project, these simplifications are both necessary and justified.

It appears that the Toaroha River event of 1717 AD resulted in relatively high shaking intensities for locations in the central and southern areas but slightly lower intensities for the northern locations such as Christchurch. This reflects the inferred northern limit of the rupture at the Haupiri River. In contrast the Crane Creek event, which the forest disturbance record indicates extended north past Haupiri another 50 km to at least the Rahu Saddle, is a more significant event for most northern locations. Unfortunately we have no data from north of the Rahu Saddle to provide a basis for the possible extension of the model. However, a rupture extending further north
than the Rahu starts to also become a significant earthquake for such locations as Nelson and Blenheim.

Table 8.1 below summarises the intensities which are predicted from Figures 8.2 & 8.3 at a range of locations for each scenario.

<table>
<thead>
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<th>Location</th>
<th>Predicted Intensity (MM units)</th>
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<th>Toaroha River event</th>
<th>Modal Intensity</th>
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<td>Hanmer</td>
<td>8</td>
<td>7</td>
<td>7 - 8</td>
<td></td>
</tr>
<tr>
<td>Amberley</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td></td>
</tr>
<tr>
<td>Kaiapoi</td>
<td>7</td>
<td>6 - 7</td>
<td>6 - 7</td>
<td></td>
</tr>
<tr>
<td>Christchurch</td>
<td>7</td>
<td>6 - 7</td>
<td>6 - 7</td>
<td></td>
</tr>
</tbody>
</table>

Table 8.1 - Predicted intensities for the two most recent Alpine Fault earthquakes based on Figures 8.2 and 8.3. The modal intensity from these is included to provide a guide to the most likely intensity in a future earthquake. No increase for amplification has been included in this table but local site conditions may significantly change the actual intensities experienced.

Figure 8.4 shows graphically a comparison of these predicted intensities with those recorded historically during other significant earthquakes. To make a meaningful comparison for Christchurch and Kaiapoi we have made an arbitrary increase of one intensity unit to allow for amplification of the
expected bedrock intensities. This has been demonstrated both theoretically (Elder et al., 1991; Haines et al., 1994) and directly (Toshinawa et al., 1997).

It is apparent from Figure 8.4 that at most locations the next Alpine Fault earthquake will be significantly larger than other earthquakes experienced at these locations in the last 150 years.

**Figure 8.4** - A comparison of predicted intensities associated with the next Alpine Fault earthquake and those experienced in other significant earthquakes over the last 150 years. For the predicted intensity of the next Alpine Fault earthquake at Christchurch and Kaiapoi an arbitrary increase of one MM intensity unit has been added to the value from Table 8.1 to allow for amplification by the soft sediments at these locations which has been observed in historical earthquakes and is predicted by theoretical modelling (refer Elder et al., 1991; Haines et al., 1994, Toshinawa et al., 1997).
8.4 LIKELY GROUND ACCELERATIONS, VELOCITY AND DURATION

There are numerous published correlation's which broadly convert MM Intensity to ground acceleration (e.g. those of Richter, 1958; M.S.O.P., 1979; Hunt; 1984). Table 8.2 provides a general guide however it should be noted that local amplification due to ground effects and topography, as well as spectral variations from the rupture process, will result in considerable variation.

<table>
<thead>
<tr>
<th>MM Intensity Unit</th>
<th>Ground Acceleration (cm/sec²) for periods 0.1 - 0.5 sec</th>
<th>Ground Velocity (cm/sec) for periods 0.5 - 2.0 sec.</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>12 – 25</td>
<td>1.0 - 2.0</td>
</tr>
<tr>
<td>6</td>
<td>25 – 50</td>
<td>2.1 - 4.0</td>
</tr>
<tr>
<td>7</td>
<td>50 -100</td>
<td>4.1 - 8.0</td>
</tr>
<tr>
<td>8</td>
<td>100 - 200</td>
<td>8.1 - 16.0</td>
</tr>
<tr>
<td>9</td>
<td>200 - 400</td>
<td>16.1 - 32.0</td>
</tr>
<tr>
<td>10</td>
<td>400 - 800</td>
<td>32.1 - 64.0</td>
</tr>
</tbody>
</table>

Table 8.2 – The likely range of horizontal peak ground acceleration and velocity typically associated with each MM Intensity class for average ground conditions (after M.S.O.P., 1979).

The duration of strong ground motion from earthquake shaking can have a major influence on earthquake damage. Both the degradation in the stiffness and strength of certain types of structures, and the buildup of porewater pressures affecting liquefaction and slope stability, are sensitive to number of load or stress reversals that occur during an earthquake.

Strong ground motion is normally defined as the time between the first and last exceedance of some threshold acceleration. The most commonly used
measure is the *bracketed duration* of Bolt (1969) which adopts a threshold acceleration of 0.05g (Kramer, 1996). Because the duration of strong ground motion is related to the time required for release of accumulated strain energy by rupture along the fault, and as the fault area increases with increasing magnitude, the duration also increases. The duration is also affected by local site conditions, with rock sites having a generally shorter duration. For example a M 8 earthquake in the epicentral area (i.e. an epicentral distance of less than 10 km) would be expected to have a bracketed duration of around 30 seconds at a rock site but twice this value at a soil site (Chang & Krinitisky, 1977).

Due to the effects of ground damping, ground acceleration decreases with distance, and at some distance ground acceleration drops below the threshold value and the bracketed duration becomes zero. Various predictive relationships exist that relate bracketed duration to epicentral distance. Table 8.3 utilises the most widely adopted relationship of Chang & Krinitisky (1977) to summarises the likely range of bracketed duration for an M 8 earthquake on the central Alpine Fault at a range of towns in the central South Island. Most, or all, of the towns are built on alluvial materials and so in these cases the expected bracketed duration for soil sites provides the best general guide. However some parts of these towns, for example the hill suburbs of Greymouth and Christchurch, better approximate to the hard rock sites. In other cases, for example Tekapo Village, the foundation soils are relatively stiff glacial till and the resulting duration is likely to be some intermediate value.
<table>
<thead>
<tr>
<th>Location</th>
<th>Epicentral distance</th>
<th>Bracketed duration for soil sites</th>
<th>Bracketed duration for rock sites</th>
</tr>
</thead>
<tbody>
<tr>
<td>Franz Josef</td>
<td>10</td>
<td>65 seconds</td>
<td>NA</td>
</tr>
<tr>
<td>Mt Cook Village</td>
<td>20</td>
<td>52 seconds</td>
<td>28 seconds</td>
</tr>
<tr>
<td>Hokitika</td>
<td>35</td>
<td>40 seconds</td>
<td>NA</td>
</tr>
<tr>
<td>Greymouth</td>
<td>50</td>
<td>26 seconds</td>
<td>16 seconds</td>
</tr>
<tr>
<td>Tekapo Village</td>
<td>60</td>
<td>24 seconds</td>
<td>12 seconds</td>
</tr>
<tr>
<td>Darfield</td>
<td>90</td>
<td>17 seconds</td>
<td>NA</td>
</tr>
<tr>
<td>Christchurch</td>
<td>130</td>
<td>12 seconds</td>
<td>Zero</td>
</tr>
</tbody>
</table>

**Table 8.3** - The expected bracketed duration of strong ground shaking (Bolt 1969) at a range of locations resulting from a Moment Magnitude 8 earthquake on the central section of the Alpine Fault (based on the predictive relationship of Chang & Krinitsky 1977).

### 8.5 DIRECTIONAL AND LOCAL EFFECTS

There will be complex near source effects associated with an Alpine Fault earthquake. In particular on such a long fault there are likely to be directivity effects associated with the propagating rupture front (Benioff, 1955; Ben-Menachem, 1961; Kramer, 1996). Rupture frequently progresses along a fault as a series of linked dislocations and the constructive interference of earthquake waves produced by successive dislocations can produce strong pulses of large displacement at nearby sites to which the rupture is progressing (Benioff, 1955; Singh, 1985). This tends to focus seismic energy in the direction of rupture.

There is no way to be sure of the direction of propagation of a prehistoric earthquake. If a systematic reduction in single event slip can be demonstrated from one end of the fault to the other then this may provide some guide. At present this information is lacking for the Alpine Fault.
There is a line of reasoning that the rupture is more readily initiated in relatively stiff crust where the seismogenic zone is deeper rather than in thin crust which is widely sheared and jointed (Sibson, pers. comm. 1996). This would tend to suggest initiation at the southern end of the fault in Fiordland and south Westland, with propagation northwards, and a progressive slip reduction into the Marlborough splay faults. This in turn implies directivity effects focussed around the north central section. If this was the case the isoseismals of the MM 9 and greater shaking would tend to be more pear shaped and larger in the north, rather than the relatively symmetrical shape predicted by the Smith model and adopted in Figures 8.2 & 8.3.

The alternative argument is that individual events can initiate at either end, for example at the north end triggered by movement on one of the Marlborough faults, and it is possible the Crane Creek event was an earthquake of this type. In this hypothesis the north end is subject, at different times, to ruptures which propagate from both the south and north ends. Without more detailed paleoseismic work on the rupture limits of the Crane Creek event this possibility is conjecture and can not be usefully incorporated into the modeling.

Topographic effects are also an important control on strong shaking effects in the epicentral area (Trufunac & Hudson, 1971; Aki, 1988; Gelli et al., 1988; Faccioli, 1991). Damage patterns in historical earthquakes have suggested ridges tend to amplify ground motion (for recent examples in Italy and Chile refer Finn, 1991). This was demonstrated most effectively in five earthquakes which occurred in Matsuzaki, Japan, where instrumentation revealed the average peak ground acceleration at the ridge crest was 2.5 times higher than the average peak ground acceleration at the base of the slope (Jibson, 1987). Modelling by Gelli et al., 1988 suggests this effect is most pronounced for ridges with relatively rounded profiles, rather than sharp ridge crests. However, analysis of topographic irregularities is a
complicated problem and predictive methods are still very limited (Sanchez-Sesma & Campillo, 1993).

In an Alpine Fault earthquake in a mountainous environment such as the Southern Alps topographic amplification may be a significant factor in the triggering of rock avalanches, with ridge shape providing some control on their location. The distribution of known rock avalanches in relation to the Alpine Fault is discussed in more detail in section 9.3.3.

Sedimentary basins are another setting in which ground motion can be amplified. The curvature of a basin in which softer alluvial soils have been deposited can trap body waves and cause some incident body waves to propagate through the alluvium as more damaging surface waves (Vidale & Helmberger, 1988) and can also lead to substantial increases in the bracketed duration (Silva, 1988).

There are often significant differences in the amount of amplification between the centre and edges of a valley or basin and these differences have been observed in the damage patterns of numerous historical earthquakes (Kramer, 1996). For alluvial valleys of irregular shape, as are typical in the Southern Alps and adjoining areas, theoretical studies indicate such combined convex/concave areas are likely to result in very complex, even chaotic, ground motion (Rial et al., 1992).

The detailed evaluation of the numerous effects created by directivity, topography and subsurface geometry requires site specific two dimensional, and in some cases, three-dimensional analysis. However, even using such methods, in many cases these effects remain unpredictable (refer the comments on quantitative predictability in Silva, 1988). While this detail is neither feasible nor warranted in a regional study such this, it is important to note that such effects will occur, particularly in the epicentral areas of strongest ground motion in Westland and the Southern Alps.
8.6 SUMMARY AND CONCLUSIONS

The next Alpine Fault earthquake is likely to have a Moment Magnitude of approximately 8. It is most likely to rupture the central section of the Alpine Fault from Haast to Inchbonnie, but the rupture could also extend south to Milford Sound. The extent to which the northern section of the fault ruptures is still difficult to assess.

Attenuation models can be used to estimate the shaking intensity at various locations based on the previous events. The strongest shaking will occur close to the fault where Modified Mercalli Intensities will exceed MM 9. At greater distances, for example Christchurch, the intensities are still likely to be relatively strong (MM Intensity 7 - 8) due in part to the likely amplification of the incident earthquake waves by the soft sediments under parts of the city.

