DEVELOPMENT OF MULTI-CHANNEL ANALYSIS OF SURFACE WAVES (MASW) FOR CHARACTERISING THE INTERNAL STRUCTURE OF ACTIVE FAULT-ZONES AS A PREDICTIVE METHOD OF IDENTIFYING THE DISTRIBUTION OF GROUND DEFORMATION.

2008

RESEARCH UNDERTAKEN BY:

BRENDAN DUFFY

with

MRS JOCELYN CAMPBELL and

DR MICHAEL FINNEMORE

November 2008

1 Department of Geological Sciences, University of Canterbury
2 Southern Geophysical Limited

RESEARCH FUNDED BY:

Earthquake Commission (Research report number 07/US37)

Brian Mason Trust Fund, University of Canterbury
EXECUTIVE SUMMARY

Much of New Zealand lies within the seismically active broader plate boundary deformation zone. Built structures are often, therefore, located in areas where there are undetected or poorly defined faults. Where these structures are sited near to, or across, such faults or fault zones, they may sustain both shaking and ground deformation damage during an earthquake. Within this zone, management of seismic hazards needs to be based on accurate identification of hidden faults including the potential fault damage zone associated with the likely width of off-plane deformation bordering the principal rupture surface. The Multi-channel Analysis of Surface Waves (MASW) shallow seismic survey method is used here to image buried fault zones in a range of lithologies. The technique is calibrated to optimise the technique for fault location and the limits of resolution in soft rocks are explored. Particular objectives were to define within-zone variability in shear wave velocity that could be readily correlated with measurable rock properties and with the distribution of ground deformation.

An MASW survey records dispersive Rayleigh waves. Rapid data acquisition results from the use of a towable land streamer carrying a multichannel geophone string, providing both data redundancy and reduction in labour. The fundamental mode wave velocities are plotted against their frequencies to produce a fundamental mode dispersion curve. The curves for each record are then inverted to determine s-wave velocity with depth, and the resulting 1D inversions are interpolated to construct a 2D shear wave velocity profile. Horizontal variation in s-wave velocity in these pseudo-sections defines fault zones and acts as a proxy for changes in rock mass properties. The property variations may be due to intensity of fracturing and/or fundamental strength differences related to juxtaposition of different lithologies.

MASW-derived shear wave velocity profiles are presented for each test and calibration site, and interpreted in relation to adjacent topographic profiles, to the local geomorphology and to geotechnical characterisation of exposed bedrock in adjacent river channels. The initial Dalethorpe study (Springfield Fault, 80 km west of Christchurch) explores the optimisation of data acquisition and the influence of fracture frequency on s-wave velocity in hard rock by correlation of MASW with laboratory and outcrop characterization. Predictable limitations include modification of the MASW profile velocities with depth by lithostatic loading conditions relative to outcrops and laboratory
samples. Intense fracturing also plays a role by severely limiting the sampling of intact material for laboratory characterization. This is compounded by unpredictable scaling relationships between laboratory and survey velocities. That said, the Dalethorpe profiles correlate well with geological and geotechnical mapping in parallel bedrock and fault zone riverbank exposures. In the greywacke lithologies found at Dalethorpe, shear zones were shown to be characterized by low s-wave velocities compared with adjacent competent bedrock. The velocity drop is consistent with fracture development and weathering observed on site, with fault zone velocities measured in horizontal crossholes, and with detailed surveys of stream bed and strath morphology (Figure E1). It is also consistent with laboratory observations of velocity decline in artificially fractured greywacke.

![Figure E1: A northeast facing across site correlation of the MASW profiles with detailed geomorphological surveys and structural interpretations. Significant detail is evident here of shear wave velocity variability within the wider fault zone that can be closely correlated with individual shears and secondary faulting or off-plane deformation. Features to note include the warped surface of Terrace B(13) along the fault scarp profiles (7, 10-12), lesser warping of the younger Terrace A strath and subtle gradient changes in the modern river bed coinciding with the high velocity s-wave zone of differentially uplifted bedrock. These features are indicative of persistent deformation for 300-400metres east of the obvious scarp. The shear zone extends well into the footwall. Note: Fault dips exaggerated by 10x VE. Also, a regional gradient to the NE affects precise superimposition of corresponding surfaces.](image1)

In order to explore the boundaries of resolution of the MASW technique, soft rock sites were selected at Boby's Stream (a tributary of the Waipara River, 50 km NW of Christchurch) to be representative of the problems likely to be encountered in distinguishing complex faulting in mixed lithologies. Shear wave propagation in the
Boby's Stream area showed generally lower velocities and consequently smaller contrasts induced by cataclasis. Laboratory soft rock characterization showed that a geotechnical threshold exists at an unconfined compressive strength value of ~ 20 MPa, below which rock strength rather than fracturing dominates shear wave velocities. This result was compared with MASW survey results and found to be in good agreement. Faulting in softer rocks was generally reasonably well imaged where differing lithologies were juxtaposed (Figure E2). This was despite overlapping velocity ranges between lithologies (Table E1). Although primary lithological variation provided the best contrasts, in some cases it could be seen to be slightly overprinted by cataclastic velocity declines (Waikari/Amuri fault contact, Figure E2). Interpretation was generally aided by outcrop control and could be expected to be more problematic if contrasts are weak and control is sparse. Faults were located and imaged in all but the weakest lithologies at Dalethorpe and Boby's Stream.