For most locations in the study area the next Alpine Fault earthquake will be larger than any other historical earthquake experienced there in the last 150 years. The impacts will also be on a much larger scale, and result in simultaneous damage to a very large area of the South Island. This will compound the existing access difficulties created by the topography of the rugged central South Island region. Chapter 9, which follows, considers the likely effects of an Alpine Fault earthquake in more detail.
Chapter 9

LIKELY CONSEQUENCES OF AN ALPINE FAULT EARTHQUAKE

9.1 INTRODUCTION

The previous chapter has outlined the most likely range of seismic parameters for an Alpine Fault earthquake, in particular the degree of earthquake shaking and ground acceleration which can be expected. This chapter considers the range of direct and indirect effects likely to be caused by the surface rupture and shaking associated with such an earthquake. Variability in geological materials and conditions prevents the prediction of where exactly many of these effects will occur. The exception to this is ground rupture along the fault trace, the location of which can generally be predicted to within approximately 50 m.

This consideration of effects takes as a starting point a description of the nature and extent of the damage caused by the 1929 Buller earthquake of $M = 7.8$. Similarities have already been noted between the terrain affected by this event and the Southern Alps range front. While the magnitude of the 1929 Buller event is at the low end of the expected range for an Alpine Fault earthquake, the historical analogy provided by this earthquake is still very useful. Analogies can also be drawn with the $M_s 7.01$ 1929 Arthurs Pass earthquake (Dowrick & Roades, 1998) which generated maximum intensities in excess of MM 8 in the mountainous area northeast of Arthur's Pass (Cowan et al., 1996) and triggered rock avalanches in this epicentral area such as Falling Mountain (Hancox et al., 1997). The $M_s 7.4$ 1969 Inangahua earthquake (Dowrick & Smith, 1990; Anderson et al., 1994), also resulted in numerous
9.2 THE 1929 BULLER EARTHQUAKE – THE CLOSEST HISTORIC PARALLEL TO AN ALPINE FAULT EARTHQUAKE

The M₈ 7.8 1929 Buller earthquake is the largest South Island earthquake this Century. Figure 9.1 shows the source region and epicentral area.

The epicentre of this earthquake has traditionally been assigned close to the Buller River, at the point of surface rupture on SH 6, approximately 10 km west of Murchison. This is the location where reverse vertical movement of 4.5 m could be most clearly seen on the White Creek fault. However Adams (1981b) shows an elliptical distribution of landslides with a much more northern bias which would tend to shift the epicentre northeast by around 50 km (i.e. closer to Seddonville and Karamea). This discrepancy was more recently also highlighted by Wellman (1996), and Hancox et al. (1997) have redrawn the isoseismals after a more detailed analysis of the landslide distribution. The redrawn isoseismals also strongly suggest the epicentre should be revised and moved northwards. This shift becomes significant in the derivation of attenuation relationships for New Zealand conditions based on historical earthquakes such as that used in Dowrick et al. (1998).

Intensity estimates for the Buller earthquake are hindered by the lack of population in the epicentral area. MM 10 was reached closest to the fault rupture (Hancox et al. [1997]) although the isoseismic map of Dowrick (1994) does not show this area. Murchison was the most strongly shaken population centre with an estimated intensity of MM = 9, and considered comparable to a peak ground acceleration of 0.5 - 0.9g. Westport was the next most seriously affected large town. Table 9.1 summarises the material damage from the event under the categories of buildings, services, and transportation.
Figure 9.1 - Isoseismals for the 1929 Buller earthquake of $M_s = 7.8$. Note the asymmetry of the epicentre with respect to the area of strongest shaking leading to suggestions that this should be plotted further north. From Downes (1995) based on Dowrick (1994). The MM intensity of 5 at Christchurch is from observatory files but specific investigations of felt intensities (Dibble et al. [1980]; Stirling et al. [1999]) indicate MM 6 for this event in many parts of Christchurch.
<table>
<thead>
<tr>
<th>Type of Damage</th>
<th>Description</th>
</tr>
</thead>
</table>
| **Buildings**  | **Murchison**: MM = 9. The main damage to buildings, apart from Hodgsons Store and the BNZ, was practically confined to the universal loss of chimneys. Some houses fell off piles. The BNZ was partly damaged, and the two storey Hodgsons Store was near collapse, due to large sway deflection of the top floor.  
**Westport**: MM = 8.  
**Greymouth**: MM = 7  
**Nelson**: MM = 8 |
| **Services**    | Not well documented but moderate. Murchison lost its power station for several months. Westport lost electricity for 13 hours, water for 9 days and the telephone exchange for 3 days. Gas mains broke in 4 places. Various of the smaller communities lost telephone and power for lengthy periods because of landslides sweeping the poles away. |
| **Roads and Bridges** | Major damage from landslides occurred within the MM 9 zone, in particular SH 6 between Inangahua and Murchison. This section of road took 22 months to repair and reopen. Landslides also caused major damage to roads in the Matakitaki valley, Maruia Saddle and Karamea Bluff (Corbyvale Pass). Subsidence and lateral spreading of embankments and bridge approaches occurred in many local areas as far away as Greymouth (approx. 120 km from epicentre) and Takaka and Collingwood (a similar distance).  
Structural damage to bridges was not widespread and was generally confined to displacements of the abutments or piers (i.e. the Matakitaki bridge at Murchison, the Lyell Creek bridge, the Newton Creek bridge and the Little Wanganui bridge). |

**Table 9.1** Summary of damage to the built environment during the 1929 Buller earthquake (after Henderson, 1937; Adams, 1981a; Dowrick, 1994).
Ground rupture occurred along the White Creek Fault but the only area easily accessed and investigated was adjacent to the Buller River. The trace was followed northwards in the forested and rugged terrain for a total length of around 8 km but the total length may have been much greater, particularly given the northern bias of the landslide ellipse. The main consequence of the surface fault rupture was disruption of SH 6 at White Creek where the road was offset vertically by around 4.5 m.

While damage to buildings, bridges and services was substantial, the epicentral area was very sparsely populated. As a result it was the natural landscape and its few rural occupants that suffered the greatest damage and this was caused by countless landslides. Pearce & Watson (1986) estimate 410 landslides occurred in just the c.100 km² Matiri catchment. However, landslides extended over a total area which was at least 50 times larger, estimated to be more than 5,000 km² in total (Adams, 1981a). Landslides were more numerous in this earthquake than for any other historic event in New Zealand. This reflects the altitude range and very steep nature of the terrain, with many slopes close to their natural equilibrium.

The extensive landslide damage is also in part attributed to high antecedent groundwater levels in the region prior to the earthquake. The earthquake occurred in June and Henderson (1937) notes the soil and subsoil were saturated. Dowrick (1994) also demonstrates this was a much wetter June than normal based on the weather stations in the region. The great majority of the landslides were surficial failures of residual soil or thin colluvium sliding on top of deeper harder bedrock, but most of the largest landslides were deep-seated failures in the Tertiary units, often parallel to bedding planes.

The importance of landslide impacts is reflected in the record of fatalities. Of the 17 people killed as a result of the earthquake, fourteen were killed by landslides. In nearly all cases they were in houses built on relatively flat low ground but near the toe of steep and high slopes. Debris runout from the
landslides destroyed the houses and killed or injured their occupants.
Henderson (1937) describes landslides in the Matakitaki valley 3 km from Murchison. A deep-seated failure in the Tertiary muddy sandstones, which dip at 30 - 40 degrees towards the valley floor, produced a debris flow which travelled across the river terraces burying these in up to 60 m of debris. Two houses were destroyed at this location with 5 people losing their lives. The Matakitaki River was dammed as result and a 4 km long lake was created which lasted for many years.

Four more people died in a very similar landslide in the nearby Maruia Valley, where slide debris from dipping Tertiary sandstones overwhelmed a homestead. Once again a landslide lake was formed, which was over 5 km long, but in this case the river diverted around it within several days.

Numerous landslide dammed lakes formed as a result of the Buller earthquake. The Buller, Mokihinui, Karamea, Matakitaki, Matiri and Maruia Rivers were all dammed. Most of these dams have subsequently been removed by river erosion. For example the landslide dams on the Karamea River are now marked by steep, boulder-strewn sections of river rapids, with intervening sandy, low gradient reaches (Adams, 1981b). The Mokihinui River gorge above Seddonville was also dammed by landslide debris, killing two prospectors on the track, and forming an 11 km long lake (Lake Perrine) which was up to 20m deep. Three weeks after the earthquake, part of the dam washed out, and the lake suddenly lowered 8m thereby flooding Seddonville. During the flood houses were shifted off their foundations, the hall drifted some tens of meters, and the whole of the Seddonville alluvial flat was deeply covered in water (Henderson, 1937).

Liquefaction was not well understood as a phenomena at the time, but Henderson (1937) describes damage to alluvial ground which is very likely to have been liquefaction induced. He notes the earthquake caused:
"numerous sections of river bank to collapse, rents more or less parallel with them to appear - perhaps many chains back - and large masses of poorly consolidated material to move bodily to the unsupported sides. Movements of this kind occurred commonly throughout the region - roads were fissured and depressed; fences were displaced, and railway lines buckled. A good deal of damage was done in Greymouth and Westport, where cracks opened in the streets, water-mains, gas-mains, and sewers were broken, and houses were distorted. At Westport the wharf was slightly damaged. At Karamea a considerable area north of the township moved towards the river, distorting the wharf and training wall and disrupting a tramway. Many fissures opened along the foot of a terrace of older alluvium on the north side of the area, and also on the east side .......... the area sank by 2 ft 3 in. It is the youngest estuarine flat built by the river and is raised but a few feet above sea level.

- (Henderson, 1937: page 89)

9.3 EXPECTED IMMEDIATE EFFECTS OF THE NEXT ALPINE FAULT EARTHQUAKE

9.3.1 Ground rupture and warping

While much of the Alpine Fault trace is in heavily forested rugged terrain there are a number of points where roads and rail routes cross the active trace. The oblique dextral horizontal offset is likely to vary along the fault trace as has occurred in comparable historical ruptures (refer again Chapter 6, Fig. 6.2). The horizontal offset will increase from zero at the surface rupture limits to around 8 ± 2 m in the south and central area decreasing to around 1.8 ± 1 m in the north. This assumes the range of slip rate of 25 - 30 mm/yr, a lapsed time of approximately 280 years, and a component of natural variance in slip of 30 %. The vertical component will be considerably less, and probably in the range between 0.5 - 1.5 m. The amount of ground distortion and tilting immediately adjacent to the trace will vary in extent with location but will commonly occur within 50 - 100 m of the trace. In some locations (eg trace step-overs, areas
between multiple traces etc) this zone of deformation may locally be more extensive.

In a few locations buildings and houses have been built on the fault scarp or trace. The most obvious example is the small town of Franz Josef (Figure 9.2) where the trace is apparent crossing the main road and projects beneath a service station. More detailed work is required to define the trace through the remaining town area.

At a few rural locations further north (e.g. Inchbonnie, Haupiri River) buildings are at, or near, the top of the scarp. Trenching completed as part of this study suggests the renewed rupture generally occurs near the base of fault scarps, so these buildings are probably not at the most likely new rupture point. While these particular houses will be very severely shaken, and possibly affected by permanent ground warping, they may not be physically ruptured by fault offset within their foundation perimeter.

As noted above the main direct impact of fault rupture will be to roading and rail routes where these cross the trace. Table 9.2 summarises the likely rupture points. While most of the locations listed are virtually certain to be rendered at least temporarily impassable, by far the greatest proportion of road and rail disruption will be caused by secondary effects such as landsliding, liquefaction and structural damage to bridges. These effects are discussed separately below.
Figure 9.2 - The Alpine Fault trace passing through SH 6 at Franz Josef (GR H35/818538). The trace is marked by the slight rise in the road immediately in front of the red 4WD. The fault trace crosses the road and passes under the service station forecourt.
<table>
<thead>
<tr>
<th>Road or rail route</th>
<th>Grid Reference, NZMS 260</th>
</tr>
</thead>
<tbody>
<tr>
<td>SH 7, Lake Daniels rest area</td>
<td>L 31/458725*</td>
</tr>
<tr>
<td>Amuri – Haupiri Rd, Ahaura River</td>
<td>L 32/132484*</td>
</tr>
<tr>
<td>Inchbonnie – Rotomanu road and rail</td>
<td>approx. only at K32/877323</td>
</tr>
<tr>
<td>Inchbonnie – Rotomanu road , Lk Poerua</td>
<td>K32/863314</td>
</tr>
<tr>
<td>Jacksons – Inchbonnie Rd, Inchbonnie</td>
<td>K 32/847304</td>
</tr>
<tr>
<td>SH 73 Rocky Point</td>
<td>K33/825291</td>
</tr>
<tr>
<td>Lake Arthur Road, Kokatahi</td>
<td>J33/588116</td>
</tr>
<tr>
<td>SH 6 Whataroa River</td>
<td>I 35/004667</td>
</tr>
<tr>
<td>SH 6 Whataroa River</td>
<td>I 35/991658</td>
</tr>
<tr>
<td>SH 6 Franz Josef township</td>
<td>H35/818538</td>
</tr>
<tr>
<td>SH 6 near Franz Josef</td>
<td>approx. only at H35/811534</td>
</tr>
<tr>
<td>SH 6 Paddy Creek</td>
<td>H34/744492</td>
</tr>
<tr>
<td>SH 6 Cook Saddle</td>
<td>H34/713463</td>
</tr>
<tr>
<td>SH 6 Karangarua</td>
<td>H34/715466</td>
</tr>
<tr>
<td>SH 6 Karangarua</td>
<td>H34/524322</td>
</tr>
<tr>
<td>SH 6 Haast</td>
<td>F37/902890</td>
</tr>
</tbody>
</table>

* denotes areas which will only be affected if the far northern section ruptures

**Table 9.2**: Points of likely fault rupture of existing road and rail routes in Westland.

### 9.3.2 Ground shaking

Chapter 8 outlines the likely range of intensities and the typical ground accelerations associated with these intensities which can be reasonably expected in the next Alpine Fault event. Ground shaking will cause direct
damage to buildings, contents, and services through cyclic vibration at a wide range of frequencies.

Unfortunately even if the exact characteristics of the seismic waves generated by an Alpine Fault earthquake were fully known these waves will be attenuated, reflected and refracted in a complex and largely unpredictable manner by near surface geological conditions and ground topography. The geology along every wave path varies, and the final outcome, which is ground shaking at a specific surface location, is subject to the sum of all the variability. Additional to this is the wide variety in building type, building age, and natural period which then governs the response of structures to the shaking. However, it is possible to use historical precedent and the current categories of building type to make some general predictions regarding damage levels (Table 9.2). The general locations which could reasonably expect this intensity of shaking are also noted.

Variations in foundation conditions within a town are likely to result in a range of intensity of at least one intensity unit, and possibly up to two. For the case of Christchurch where amplification has been well documented (for example Elder et al., 1991; Haines et al., 1994; Toshiniwa et al., 1997; Stirling et al., 1999) allowance is made for amplification of up to one intensity unit (i.e. the predicted bedrock intensity of MM 7 becomes MM 7 + MM8). The estimates in Table 9.3 should be viewed as general approximations to provide an appropriate scenario for District and Regional Councils and infrastructure providers.

As noted in section 9.1, shaking of MM = 9 has already been experienced on the west coast in the epicentral area of the Buller earthquake. Other New Zealand historical earthquakes which have produced MM = 9 shaking in towns or cities include the Wairapa earthquake (1855, MM = 10 in Wellington and MM = 9 in the Wairapa); the Inangahua earthquake (1968, MM = 10 at
<table>
<thead>
<tr>
<th>Intensity</th>
<th>Impact on People &amp; fittings</th>
<th>Impact on Structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>MM = 7</td>
<td>General alarm</td>
<td>Unreinforced stone and brick walls, and poorly built rammed earth and mud houses (Type I) cracked. Some damage to old but well built unreinforced masonry buildings (Type II). Unbraced parapets and architectural ornaments fall. Roofing tiles dislodged. Many chimneys broken. A few instances of damage to brick veneers and plaster or stucco linings. Water cylinders move or leak. Some cracked windows.</td>
</tr>
<tr>
<td></td>
<td>Difficulty in standing</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Noticed by drivers of cars</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Furniture moves on smooth</td>
<td></td>
</tr>
<tr>
<td></td>
<td>floors and may move on</td>
<td></td>
</tr>
<tr>
<td></td>
<td>carpeted floors. Some</td>
<td></td>
</tr>
<tr>
<td></td>
<td>contents disrupted.</td>
<td></td>
</tr>
<tr>
<td>Most of:</td>
<td>Ashburton</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Timaru</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rangiora</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Amberley</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Hanmer</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Darfield</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Fairlie</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Queenstown</td>
<td></td>
</tr>
<tr>
<td>MM = 8</td>
<td>Alarm may approach panic</td>
<td>Well built but old unreinforced masonry buildings (Type II) damaged, some severely. Some cases of damage to Pre 1970 - 1980 buildings (Type III). Monuments and external tanks fall. Brick veneers damaged, some post 1980. Weak piles damaged and houses not secured to foundations may move.</td>
</tr>
<tr>
<td></td>
<td>Steering of cars greatly</td>
<td></td>
</tr>
<tr>
<td></td>
<td>affected.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Furniture and contents</td>
<td></td>
</tr>
<tr>
<td></td>
<td>damaged.</td>
<td></td>
</tr>
<tr>
<td>Most of:</td>
<td>Greymouth</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Reefton</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Murchison</td>
<td></td>
</tr>
<tr>
<td>MM = 9 or more</td>
<td>General panic.</td>
<td>Well built but old unreinforced masonry buildings (Type II) heavily damaged, some collapsing. Pre 1970 - 1980 buildings damaged, some seriously. Damage and distortion to modern buildings and bridges. Houses not secured to foundations shift off them. Brick veneers fall and expose framing.</td>
</tr>
<tr>
<td></td>
<td>Furniture and contents</td>
<td></td>
</tr>
<tr>
<td></td>
<td>greatly damaged.</td>
<td></td>
</tr>
<tr>
<td>Most of:</td>
<td>Hokitika</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ross</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Hari Hari</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Whatereoa</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Franz Josef</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Fox</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Haast</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Moana</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Otira</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Arthurs Pass</td>
<td></td>
</tr>
</tbody>
</table>

Table 9.3: Average shaking intensities for towns and cities likely to experience significant ground shaking in the next Alpine Fault earthquake. The method of intensity prediction is outlined in Chapter 8. Correlations between Modified Mercalli intensity and damage from Downes (1995).
Inangahua Junction) and the Edgecumbe earthquake (1987, MM = 9 in Te Teko and parts of Kawerau and Edgecumbe).

It is likely a considerable area of MM = 10 shaking will also result from a Magnitude 8 Alpine Fault earthquake but this can not be modelled with any confidence because of the sensitivity of this narrowest isoseismal area to assumptions about fault rupture depth, rupture propagation direction and the total length of rupture. Thrust faults commonly have very high levels of shaking in the hanging wall zone (Kramer, 1996; Archeletta, pers. comm, 1999) and for the Alpine Fault the hanging wall is the steep and elevated Southern Alps area. In general the east dip of the fault is most likely to shift both the epicentre and the area of maximum shaking in the order of 5 – 10 kilometres east of the fault trace.

9.3.3 Earthquake triggered landslides

The analogy provided by the Buller earthquake indicates landslides are likely to be the main impact of an Alpine Fault earthquake on the natural environment, and these will also affect roading, services and property located close to the range front or in other areas of steep topography. The high number of landslides will result from the juxtaposition of the steepest and most elevated area of relief in New Zealand with the epicentral region of strongest earthquake shaking.

Keefer (1984a&b) noted the relative abundance of relatively shallow rock falls and disrupted soil slides during earthquakes. This type of failure will dominate, as it did in the Buller earthquake. Less common are the deeper seated failures, such as the translational block slides which caused so much damage in the Buller earthquake, and the least common are rock avalanches. This study first considers the likely geographic variation in the distribution of all types of landslides, before considering rock avalanches in more detail.
Aggradation impacts associated with landslides are considered later in section 9.4, which deals with the likely long term effects of the next Alpine Fault earthquake.

*Geographic variation in the likely extent of earthquake triggered landslides*

Various magnitude-distance relationships have been derived from landslides triggered in historical earthquakes to estimate the likely distance from the fault rupture that earthquake triggering of landslides can occur. The broadest classification is that of Keefer (Keefer, 1989; Keefer, 1984) which attempts to define the maximum distance various categories of landsliding will be triggered for a given magnitude of earthquake. A similar general relationship has been derived by Yasuda & Sugitana (1988), mainly from the extensive Japanese historical earthquake record. Most recently Hancox et al. (1998) reviewed New Zealand examples of earthquake triggered landslides to test the applicability of these various relationships.

All the general relationships suggest landslides will be triggered at distances as far away from the Alpine Fault epicentral region as 170 km (Banks Peninsula). This also matches estimates which suggest triggering occurs at relatively low intensities of around MM 5 and 6 (Keefer, 1989).

While some triggering at these most distant limits will occur, the total number of landslides triggered at such distances from the epicentre will generally be very small. Of more practical application are the few studies in which the number and/or areal proportion of landslides triggered at particular epicentral distances have been assessed. Unfortunately Hancox et al. (1998) do not quantify this relationship from the New Zealand experience however other internationally derived relationships are available.

Table 9.4 outlines predictions of the extent of landsliding by a range of these methods at various locations. The towns included in Table 9.4 are not intended
as a full list of those locations likely to be affected but provide a range of epicentral distances which to can be used to establish the likely general pattern.

<table>
<thead>
<tr>
<th></th>
<th></th>
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<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Otira</td>
<td>5</td>
<td>80 – 90 %</td>
<td></td>
<td></td>
<td>Destructive</td>
</tr>
<tr>
<td>Franz Josef</td>
<td>10</td>
<td>80 – 90 %</td>
<td></td>
<td></td>
<td>Failure</td>
</tr>
<tr>
<td>Arthurs Pass</td>
<td>10</td>
<td>80 – 90 %</td>
<td>&gt; 60 %</td>
<td></td>
<td>Impact</td>
</tr>
<tr>
<td>Main Divide</td>
<td>1- 20</td>
<td>50 – 80 %</td>
<td></td>
<td></td>
<td>Common</td>
</tr>
<tr>
<td>Mt Cook Village</td>
<td>20</td>
<td>50 – 60 %</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hokitika</td>
<td>35</td>
<td>40 – 50 %</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Greymouth</td>
<td>50</td>
<td>15 %</td>
<td></td>
<td></td>
<td>Failures</td>
</tr>
<tr>
<td>Tekapo Village</td>
<td>60</td>
<td>&lt;15 %</td>
<td>15 – 60 %</td>
<td></td>
<td>Occur</td>
</tr>
<tr>
<td>Cant. Foothills</td>
<td>60 - 80</td>
<td>&lt; 5 %</td>
<td>15 – 60 %</td>
<td>But Most</td>
<td>Impact</td>
</tr>
<tr>
<td>Queenstown</td>
<td>75</td>
<td>Very few</td>
<td>&lt; 15 %</td>
<td>Are</td>
<td></td>
</tr>
<tr>
<td>Buller gorge</td>
<td>100</td>
<td>Very few</td>
<td>&lt; 15 %</td>
<td>Smaller</td>
<td>Scale</td>
</tr>
<tr>
<td>Port Hills, ChCh</td>
<td>130</td>
<td>Very few</td>
<td>&lt; 15 %</td>
<td></td>
<td>Impact</td>
</tr>
<tr>
<td>Banks Peninsula</td>
<td>170</td>
<td>Very few</td>
<td>Very few</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Distance from the long axis of the maximum likely isoseismal

1 Percentage of slope area affected by landslides, including the run out area.

Table 9.4: Comparison of locations within the likely zone of slope failure triggering during an Alpine Fault earthquake. The locations listed fall within the epicentral distance at which landslides are likely to be triggered based on the general relationships of Keefer (1984a&b); Yasuda & Sugitani (1988); and Hancox et al. (1997). Columns three to five review the predicted relative abundance and impact of the landslides at indicative locations by the more detailed methods of Ishihara & Nakamura (1987); Mora & Mora (1993) and ISSMFE (1993).
Ishimara & Nakumura (1987) studied the distribution of slopes which failed during the 1987 Equador earthquake and make predictions based on this on a percentage basis. Mora & Mora (1993) studied slope failures in Costa Rica during 11 earthquakes from 1888 to 1991 and estimated the maximum distances from epicentres for two different categories: areas where more than 60% of slopes failed, and areas where less than 15% of slopes failed but in which at least one slope per square kilometre failed. This latter category is still a large number of landslides but it marks the transition to areas where the impact of landslides becomes more minor.

Because the Mora and Mora (1993) system has just two categories it is not easy to compare with Ishimara & Nakumura (1987). At moderate distances of around 30 - 50 kilometres, Mora and Mora appear to predict considerably more landsliding. However, the Mora and Mora (1993) approach has the advantage of being based on 11 earthquakes, not just one as for Ishimara & Nakumura. These 11 earthquakes span a range of pre-existing climatic conditions and geology and in general it is likely to be a more reliable estimate.

Climatic factors are important in applying these relationships. Section 9.1 outlines how the wet weather prior to the 1929 Buller earthquake is considered to have been a factor which increased the landslide impact. I.S.S.M.F.E. (1993) also reviews existing relationships between epicentral distance and landslides and suggests a distinction be made between wet and dry areas. They adopt two categories; the most severe being "destructive slope failures" and a lesser category of simply "slope failures". This appears to be similar to the 60 % and < 15 % percent category of Mora and Mora (1993) but with a more subjective basis. In applying their method a climate on the wet side of average has been adopted for areas west of the main divide and dry of average on the east. Table 9.4 presents these results and the two categories for the selection of locations.
It appears that no matter what method is used there is consistent agreement that the areas of slope within 30 km of the likely epicentral region of maximum shaking will experience major impact from slope failures. Fortunately the majority of the area likely to be worst affected by landslides, including the rock and debris avalanches discussed earlier, are the Southern Alps which are largely uninhabited. The exceptions are the farms and small towns along SH 73 (Arthurs Pass), SH 6 (Kumara - Haast), and to a lesser extent SH 80 (Mt Cook). While none of the towns listed in Table 9.4 in these areas are built directly on significant slopes many are close to the toe of some very high slopes. The risk at these locations is from debris inundation associated with run out from the landslides.

The larger towns are generally outside the most severely affected region, and some (e.g. Hokitika and Greymouth) have only a small proportion of the town built on, or near, hill slopes. It is likely that landslides will also cause moderate to minor damage to slopes throughout the inland mountain basins (for example the Mackenzie Country, Castle Hill Basin) and also the eastern foothill areas of Canterbury. Although the level of landslide damage on Banks Peninsula is likely to be low, and restricted mainly to minor property damage, the total number of properties affected may be significant due to the higher housing density.

Rock Avalanches

The least common type of earthquake triggered landslide failure on a numerical basis are rock avalanches (Keefer, 1984a&b). These are large scale, rapid and flow-like movements of broken rock downslope and are characterised by relatively long runout distances. They are most frequently triggered by an earthquake (Voight & Pariseau, 1978; Whitehouse & Griffiths, 1983), although it must be noted that the Mt Cook rock avalanches of 1991 & 1992 are the most recent examples of rock avalanches in the Southern Alps which have occurred with no earthquake trigger (McSaveney, 1998).
The best known New Zealand historical example of a rock avalanche in the Southern Alps, and associated with an earthquake is Falling Mountain, near the Aickens Range, which was triggered during the Ms 7.1 1929 Arthurs Pass earthquake (Figure 9.3). This rock avalanche flowed 3 kilometres down the west branch of the Otehake River producing an estimated 88 million tonnes of debris (Cave, 1987; Cave 1998).

Figure 9.3 - The Falling Mountain rock avalanche in the Otehake River valley near Arthurs Pass (K33/020115) which occurred in the 1929 Arthurs Pass earthquake. This type of large-scale rock avalanche is likely to occur in an Alpine Fault earthquake. Photo courtesy of T. Davies and L. Homer.
Whitehouse (1983) compiled information on 46 Holocene rock avalanche deposits in the Southern Alps between Mount Cook and Arthurs Pass. The age estimates for these rock avalanches are based mainly on the weathering rind dating technique for greywacke (Chinn, 1981) which has recently been recalibrated (McSaveney, 1992).

Figure 9.4 - All known rock avalanche locations in the Southern Alps and Westland between Mount Cook and Arthurs Pass. Information from Whitehouse (1983), Whitehouse and Griffiths (1983), Basher (1986), Wright (1994), Chinn, pers. comm. (1996); McSaveney (1998), and this thesis.
<table>
<thead>
<tr>
<th>Locality</th>
<th>Recalculated age in cal. years AD (Nichols pers. comm., 1996)</th>
<th>Possible Alpine Fault event</th>
</tr>
</thead>
<tbody>
<tr>
<td>Craigieburn Range</td>
<td>1700 AD ± 70</td>
<td>Toaroha River event</td>
</tr>
<tr>
<td>Clyde River</td>
<td>1626 AD ± 28</td>
<td>Crane Creek event</td>
</tr>
<tr>
<td>Havelock River</td>
<td>1620 AD ± 90</td>
<td>Crane Creek event</td>
</tr>
<tr>
<td>Cropp River (Basher, 1986)</td>
<td></td>
<td>Crane Creek event</td>
</tr>
<tr>
<td>McTaggart Creek</td>
<td></td>
<td>Crane Creek event</td>
</tr>
<tr>
<td>Lawrence River (Herm. Hut)</td>
<td>1365 AD ± 160</td>
<td>Geologists Creek event</td>
</tr>
<tr>
<td>Geologists Creek (Chinn, 1996)</td>
<td></td>
<td>Geologists Creek event</td>
</tr>
<tr>
<td>Lawrence River</td>
<td>1290 AD ± 180</td>
<td>Muriel Creek event</td>
</tr>
<tr>
<td>Cass River</td>
<td>1200 AD ± 200</td>
<td>Muriel Creek event</td>
</tr>
<tr>
<td>Roundtop Debris Avalanche</td>
<td>950 AD ± 50</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>N. Ashburton River</td>
<td>960 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Rangitata River (Lake Camp)</td>
<td>930 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Rangitata River (Pudding Valley)</td>
<td>880 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Havelock River (Fan Stream)</td>
<td>960 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Rangitata River (Forest Creek)</td>
<td>960 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Godley River (Bloody Point)</td>
<td>880 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Jollie River (Arthurs Pass)</td>
<td>880 AD ± 270</td>
<td>Roundtop event</td>
</tr>
<tr>
<td>Mathias River (Boundary Ck)</td>
<td>1000 AD ± 270</td>
<td>Roundtop event</td>
</tr>
</tbody>
</table>

Table 9.5: Possible correlation between the recalculated rock avalanche ages of Whitehouse & Griffiths (1986) and the inferred Alpine Fault events in this thesis. Ages recalibrated adopting McSaveney (1992) by Nichols, pers. comm. 1996.

Cowan et al. (1996) used a recalibrated data set to demonstrate that in general these rock avalanche ages do not coincide with inferred earthquakes in the Porters Pass - Amberley fault zone and they attribute these to possible earthquakes on faults further to the west. Figure 9.4 shows the distribution of all known rock avalanches between Mouth Cook and Arthurs Pass.
Table 9.5 summarises the age determinations and notes the possible connection between the known rock avalanches and the inferred Alpine Fault earthquake chronology presented in this thesis. Table 9.5 includes the Roundtop debris avalanche near the Toaroha River, which has been recently recognised and described (Wright 1994, 1996), and has been dated in this thesis. This debris avalanche has an estimated deposit volume of 45 ± 28 million cubic metres and an extreme run out of approximately 3.5 km, measured from the base of the range front to the distal end of the deposit. Figures 9.5 and 9.6 show the source area on Roundtop and the runout material.

**Figure 9.5** – The Roundtop debris avalanche. This view, looking south with Roundtop on the left, shows the reforested scarp creating an obvious “hole” in the side of the hill. The run-out material from the debris avalanche forms the hummocky ground at the right, and this extends past the photo right margin to the banks of the current Kokatahi River (see also below). Photograph taken from grid reference J33/550110.
As part of this study an intact section of broken kahikatea log was obtained from the abundant trees scattered among the basal debris exposed in a recent diversion of Cunninghams Creek (G.R. J33/535095). Radiocarbon dates were obtained for sections from both the sapwood, and wood closer to the heart, after first counting the rings bounding these. The possible calendric age is then restricted to the adjusted overlap in the respective calendric ranges and the error is substantially reduced. The age of the deposit becomes 950 AD ± 50 yrs. Bull (1996) previously noted the clustering of eight rock avalanches ages at approximately 1000 years ago (with a mode of 963 AD) and suggests they have been triggered by an Alpine Fault earthquake.

Figure 9.6 – View looking north showing the avalanche debris near the Kokatahi River forming Lake Arthur and adjacent lakes. The very smooth surface of the debris under pasture is a result of recent regrading with a bulldozer by the landowner, Nelson Cook. Photograph taken from grid reference J33/545092.
Figure 9.7 shows the distribution of the known rock avalanches from this time. The location of the Roundtop deposit close to the Alpine Fault trace does provides some additional support for Bull's inference.

Figure 9.7 - Locations of known rock avalanches in the Southern Alps and central Westland which have ages broadly matching the inferred Roundtop Alpine Fault earthquake event of 950 AD ± 50 yrs.
Based on the isoseismal patterns presented earlier in Chapter 8, the locations of rock avalanches for the 950 ± 50 AD event coincide with the epicentral area of Alpine Fault earthquakes. However, this area is also the steepest part of the Southern Alps, where the pre-conditions of rapid and sustained uplift and a dissected mountainous topography also favour their occurrence, regardless of the trigger.

Although the most recent Toaroha River event appears to be poorly represented, the possible connection to this event is hampered by the large radiocarbon age uncertainty relative to the short time interval between this and the penultimate Crane Creek event. The Crane Creek event has four possible matches. The Geologists Creek event, and the preceding more poorly defined Muriel Creek event, each have two possible matches.

If many of the rock avalanches in Table 9.5 were triggered by Alpine Fault earthquakes, and if the relative abundance of the deposits is a true reflection of the shaking intensity, then this would suggest that the largest event of the four was the Roundtop event. This appears to be the best represented despite it also being the oldest. Conversely the smallest, at least in this area north of Mount Cook, appears to have been the most recent Toaroha River event.

Unfortunately the rock avalanche age record will never be a reliable guide to the Alpine Fault earthquake chronology without much higher dating precision. Even with a higher precision other factors such as the pre-existing groundwater conditions, and the time available for weathering since the previous earthquake, will affect the relative frequency of rock avalanches in each successive earthquake.

The Roundtop debris avalanche does indicate a possible hazard which exists along the Southern Alps range front, particularly to property within 500 - 1000 metres from the foot of the mountains. However, even in a large Alpine Fault earthquake, it is unlikely that many long run out landslides on this scale will
occur. It is not practical to exclude development along the range front just in case one does occur, and from a geomorphic and engineering geological perspective, virtually the entire range front from Haast to Haupiri River has the pre-existing conditions necessary to generate one of these large scale failures.

Landslide Dams

Landslide dams have resulted from most of the large historical earthquakes experienced in New Zealand, including the 1855 Wairarapa earthquake, the 1929 Arthurs Pass earthquake, the 1931 Hawkes Bay earthquake, and the 1942 Masterton earthquake. It was noted earlier that numerous landslide dams formed following the Buller earthquake, and the 1968 Inangahua earthquake also triggered landslides in the Buller gorge area, one of which formed a large landslide dam.

The 1968 Inangahua Earthquake triggered landslide contained more than 4 million cubic metres of material which fell from a ridge 600m above the river, crossed the river, and rose 100m up the other side before falling back to form a dam 12 m high (Johnson, 1974; Adams, 1981a). In this case the ponding was not considered dangerous because the dam height was low and when the dam failed 21 hours later no damage was caused by the flood of 1.5 million cubic metres of water travelling at 5.4 m/sec (Sutherland, 1969).

Perrin & Hancox (1992) describe landslide dammed lakes which have survived in New Zealand, many of which are inferred to have been formed during earthquakes, with several examples from Westland and others from east of the main divide. However, most of the landslide dams which survive are the relatively small ones, with water volumes comparable to (or smaller than) that of the landslides which created them.

An Alpine Fault earthquake is very likely to create numerous landslide dams in the 25 - 30 major river valleys draining the western side of the Southern Alps.
Most of these landslide dams are likely to fail quickly due to the high rainfall (i.e. days to weeks) but some may remain for longer periods. In general the steep river gradients will tend to reduce the potential stored water volume for a given landslide dam height. However, the combination of very narrow gorge sections of many of the rivers near the range front, and typical ridge to valley elevation differences in excess of 1000 metres, will generate at least some high dams and correspondingly the potential exists for large volumes of ponded water.

Davies & Scott (1997) have modelled dam break flood hazard from landslide dams in the Callery River, a tributary of the Waiho River near Franz Josef. They conclude there is evidence for such events having occurred in the past and predict dam-break floods in the order of several thousand cubic metres per second. In their opinion the impact of such an event on Franz Josef would be severe and sufficient early warning to allow evacuation can not be guaranteed. They suggest the most realistic mitigation strategies involve restrictions on land-use and development.

On 6 October 1999 a rock avalanche occurred at Mt Adams, in the Poerua River valley, in Alpine Schist rocks approximately 5km south east of the Alpine Fault trace. Although several small earthquakes occurred in the general area 10 days prior to the failure, these do not appear to have been a direct trigger for this rock avalanche (Davies, pers. comm. 1999). The landslide debris formed a dam across the Poerua River around 100m high which in turn created a lake approximately 1.2 km long (volume 5 – 7 million cubic metres). The dam filled for a day before overtopping and the initial overtopping and scour of the crest resulted in some flooding of the river valley and damage to river protection works downstream of the Alpine Fault. Several families were evacuated but after inspection the dam was considered stable at low river flows and most of the residents returned to their houses. It was not possible to predict how long the dam would last during higher river flows.
Five days later the dam breached during rain in the area and an estimated maximum discharge of 500 hundred cubic metres per second of water and debris created significant damage to downstream areas bordering the river. No one was injured and no lives were lost. Further details are included in Appendix 3. This recent event provides an individual example of the type of feature which is likely to form simultaneously, and on a regional scale, following an Alpine Fault earthquake. In many cases this type of dam is likely to form in a series along a river valley, and the additional potential cascade effect that this creates will greatly complicate the assessment of dam-break hazard.

No landslide dams have been reported during historical earthquakes in rivers east of the main divide with the exception of the 1929 Arthurs Pass earthquake. However an Alpine Fault earthquake is likely to cause these. The river gradients are generally less steep, and it possible that although the landslide dams may be fewer in number, the reservoir volumes could be relatively large. Fortunately many of the major river valleys are very wide in most places. However landslide dams may also form in the gorge sections of the major rivers close to the Canterbury Plains, with a corresponding flood hazard lower in the river. While it is statistically unlikely that an earthquake will coincide with heavy rain, the majority of subsequent landslide dam failures occur when rainfall is extreme.

In section 9.4 the long-term impacts of river aggradation are noted, including the potential effects on towns and vulnerable transportation routes adjacent to rivers, and effects on hydro-electric projects.

9.3.4 Liquefaction

Liquefaction results from the earthquake induced consolidation of loose saturated sand and silty sand sediments which, when vibrated, expel porewater at a rate faster than they can drain. As a result the porewater pressure
exceeds the overburden weight and the sediment behaves as a liquid and therefore has no shear strength. Foundations sink into the soil and sloping ground at river, lake and lagoon margins tends to flow towards the lowest area (lateral spreading).

The important role of liquefaction in earthquake damage has been increasingly recognised since 1964 when both the Niigata, Japan and the Anchorage, Alaska earthquakes drew widespread attention to the phenomena. Considerable research followed (for example Seed & Lee, 1966; Yamada, 1966; Seed & Idriss, 1971; Long, 1973; Castro, 1975; Peck, 1979; Davis & Berrill, 1982; Youd, 1984) and predictive methods to assess liquefaction susceptibility rapidly advanced (Ishihara, 1985; Iwasaki, 1986; Liao et al. 1988; Davis & Berrill, 1996; Berrill, 1997).

Preconditions which favour liquefaction include:

- Geologically young sediments of a Late Holocene or Recent age (these are often uncemented and relatively unconsolidated)

- Sediments with a low bulk density and shear strength

- Sediments with an abundance of clean fine - very fine sand and coarse silt

- The presence of a high watertable (implying either a location close to river and sea level and/or an area with a high rainfall).

The lower floodplains of most west coast rivers generally match all of these preconditions and many of these areas are vulnerable to liquefaction. Higher river terraces (which tend to be slightly older and have lower watertables), and the coarser grained sediments of the upper catchments, are normally less susceptible. Moraine, alluvial fans and colluvium slopes on bedrock are also unlikely settings for liquefaction damage. The most at risk locations are close
to the mouths of the major rivers, particularly the lagoonal, lake delta and estuarine areas, and also parts of the dune and beach area. Historically the river mouth sites offered potential for ports and river transport. As a result significant areas of the largest west coast towns such as Karamea, Westport, Greymouth and Hokitika, have a geological setting which makes them potentially susceptible to liquefaction.

**Historical examples**

Liquefaction has occurred in all the west coast historical earthquakes, even the relatively small ones (Table 9.6). To date Karamea, Westport and Greymouth have been the closest towns to the earthquake epicentres.

<table>
<thead>
<tr>
<th>Earthquake and Magnitude</th>
<th>Summary of damage</th>
</tr>
</thead>
<tbody>
<tr>
<td>1913 Westport M = approx. 6</td>
<td>Cracks in the ground and ejection of mud at Cape Foulwind.</td>
</tr>
<tr>
<td>1929 Buller M = 7.8</td>
<td>Fissures and subsidence and water spouts in Karamea, Westport, Greymouth, Inangahua and Greenstone.</td>
</tr>
<tr>
<td>1968 Inangahua M = 7.1</td>
<td>Sand boils, fissures and subsidence. Sand boils sufficient to lift and rotate one house in Westport on its foundations (Romilly Street).</td>
</tr>
<tr>
<td>1991 Westport M = 6</td>
<td>Minor liquefaction at Nine Mile Beach, Charleston causing subsidence and sand boils.</td>
</tr>
</tbody>
</table>

Table 9.6 Historical examples of liquefaction on the west coast during earthquakes (from Fairless & Berrill, 1984; Berrill et al., 1988 and Benn, 1992).

Other east coast South Island historical examples of liquefaction have been reported from the lower Wairau River (both the 1848 and 1855 earthquakes); Hanmer basin (1888 Amuri earthquake); Kaipoi, Hurunui River and Cheviot (1901 Cheviot earthquake); and Lake Sumner (1929 Arthurs Pass earthquake).
Generally liquefaction presents the greatest hazard in areas where liquefaction in sloping ground makes lateral spreading (low angle landsliding) possible. The common juxtaposition of vulnerable saturated sediment and moderate ground slopes at river and lake margins make these areas particularly vulnerable to lateral spreading. In Westland the most common structures likely to be damaged in such locations are bridges and road embankments.

The likely extent of liquefaction in susceptible sediments

The furthest liquefaction has been recorded from an earthquake epicentre on the west coast is 122 km (Greenstone River, during the 1929 Buller earthquake; Benn, 1992). This compares with the maximum New Zealand recorded example of 193 km during the M = 7.1 Marlborough earthquake of 1848. The larger 1855 Wairapa earthquake induced liquefaction up to 160 km from the epicentre but the country was sparsely populated and other examples may not have been recognised (Fairless & Berrill, 1984).

Overseas examples of liquefaction have been analysed and a relationship derived between magnitude and epicentral distance. Kuribayashi & Tatsuoka (1975) considered 32 historic Japanese earthquakes and suggest a Magnitude 8 earthquake on average would produce liquefaction of susceptible sediments at distances of up to approximately 250 kilometres. Berrill & Fairless (1984) review the historical record of liquefaction in New Zealand and compare this with the relationship between earthquake magnitude and epicentral distance of Kuribayashi & Tutsaoka, 1975. In so doing they discount some of the most extreme New Zealand examples but conclude that within the limited data set this relationship appears to also hold for New Zealand conditions.

Ambraseys (1988) used an enlarged international data set and derived a relationship which extends the predicted distance over which liquefaction will occur to approximately 350 km for M = 8. More recently Wakumatsu
(1993;1991) showed that at high magnitudes (such as M = 8) the Ambraseys relationship is appropriate for Japanese conditions only if minor signs of liquefaction are included. However if the data-set is restricted to cases of significant (or potentially damaging) liquefaction, then 250 km is the more appropriate limit.

Hancox et al. 1997 reviewed 22 historical New Zealand earthquakes and found the general threshold for liquefaction was MM 7 for sand boils (sand and water ejections) and MM 8 for fissuring and lateral spreads, but noted that both may occur at one intensity unit lower in highly susceptible soils. Figures 8.2 & 8.3, the Smith 1995a & b isoseismal constructions for the most recent two Alpine Fault earthquakes, indicate that MM 8 and MM 7 shaking in an M 8 Alpine Fault earthquake is likely to extend approximately 85 km and 165 km respectively from the epicentre.

The locations in the South Island of main interest can be divided into four groups based on their distance from the likely epicentral region (refer Table 9.7). Applying the various observations of the liquefaction damage in historical earthquakes from both New Zealand and overseas discussed above (i.e. Kuribayashi & Tatsuoka, 1975; Ambraseys, 1988; Wakumatsu, 1991, 1993; Hancox et al., 1987) some general comments about areas of potential damage can be made (Table 9.7).

**Defining susceptible soil**

The previous discussion demonstrates that historical precedent suggests liquefaction is likely to be a significant hazard in susceptible soils at locations in the first three groups of Table 9.7. While in many cases susceptible soils are not proven to be present at the sites listed, these locations are all areas where the geological setting suggests this is a strong possibility. The list in Table 9.7 includes examples from coastal areas; lake margins, swamps and
river margins; and/or areas that have experienced liquefaction in previous earthquakes.

<table>
<thead>
<tr>
<th>Group</th>
<th>Towns &amp; locations, with distances*</th>
<th>Likely extent of liquefaction in susceptible sediments</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Group one</strong></td>
<td>Locations such as Okuru (15 km); Paringa (15 km); Okarito (30 km); Kokatahi (20 km); Hokitika (30 km); Rotomanu (20 km); Ivieagh Bay (20 km); Greymouth (50 km)</td>
<td>Extreme liquefaction in susceptible sediments. At the very closest distances some liquefaction possible in more gravel rich sediments not normally considered liquefiable.</td>
</tr>
<tr>
<td>&lt; 50 km</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Group Two</strong></td>
<td>Lakes Coleridge, Pukaki, Tekapo and Ohau (50 - 60 km); Inangahua valley (50 - 80 km); Queenstown (75 km);</td>
<td>Liquefaction of loose lake sediments and some alluvial material is likely.</td>
</tr>
<tr>
<td>50 – 85 km</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Group three</strong></td>
<td>Westport (115 km); Kaiapoi (115 km); Christchurch (115 – 130 km); Ashburton (110 km); Timaru (140 km).</td>
<td>Damaging liquefaction is likely in the most susceptible estuarine and fluvial sediments</td>
</tr>
<tr>
<td>85 – 165 km</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Group three</strong></td>
<td>Karamea (175 km); Dunedin (240 km); Blenhiem (250 km); Nelson (230 km).</td>
<td>Liquefaction in extremely susceptible sediments may occur at closest locations within this range, but probably only of limited local impact.</td>
</tr>
<tr>
<td>165 – 250 km</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* distance as measured from the long axis of the likely maximum isoseismal

**Table 9.7:** Locations where liquefaction may occur in susceptible sediments during a Magnitude 8 earthquake on the Alpine Fault.
Research on the role of depositional environment in controlling liquefaction susceptibility (for example Tinsley & Dupre, 1992; Ishihara, 1993) indicates that sediment which has accumulated in fluvial point bars is normally the most vulnerable.

Site specific subsurface investigation is required to fully assess liquefaction potential and this includes the determination of soil size grading; density, and depth to watertable. CPT testing has progressively replaced the earlier SPT based site assessments and piezocone testing has been shown to be an excellent test method (Berrill et al., 1992; Dou Yiqiang & Berrill, 1993; Vreugdenhil et al., 1994).

This type of detailed site specific investigation has been carried out previously in a few locations on the west coast but only in the Buller region (Ooi, 1987; Bienvenu, 1988; Berrill et al. 1987). This has confirmed that liquefiable soils are common in both the coastal setting and the margins of the Buller and Inangahua Rivers. Testing has shown some sites remain vulnerable despite undergoing repeated liquefaction in both the 1929 Buller and 1968 Inangahua earthquakes.

Christchurch is the only significant population centre in Table 9.7 where much site specific investigation information on liquefaction vulnerability has been gathered (see for example McAhon et al. 1992; Berrill et al. 1994; Berrill, 1997; McAhon et al., 1997). As part of the recent Christchurch engineering lifelines project a total of 16 sites which are principle nodes in the lifelines networks in eastern Christchurch have been investigated. Of this 16, a total of 12 were found to be potentially liquefiable (Berrill, 1997). There is no reason to indicate that this is not a representative sample of the susceptibility of the sand sediments of eastern Christchurch.

If liquefaction potential is recognised at a site there are a range of mitigation options available. These include piling the foundations, densification or
replacement of the soils prior to construction, permanently lowering the water table, or avoiding the area. For many existing buildings it is frequently not practical to reduce the hazard but for critical structures (hospitals, police stations, and key utilities) remedial work may be justified.

In summary liquefication is likely to be a significant cause of damage in the next Alpine Fault earthquake. This will be particularly serious in the epicentral area, but because of the population distribution, the greatest material damage may be in densely developed vulnerable settings at relatively distant locations such as Christchurch.

9.3.5 Tsunami & seiches

Tsunami are seismically generated sea waves produced by submarine faulting or landslides, or where surface landslides enter a body of water. They travel very quickly in the open ocean (velocities in the order of 700 km/hr, Hunt [1983]) with little surface expression, but gain in wave height near shore, particularly along gently shelving coastlines and within narrow coastal embayments. The continental shelf of Westland is narrow and no significant open bays exist, although there are many river mouth settlements. In general the Alpine Fault in central Westland is not expected to produce regional submarine uplift or deformation. It is still possible submarine or coastal landslides could be triggered which may produce local tsunami.

In Fiordland, the Alpine Fault continues offshore southwest of Milford Sound. Submarine fault rupture in this area of numerous narrow fiords could create significant tsunami in Fiordland. However, the fiords tend to be steep sided and in most cases only the head of these embayments would be vulnerable. Most of these areas are uninhabited but this could potentially affect the developed area of Milford Sound.
Strong seismic shaking also induces water in lakes to oscillate (or "slop") at a particular frequency controlled by the lake size and depth. These oscillations, referred to as seiches, were generated on the west coast during the 1929 Buller earthquake.

Benn (1992) quotes from the Greymouth Evening Star (20.6.1929) an account of damage to Lake Rotoroa and the Gowan River, which drains the west end of the lake:

"Lake Rotoroa rocked from side to side like a huge basin of water being tipped out. Half an hour after the main shake, the water receded from the hotel shore and exposed the lake bed for 50 yards. It then came back in a series of large waves. The bridge over the Gowan River at the lake was torn from its piles and the banks of the river and was hurled upstream. The wrecked structure was carried still further upstream by the Gowan waters, which were temporarily flowing back into the lake."


Lake Moana (Brunner) was also reported to have "sank down in the middle then come up like a typhoon" during this earthquake.


In general lake seiches are more likely to cause damage during an Alpine Fault earthquake than are tsunami. These lakes are closer to the area of strongest shaking and there are a large number on both sides of the main divide. On the west coast these include several small northern lakes (i.e. Lady & Kangaroo Lake and Lakes Haupiri, Ahaura and Hochstetter) but also include the much larger and relatively populated Lake Moana and Lake Kaniere, both with lakeshore settlements and dwellings. There are also numerous lakes and lagoons further south along SH 6 with generally less potential for material damage (for
example Lake Mahinapua, Lake Ianthe, Saltwater Lagoon, Lake Wahapo, Okarito Lagoon, Lake Mapourika, Lake Matheson; Lake Paringa, Lake Moeraki and many others).

Of more concern than a shaking induced seiche for Lake Brunner and Lake Kaniere, is the more remote possibility of a significant landslide falling into one of these lakes, and thereby creating a damaging water wave. Both these lakes are very close to the fault trace (7 km and 500 m respectively), and both have steep peaks along their shorelines which reach elevations of more than 1000 metres above lake level (i.e. Mt Te Kinga at Lake Brunner, elevation 1204m and Mt Tuhua at Lake Kaniere, elevation 1125m). Hawley (1984) and Benn (1992) have also noted this possibility.

East of the main divide there are the large moraine dammed lakes of the Mackenzie Basin (i.e. Lakes Tekapo, Pukaki, Ohau) as well as Lake Benmore, which is impounded behind a hydroelectric dam. All four are close to the likely epicentral region of Alpine Fault earthquakes and seiches are likely to be generated in these lakes. The extent of the seiches will be controlled by the lake shape, depth, proximity to the epicentre, and the frequency of the earthquake waves. However seiches in these areas could have impacts for the hydroelectric generation facilities on these lakes.

The larger southern lakes (i.e. Lakes Wanaka, Hawea and Wakatipu), are all within 100 km of the fault trace, but may also experience seiches, although probably to a lesser extent.
9.4 LONG-TERM EFFECTS

This section considers the various long-term impacts which may follow an Alpine Fault earthquake. The time frame for these impacts ranges from a matter of days following the main shock, to years or possibly tens of years after the earthquake event.

9.4.1 Effects on forests and vegetation

Chapter 5 outlines the evidence preserved in the age structure and growth rings of the current forest which indicates the impact of past Alpine Fault earthquakes. Vegetation in general will be killed or damaged immediately by strong shaking, landsliding, and liquefaction; and more slowly in the post-seismic period by root damage and subsequent rainfall triggered landsliding, aggradation, river channel avulsion and flooding.

Pearce & Watson (1986) document the burial of forest on valley floors by the Buller earthquake. More recently Grapes & Downes (1997), quoting newspaper accounts from the time, note that the 1855 Wairarapa earthquake removed nearly one third of the vegetation on the western face of the Rimutaka Range as viewed from Wellington. McKay (1901) describes landslides during this earthquake which:

"carried away hundreds of acres of bush, burying it and piling it in utter confusion"

McKay (1901), page 33.

Grapes (1988) looked in detail at the vegetation remaining on one large landslide from the 1855 earthquake and concluded only one pre-earthquake tree has survived on the slide block.
Aquatic vegetation was also affected by seiches associated with the 1855 Waiararapa earthquake. Grapes and Downes (1997) outline reports of raupo, flax and toi toi torn up by the seiche wave action and left floating on Lake Waiararapa.

9.4.2 Effects on rivers

The earlier Chapter 4 included a discussion of the typical impact of earthquake triggered landslides on the sediment balance of a river. In the longer term the impact of an Alpine Fault earthquake on the behavior and character of rivers may prove to be one of the most significant overall effects.

It is likely to induce a progressive "wave" of aggradation with associated channel avulsion, local bank erosion, and flooding which will move down through the river systems on both sides of the main divide. The greatest impact will be on the rivers draining the west coast because these have the largest flows, steepest catchments, and the largest current sediment budgets. The distance down river that aggradation will extend will vary from catchment to catchment and be controlled by the proportion of landslides in the upper catchment with respect to the average river flow.

In addition aggradation is normally most severe in the sections of river dominated by bedload, as opposed to suspended load. This is generally the steepest gravel dominated sections of river close to the main divide and the range front. Even relatively small streams which drain steep slopes, and which are within or near the epicentral region, are likely to experience rapid fan building and associated channel avulsion. The impact is likely to be the least in areas near the coast and relatively far from the mountains where river gradient is lowest and suspended load is the dominant transport process. But severe aggradation may ultimately lead to channel avulsion and relocation of the river.
In general the rivers which give the greatest problems under the present conditions are also likely to be the most sensitive to the increased sediment budget. Obvious examples include the Waiho River, which is aggrading rapidly at present probably due to glacial fluctuation (Davies, pers. comm., 1997); the Waitangitaonga River near Lake Wahapo (Griffiths & McSaveney, 1986); and possibly the Grey and Hokitika Rivers in sections where flooding has been a historic problem.

**Figure 9.8** - The Waiho River at Franz Josef (GR H35/817533) which has had a history of flooding and aggradation since settlement of the district. The old bridge has had to be lifted to maintain the waterway, while a new temporary bridge has been put in immediately upstream. Aggradation such as this is likely in many of the rivers draining the Southern Alps following an Alpine Fault earthquake.
Figure 9.9 - The Toaroha River immediately downstream of the Alpine Fault trace (GR K33/573100, refer also Figure 2.2 in back pocket). Note that the current bed level is such that coarse gravel bed load is currently being deposited on terrace surfaces which were previously only reached by suspended sediment. Locations such as this are extremely vulnerable to any future increase in bed levels as a result of earthquake induced aggradation.

An example of potential channel avulsion – the Taramakau River at Inchbonnie

The Taramakau River could ultimately shift channel at Inchbonnie in response to earthquake induced aggradation with very serious consequences for Greymouth and areas of the Grey valley. Figure 9.10 is a map of the middle and lower Taramakau River catchment, including Inchbonnie. Figure 9.11 is an
aerial photograph of part of this same area. At various times the Taramakau river has flowed northwest, straight through to the south end of Lake Brunner, and at other times northeast around through the current Lake Poerua, to the north end of Lake Brunner.

During detailed field investigation of the Alpine Fault trace at Inchbonnie as part of this thesis I collected a sample of buried grass from a 1.2 m deep excavation near the Inchbonnie Recreation Centre (Figure 9.11). This building is on the Inchbonnie - Rotomanu Road, which is north of the current Taramakau channel, and within the old river route to the south end of Lake Brunner. This grass had been growing on a surface of very coarse, well rounded, greywacke cobbles from the Taramakau River, which is the only source of greywacke amongst the nearby schist and granite slopes. The grass had then been rapidly buried by over one metre of uniform sandy silt, typical of the overbank flood material of a large river. This grass sample was radiocarbon dated (Wk 5509) as “modern”, indicating river occupation of this relatively low area since approximately 1700 AD. Eroded into the surface of this overbank silt are shallow braided paleo- river channels which lead to Lake Brunner, and which are still clearly visible in modern aerial photographs of the area (Figure 9.11).

The recent occupation of these alternative channels by the Taramakau River is supported by local farmers accounts that the old saw-millers reported that the really large trees in the Inchbonnie area were only found growing above the “terrace” (the Alpine Fault scarp). The main consequence of water from the Taramakau River entering Lake Brunner from either of the possible alternative flood routes is that this water will then drain down the Arnold River to the Grey River, which has a history of serious flooding at the coastal town of Greymouth.
Figure 9.10 – The middle and lower Taramakau River catchment with Lake Brunner, and the lower Grey River catchment, also shown. Two potential alternative flood routes of the Taramakau River in response to earthquake induced aggradation are marked by arrows. The most direct route, from Inchbonnie north to Lake Brunner, has a gradient which is significantly steeper than the current river bed. The second alternative route, from Inchbonnie northeast via Lake Poerua to Lake Brunner, has a gradient only slightly less than the current river bed. If the river does change course, then water from the Taramakau River will enter the Grey catchment, with major potential consequences for the flood prone coastal town of Greymouth.
Figure 9.11 – An aerial photograph of the area around Inchbonnie (refer Figure 9.10 for location details) with the current Taramakau River just visible flowing west to Kumara at the left hand side of the photograph, and Lake Poerua in the centre. Note the paleo-channels, which are visible in the air photo extending towards the top right corner, leading on to Lake Brunner (which is another 4 km from the top photograph margin). There are also paleo-channels swinging northeast via Lake Poerua to Lake Brunner. The white pointer shows the location of a buried grass sample, obtained as part of this thesis, 1.2 m below the current ground surface. This returned a post 1700 AD date, and indicates that the Taramakua River has flowed along these alternate routes within the last 300 years. Scale 1:30,000 approximately from SN 8312 K8,1984. Reproduced courtesy of Aerial Mapping Ltd.
There is only a low probability that an Alpine Fault earthquake will coincide with high river flows and earthquake induced aggradation will take several weeks to develop following an earthquake. However, any abrupt change in river route is likely to occur during a flood event, and therefore coincide with high flow in the Arnold and Grey River. The Taramakau River flow will be very difficult to reverse once a new channel has become established, particularly at a time when resources are stretched by the regional impact of an Alpine Fault earthquake. It would be prudent to progressively implement river protection works in this area as soon as possible.

Because of the narrow width of central and south Westland, and the limited local population, there is generally only one road, power and telephone line through most of the district. These services traverse hundreds of small river fans and scores major rivers, most of which drain the Southern Alps. The Inchbonnie location described here is just one example of many locations where the consequences of river avulsion following aggradation are likely to be serious for the local towns and settlements, and to have a major impact on the communication, transportation and lifeline infrastructure of the district.

9.4.3 Effects on the coast

The coastline in Westland is likely to accrete rather than erode in response to the sediment pulse associated with an Alpine Fault earthquake. This will increase both the volume of material carried northward by littoral drift, and the sediment discharge from each individual river mouth. River bars may tend to enlarge in response to this but the interaction is likely to be complex and difficult to predict.

For example Hokitika has a history of cyclic coastal erosion of the area immediately north of the river mouth. Gibb (1987) noted that the erosion is controlled largely by the orientation of the river mouth within the coastal delta. If the river mouth and associated river discharge is offset to the south, then
there is generally erosion immediately to the north of the mouth, which affects the town. It is difficult to predict the influence that earthquake generated sediment pulses will have on river mouth alignments.

9.5 SUMMARY AND CONCLUSIONS

An Alpine Fault earthquake will have effects similar to the 1929 Buller earthquake, but on a larger scale. Immediate effects of the earthquake will include:

- Rupture of the fault trace where this crosses State Highway 6 and several other more minor roads. Franz Josef is the only town where rupture of the active fault trace will cause direct damage.

- Ground shaking, which will be very strong along the west coast and in the central Southern Alps. This shaking will cause structural damage to buildings and infrastructure. Strong to moderate shaking will occur in most South Island locations within 150 kilometres of the Alpine Fault.

- Landslides will be triggered by ground shaking over a very large area. The largest landslides will occur in the central and western Southern Alps, where it is likely several rock and debris avalanches will occur.

- Landslide dams will be formed simultaneously in many river valleys. While most will fail relatively quickly, many will take weeks or months to clear. Some may briefly form again in response to aftershocks and subsequent rainfall triggered failures of the earthquake weakened slopes.

- Liquefaction in susceptible sediments will cause significant damage to buildings, services, and transport routes up to 150 kilometres from the fault, and liquefaction in some very weak materials will as far away as 250 kilometres from the fault. Susceptible sediments are likely to exist along
river, lake and swamp margins in the epicentral area, including coastal and river mouth locations such as Greymouth and Hokitika. Because of the population distribution in the central South Island, the greatest material damage from liquefaction may occur at more distant populated locations in vulnerable settings such as Christchurch, which is 130 km from the likely epicentral region.

- Seiches (high waves) will be induced by the earthquake shaking in many lakes in the general area, including the Southern Lakes and the various lakes in the Mackenzie basin.

One of the important differences between an Alpine Fault earthquake, and other earthquakes that have occurred in the area since European settlement, is the large scale regional impact associated with an Alpine Fault event. The simultaneous regional impact will be compounded by access difficulties for civil defence and emergency resources which will be created by the damage to roads and bridges, and the steep and rugged topography of much of the area.

**Longer-term effects** will continue for many months and years following the earthquake. The most significant post-seismic effect is likely to be river aggradation in response to the sediment overload from landslides. This will increase flood risk in many locations for many years after the earthquake. This will impact on both the local population, as well as the transport, communication and lifeline network in many areas of the central South Island and Westland. It will also affect the economy of those areas which are generally dependent on tourism and farming.

Another long-term effect is the likely impact on the extensive natural forest and vegetation of the region. A large amount of forest and vegetation will be killed or damaged during the earthquake by the strong shaking, landsliding, and liquefaction. The immediate damage will followed by:
• delayed impacts of root disturbance and branch loss
• post-seismic rainfall triggered landslides of the weakened ground
• river avulsion and flooding in the post-seismic period

The combination of these impacts on the existing forests of the conservation estate will be profound, although the previous earthquake record suggests this will be followed by a phase of simultaneous regeneration in the years that follow.
CONCLUSIONS AND RECOMMENDATIONS

9.1 CONCLUSIONS

1. The Alpine Fault is not "aseismic" and produces large earthquakes with associated liquefaction at the fault trace. This is demonstrated for the first time by features in the Kokatahi 2 trench excavated as part of this thesis.

2. Four independent lines of evidence are presented in this thesis which are consistent and indicate that two major Alpine Fault earthquakes have occurred in the last five hundred years. These lines of evidence are summarised below in Figure 10.1.

3. The most unequivocal and direct evidence comes from trenching of the fault. At three locations south of the Haupiri River trenching investigations indicate the most recent rupture event occurred post 1660 AD and probably between 1700 AD and 1750 AD. There is also evidence in one of these trenches (Kokatahi 1) for a second earlier rupture event post 1470 AD.

4. At two more northern locations near the Ahaura River, trenching indicates the last rupture event occurred between 1480 AD and 1645 AD. There is no evidence of the most recent post 1660 AD event.
Figure 10.1 - A summary of the four lines of independent evidence used in this thesis to establish the timing and extent of the most recent two Alpine Fault earthquakes.

5. In this type of steep terrain large earthquakes will trigger many landslides and generate river aggradation. The record of radiocarbon dated landslides and aggradation terraces in Westland over the last 750 years has been significantly increased as part of this thesis from five to a current total of nineteen. The data also suggests two earthquakes at around 1600 AD, and post 1700 AD, and are consistent with the trench rupture date ranges.

6. Indigenous forest in Westland exhibit age modes consistent with common forest re-establishment following massive disturbance events. Two regional
disturbance events are apparent in the last 500 years and earthquakes have been suggested independently by plant scientists as a possible explanation of the pattern. The inferred timing of the most recent disturbances at 1625 ± 15 years AD, and 1715 ± 15 years AD, matches the date ranges from the trenches, and also the apparent clusters of radiocarbon dates from the landslide and aggradation terraces.

7. Trees which have continued growing through these inferred earthquake events commonly exhibit abrupt fluctuations in tree ring width at dates which are once again consistent with the trench date ranges, the landslide and terrace age modes, and the forest disturbance record. These tree ring chronologies are the best method to estimate the exact timing of these events.

8. I infer the most recent earthquake occurred in 1717 AD and is referred to here as the Toaroha River event in view of it's initial recognition in the Toaroha River trenches. The associated disturbance can be traced in cross matched tree ring chronologies from Lake Te Anau to Hokitika, a distance of more than 400 kilometres. The disturbance is absent from available tree ring chronologies further north at the Ahaura River and the Rahu Saddle, which is consistent with the evidence from trenches which indicates the last rupture terminated slightly north of the Haupiri River.

9. The penultimate earthquake event is referred to here as the Crane Creek event and can be recognised in the forest disturbance record from the Rahu Saddle in the north through to the Copland Valley and Karangarua River and areas as far south as the Paringa River. We infer this earthquake occurred in 1620 AD ± 10 years and the forest disturbance record indicates a minimum rupture length of 250 kilometres.

10. Other older events can also be inferred from the available data, including the results of other paleoseismic investigations at the Waitaha River in
central Westland (Wright, 1999; Wright et al., 1998). The combined data suggests other earthquakes at around 1425 AD ± 15 yrs; 1220 AD ± 50 yrs and 960 AD ± 50 yrs.

11. The timing of previous earthquakes can be used to estimate the probability of the next earthquake occurring in the central section of the fault using a range of methods. The most basic method assumes random earthquake occurrence and estimates a 50-year probability of rupture of around 20% and a 100-year probability of rupture of around 40%.

12. Other methods assume a seismic cycle of progressive stress accumulation with a steadily increasing probability of subsequent earthquake release. Methods of this type which use only the inferred Alpine Fault recurrence sequence are hampered statistically by a lack of data, but suggest a higher 50-year probability of rupture of around 30% (range 16 – 50%) and a 100-year probability of rupture of approximately 70% (range 50 – 85%).

13. The most realistic and reliable method of probability assessment assumes that the recurrence behaviour of the Alpine Fault will resemble that of other plate boundary faults around the world. Although successive recurrence intervals will vary significantly about a mean, enough intervals exist in this international dataset to statistically predict the probability of variation. When this method is applied to the Alpine Fault the probability estimates are much higher, with a 50-year probability of rupture of 65% (range 50 – 80%) and a 100-year probability of rupture of 90% (range 80 – 95%).

14. Mechanical considerations of crustal strength, and observations of high historical strain rates, also suggest a finite limit to the possible future holding time for the central Alpine Fault in the order of 100 years.

15. The next Alpine Fault earthquake is likely to resemble the previous two earthquakes and based on this a range of likely shaking intensities can be
predicted for the central South Island. The pattern suggests that the highest likely Modified Mercalli Intensities of MM 9 or greater will be experienced in the Southern Alps and areas of central Westland. Shaking will also be strong in most west coast towns (with the possible exceptions of Westport and Karamea), and in areas of the east coast foothills.

16. Recent revisions in attenuation relationships suggest the shaking in more distant locations, such as Christchurch, will not be as severe as some previous predictions. However, the level of shaking in Christchurch is still likely to be the highest experienced in the area in the last 100 years. The likely long period and duration of the earthquake shaking will increase the likelihood of liquefaction, and the frequency of shaking will particularly affect taller buildings.

17. Landslides, including possible rock and debris avalanches, will be common in the Southern Alps and other steep slopes in Westland. Analysis of the record of previous rock and debris avalanches suggests many of these old features may have been triggered by previous earthquakes on the Alpine Fault, in the particular the Roundtop event of around 960 AD, but the dates of many of these rock avalanches have a poor resolution. Landslides on the Port Hills and Banks Peninsula will be minor in comparison to the Southern Alps area but may still affect a significant number of properties.

18. Landslide dams will be created in many river valleys on both sides of the main divide. While these may create a short-term flood risk most will fail relatively quickly. The recent Poerua River landslide dam of 6 October 1999 provides a reminder of the hazard created by a single landslide dam. Following an Alpine Fault earthquake this problem is likely to occur simultaneously in many widely separated catchments at a time when the emergency management systems are already under pressure.
19. Based on previous New Zealand and international earthquakes liquefaction is predicted to occur in susceptible sediments as far away as 200 km from the epicentral region. The towns potentially most susceptible to liquefaction are Greymouth and Hokitika (both of which are within 50 km of the likely epicentral area) and the city of Christchurch (130 km).

20. Large water waves (seiches) are likely to result from oscillation of water in lakes on both sides of the main divide in the areas of strongest shaking.

21. Vegetation damage following the Buller earthquake of 1929, and other overseas earthquakes in steep forested terrain, suggests that in the short and medium-term the landscape and vegetation impact on the indigenous forests of the Department of Conservation estate will be profound.

22. Aggradation and channel shifting following the earthquake is likely to occur in the upper reaches of most of the rivers and streams near the epicentral area on both sides of the main divide. In some cases these impacts may extend progressively down river to affect more populated areas at the river mouths.

9.2 **GENERAL RECOMMENDATIONS**

1. Potentially affected local authorities and infrastructure providers should incorporate the potential consequences of an Alpine Fault earthquake in their planning for the next 50 - 100 year time period.

2. This requires a short-term immediate post earthquake contingency plan that will apply for an initial period of minutes and continue up to several days after the earthquake.
3. Assumptions to be made in this short-term planning include:

- No vehicular access on SH 73 between Springfield and Kumara.
- No vehicular access on SH 6 between Ross and Haast.
- Possible loss of vehicular access between Lake Pukaki and Mount Cook Village.
- Loss of power and fixed telecommunications in this same general area.
- Disruption to utilities (water, power, sewer, telephone) and possibly to emergency service facilities in Hokitika and Greymouth.
- Liquefaction damage to Greymouth airport however Hokitika airport is more likely to remain serviceable.
- A short term flood hazard from breaching landslide dams on many west coast rivers (and possibly in some of the gorges of large east coast rivers) immediately after the earthquake, and especially in the first few periods of rain.

4. Planning is also required to attempt to minimise the longer-term impacts of the earthquake that can often be the most profound.

5. Such planning should allow for severely restricted vehicular access for at least several months following the earthquake.

6. There will also be only limited power generation capacity for several months on the west coast, and disruption to the power transmission network on both sides of the main divide. The power generating capacity in the Mackenzie
basin, Upper Waitaki and also Lake Manapouri may also be disrupted (the later depends on the southern limit of fault rupture in the next event).

7. River behaviour is likely to change abruptly, with aggradation of several metres in the upper reaches of most rivers on both sides of the main divide. This will present a long-term problem for bridges and land-use in these areas.

8. Over time the upstream changes in river behaviour may progressively affect communities at river mouth locations as increased sediment load affects the lower reaches of the larger rivers. This may also include the large east coast rivers draining the Southern Alps such as the Rakaia, Rangitata and Waimakariri Rivers.

9. Both the short term and long term impacts will be progressively less severe in more distant locations away from the epicentral regions. However, an Alpine Fault earthquake will be a regional event creating simultaneous damage over a much wider area than any other New Zealand earthquake since settlement of the South Island. This geographic spread, in conjunction with the rugged natural terrain, will present a severe challenge to emergency management and post-earthquake reconstruction systems.

10. Profound short-term economic impacts are likely for both Westland and Canterbury, however, in the long-term the impact of post-earthquake reconstruction and investment may be positive.
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Finally I would like to thank my wife Lisa Yetton, and my children Ben and Renee, for helping in a million ways, and putting up with it all over the last five years.
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APPENDIX ONE

CALENDRIC CONVERSION OF RADIOCARBON DATES

As noted in the footnote on page 44 the relationship between radiocarbon "years" and calendric years is not linear and only roughly approximates to a 1:1 relationship. Radiocarbon dates (or conventional dates are they often referred to) are in radiocarbon years before 1950 when the testing of nuclear bombs first influenced the global levels of $^{14}$C around the world. Figure A1 shows the calibration curve for the last 600 years from Stuiver & Becker (1993), which is the period of most interest in this project.

Note the "wiggles" in the curve in the last 450 years. These result in separate possible date ranges for some of the samples plotted in Figure 4.8 on page 63. Note also the wide fluctuations in the curve since around 1700 AD, which leads dating laboratories to simply quote "modern" in this time period.

We have noted on in section 4.5 that the stair-step shape of the calibration curve in the last 450 years does tend to cluster the radiocarbon ages of a series of events that have occurred uniformly in time. However this apparent clustering effect is largely offset by the corresponding desegregation of the "clustered" radiocarbon dates when these are converted back to the full possible calendric age ranges. For that reason we have presented the full possible calendric age ranges for samples in Figure 4.8 (page 111), not the radiocarbon dates, and have noted the similarity of the full calendric age ranges, as opposed to the number of times a particular calendric date is
represented. If date modes are adopted then McFadgen et al. 1994 have demonstrated that the periods AD 1680 – 1720; and to a lesser extent AD1525 – 1550 and 1455 - 1460 are still relatively over-represented as a relict of the calibration curve.

It is worth noting again that the precise timing of the paleo-earthquakes in this thesis is based independently on forest disturbance and tree ring patterns within each of the relatively wide radiocarbon defined date ranges from the trenching work. Because of this there is no artificial effect of the calibration curve influencing the definition of event timing.

![Diagram of decadal calibration curve](image)

**Figure A1** - The decadal calibration curve of Stuiver and Becker (1993) for the period since 1400 AD.
Southern Hemisphere offset

The southern hemisphere calibration curve appears to be offset back in time from curves derived in the northern hemisphere. Vogel et al., 1993 estimate this offset is 40 years and is constant throughout the time range. It appears more likely that the offset is somewhat less than this (around 30 years?) and that it varies through time. Until the results of work by the University of Waikato radiocarbon dating laboratory, in conjunction with various northern hemisphere institutions, which is currently being undertaken is published we have adopted the Vogel et al. offset. It makes little general difference to the calendric date ranges for the Toaroha River and Crane Creek events quoted here because any reduction from 40 years has the effect of moving both forward slightly, while keeping approximately the same time interval between them. Because in this thesis the best earthquake date estimates have come from the forest disturbance record and tree rings, any subsequent change in the southern hemisphere offset does not change the estimates of earthquake event timing.
APPENDIX TWO

PROBABILITY ASSESSMENT METHOD OF NISHENKO & BULAND (1987)

Nishenko & Buland (1987) present a method for fitting a lognormal distribution to a consolidated set of data for earthquake recurrence periods gathered from rupture segments around the world.

The lognormal distribution is a function of an independent variable and two other parameters \( \mu \) and \( \sigma \) which are the mean and standard deviation of the distribution and control its shape. In their use of the lognormal distribution Nishenko & Buland set the independent variable to be \((T/T_k)\), \(T\) being the data point return period and \(T_k\) being derived from \(\ln(T_k) = \text{average} \left[ \ln(T) \right] \) for each earthquake segment \(k\).

Into the distribution are also worked the errors associated with the derivation of the return periods, and also the natural variability of the recurrence periods. For each segment \(k\), \(T_k\) is initially derived as described above. For events derived from carbon dating and similar methods with dating errors, errors are estimated and assigned as \(\sigma_{ik}\) (i denoting the individual data point within fault segment \(k\)).

An initial estimate of the natural overall variability of the return periods is also made, being \(\sigma_D\), where \(\sigma_D^2 = \text{var} \left[ \ln(T_{ik}) \right] \). These two sources of error variability are then combined for each data point to give \(\sigma_{ik}^2\) (for more details of this derivation readers are referred to Nishenko & Buland (1987) equations 10 and 11).
At this point in the analysis a least squares fitting method is used. For each data point a parameter \( \tau_i \) is calculated (\( \tau_i = T_i/\bar{T}_k \) - i.e. the return period normalised on the rupture segment \( T \)); all \( \tau_i \) are collated in ascending order from \( j = 1 \) to \( N \), and each event is assigned a probability level \( F_{j,N} = (j - 1/2)/N \). The inverse lognormal function of each \( F_{j,N} \) is then calculated, using initial estimates of \( \mu_D \) and \( \sigma_D \). The logarithm of the result is then subtracted from \( \ln(\tau_i) \) and this is divided by a parameter \( \sigma_i^2 \) (this parameter is in turn derived from \( T_i, \bar{T}_k, \sigma_{ik} \), and \( \sigma_d \) as described in Nishenko & Buland (1987) equations 10 and 11). The results of these calculations are summed for all the data points and then this sum is minimised by iteratively varying \( \sigma_d \) and \( \mu_d \).

Once the new values for \( \mu_d \) and \( \sigma_d \) have been thus determined they are used to recalculate values for \( \bar{T}_k \) (using Nishenko & Buland, 1987, equation 9), \( \sigma_k \) (using their equation 10) and \( \sigma_{ik} \) (their equation 12).

This leads into recalculation of the \( \tau_i \) and consequent manipulations as described above, resulting in further iteration of the minimisation process to yield improved estimates for \( \sigma_d \) and \( \mu_d \). This whole cycle is repeated until stable values of \( \sigma_d \) and \( \mu_d \) are arrived at, which also allows the calculation of final values for \( \bar{T}_k \) for each fault segment.

These parameters (\( T_i, \mu_d \) and \( \sigma_d \)) can be then be used to calculate expected recurrence intervals (Nishenko & Buland, 1987, equation 14) and conditional probabilities of events within given time periods (their equation 15).
APPENDIX THREE

THE MOUNT ADAMS ROCK AVALANCHE AND POERUA RIVER
LANDSLIDE DAM OF 6 OCTOBER 1999

To date nothing has been formally published on this recent failure but we are grateful to Dr Tim Davies (Lincoln University) who has provided information for this brief summary of the key features.

Mount Adams rock avalanche

The Mt Adams rock avalanche occurred on the northeast face of the northern peak (2130 m) on the Adams Range (GR 135/095680). Between 10-15 million m$^3$ of schist and colluvial debris fell about 1790 m into the Poerua River valley, forming a landslide dam about 100 m high and impounding a 1200 m long lake 11 km upstream from the SH 6 road bridge (GR 134/104710).

The main part of the avalanche followed a steep side stream to the Poerua valley floor, while a northern lobe of mainly finer debris overtopped the low spur to the north, and formed a small temporary dam that breached shortly after the main dam was overtopped on 7 October 1999. Besides damming the Poerua River, landslide debris also infilled the channel of a side stream on the true left, creating a small debris-dammed lake. The failure path was about 3 km long, with an average slope of 37 °, but the upper half of the failure surface was much steeper (45-57 °).
As expected for this type of landslide, the avalanche was accompanied by a powerful wind and sandblast, which produced a wide zone of flattened and debarked trees that extended up to 200 m above the former valley floor. The main scarp at the top of the source area is about 900 m wide.

Slope failure probably commenced in colluvium below bush line, and extended in schist bedrock to the top of the 2130 m peak at the northern end of the Mt Adams massif.

Residents in the upper Poerua Valley (the McKenzie family) report that the rock avalanche was not a single sudden event, but took place over several hours on the morning of 6 October 1999, beginning at around 3 am. The initial loud rumbling noises from the avalanche are reported to have lasted for about 2 hours, but continuing intermittent rock falls could be heard throughout most of the day. The landslide was not obviously triggered by rainfall or an earthquake, but was most probably due to ongoing progressive weathering, erosion, and weakening of slope materials. A much smaller landslide which occurred on the same face in August 1997, about 18 months previously, is thought to have undercut the adjacent upper slope, making it more susceptible to large-scale collapse.

Two small earthquakes (M4.0 and M4.1) occurred in the area 10 days prior to the landslide (at 0301 and 0500 hrs on 26 September 1999). These events were located approximately 30 km ENE of the landslide, at a depth of less than 33 km (i.e. crustal). Since the rock avalanche did not occur during, or immediately after, the earthquakes, they cannot be regarded as the slope failure trigger. However, the local area was quite strongly shaken (at least MM5), and they may have influenced the timing of the failure by causing increased slope creep and/or joint dilation.
Poerua landslide dam and lake

Debris from the rock avalanche filled the valley to a maximum depth of about 120 m, and infilling the side stream on the true left with 60-70 m of debris. Key information about the landslide dam and lake is summarised below.

<table>
<thead>
<tr>
<th>1. Location and date of dam formation</th>
<th>Poerua River, 6 October 1999</th>
</tr>
</thead>
<tbody>
<tr>
<td>2. Type and characteristics of landslide forming dam</td>
<td>Rock and debris avalanche</td>
</tr>
<tr>
<td>(a) Vertical fall of debris</td>
<td>1790 m</td>
</tr>
<tr>
<td>(b) Length of debris runout</td>
<td>3 km</td>
</tr>
<tr>
<td>(c) Debris and wind blast travel above valley bottom</td>
<td>200 m</td>
</tr>
<tr>
<td>3. Bedrock and surficial geology in landslide source area</td>
<td>Schist and colluvium</td>
</tr>
<tr>
<td>4. Trigger or initial cause of landslide</td>
<td>No known trigger event</td>
</tr>
<tr>
<td>5. Underlying causes of landslide</td>
<td>Very steep glaciated slope; weak rock mass; previous failures; small (M4) earthquakes 10 days before failure.</td>
</tr>
</tbody>
</table>
6. Dam size and volume
(a) Height of landslide dam 120m (max)
(b) Width of landslide dam 450m at spillway
(c) Length of dam (along valley) 650m
(d) Volume of landslide dam 10 million m³
(e) Slope of dam faces downstm: 24° upstm: 12°

7. Physical characteristics of materials making up the dam.
Boulders of schist up to 4-6m, across, set in a well-compacted, silt & sand-sized matrix (colluvial and pulverised schist). Spillway overflow channel armoured with large schist boulders.

8. Lake size, depth, and volume
(a) Length 1200 m
(b) Width 350 m
(c) Maximum depth 80 m
(d) Volume of lake 5-7 million m³

9. Presence of seepage, piping, or failures of the dam face
None present prior to breaching

10. Longevity of landslide dam and lake (after formation)
6 days (5 after overtopping)

11. Type of breach
Partial, with 40-50 m deep spillway channel eroded through dam in first flood.

12. Estimated maximum dam-break discharge
Approx. 500-600 m³/sec