![Boby's Stream MASW survey RD 2](image)

**Figure E2**: Interpreted MASW profile from line RD2, run across a mapped fault at Randolph Downs, Boby's Stream. A strong contrast highlights the faulted contact between higher velocity Amuri Limestone and lower velocity Waipara Greensand. As seen in Table E1 below, the velocity fields of Amuri and Waikari lithologies, however, are almost identical. Fortunately, both Waikari and Amuri lithologies are above the 20 MPa strength threshold and the fault can therefore be picked out as a cataclastic velocity decline.

**Table E1**: (from Table 8.1) Summary table of predicted shear wave velocities for all lithologies encountered in this study. Greywacke velocities after Perrin (2008, pers. comm)

<table>
<thead>
<tr>
<th>Unit</th>
<th>$V_{s_{\text{min}}}$(m/s)</th>
<th>$V_{s_{\text{max}}}$(m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Torlesse greywacke</td>
<td>500</td>
<td>1100</td>
</tr>
<tr>
<td>Weka Pass Limestone</td>
<td>302</td>
<td>593</td>
</tr>
<tr>
<td>Mt Brown sandstone</td>
<td>227</td>
<td>596</td>
</tr>
<tr>
<td>Amuri Limestone</td>
<td>202</td>
<td>628</td>
</tr>
<tr>
<td>Waikari Formation</td>
<td>190</td>
<td>563</td>
</tr>
<tr>
<td>Waipara Greensand</td>
<td>130</td>
<td>594</td>
</tr>
<tr>
<td>Lloburn Formation</td>
<td>98</td>
<td>2-300</td>
</tr>
<tr>
<td>Conway Formation</td>
<td>47</td>
<td>2-300</td>
</tr>
</tbody>
</table>
Recent major events such as the 1999 Chi-Chi, Taiwan earthquake have shown the importance of understanding and predicting the extent of off-plane deformation. Variability in depth to bedrock, as inferred from shear wave velocity, was used at Dalethorpe to document, not only aseismic folding, but less predictable details of the internal structure of the wider deformation zone (Figure E1). These details correlated strongly with profiles of stream-bed morphology. Similarly at Boby's Stream, ongoing deformation of abandoned terraces and of the bed of the Waipara River correlated strongly with shear wave velocity imaging of small actively growing anticlines in soft Tertiary rocks.

The technique was tested by blind MASW processing of High Resolution Reflection (HRR) seismic data acquired across the concealed thrust fault bounding the eastern side of the Taieri Ridge in Macraes Flat, Central Otago. Loess-blanketed outwash slopes of Quaternary sediments, together with the lack of exhumed structures and anthropogenic land use, make the fault difficult to pinpoint even to within 10's of metres. The Taieri Ridge test appears to highlight the structural sensitivity of the technique by constraining the location of the fault, which remained ambiguous after standard processing of the reflection data. The results are promising and appear to image high velocity Otago Schist juxtaposed against the low velocity Tertiary and Quaternary units (Figure E3 A). Trenching is planned, which should confirm the MASW results.

Processing of the Taieri Ridge dataset has further led to the proposal of a novel surface wave imaging technique termed Swept Frequency Imaging (SFI). The tentatively interpreted SFI image of the Taieri Ridge seismic line (Figure E3 B) appears to show the detailed structure of the fault zone, which predictably includes considerable deformation extending forward of the fault into the softer units overlying the unconformity. The unconformity itself is also interpreted and bending moment thrusts are inferred offsetting the surface of the unconformity. Minor shears imaged using SFI appear to correlate with structures observed in an initial trench, which was sited based on interpretation of the HRR and other geophysical data, but was found to lie in the footwall only. The SFI image of the Taieri Ridge fault will be initially validated or otherwise by excavation of a second trench. The vertical scale of the image is indeterminate but should be possible to determine if appropriate simple software is developed. SFI is an exciting opportunity that awaits further development.
The outcome of this study clearly demonstrates the ability of MASW to identify the location and extent of hidden shear zones and to provide a first order indication of underlying rock mass characteristics. This is because the MASW technique is sensitive to lateral variations in shear wave velocity due to subsurface structure. On this basis, MASW is shown to provide a useful complement to geomorphologic mapping and trenching for rapid delineation of fault zones in most lithologies.

Figure E3: A) Elevation corrected conventional MASW survey results for Toieri Ridge test site. The fault is inferred to be located somewhere between stations 580 and 600. B) Swept frequency image of the northwest end of the Toieri Ridge seismic line with interpretation, suggesting a fault outcropping at around station 590, in concurrence with A. The black seismic lines are surface wave events of the frequency scaled on the left of the image. The slope of the line is inversely proportional to the velocity of the event. Red lines are interpreted as faults or shears defined by discontinuities in the stacked lines, whilst the blue line is a velocity-based pick of the location of the basement unconformity. The lateral location of the first trench is shown, together with the location of major unit boundaries observed in the trench. For discussion see text.

The document that follows contains the full text of the University of Canterbury MSc thesis of the same name, amended in the light of examiner's comments.
ACKNOWLEDGMENTS

This thesis was supported by a non-biennial grant from the Earthquake Commission and by the Mason Trust. Dr Hugh Cowan and Priscilla Cheung at EQC were ever willing to fund my penchant for conferences and the Tauranga Geological Society of New Zealand conference pushed this project forward immensely.

I wish to sincerely thank my thesis supervisors Mrs Jocelyn Campbell and Dr Michael Finnemore for their assistance in the collection and processing of data and their reviews of this manuscript. Jocelyn’s in particular were thorough, insightful and patient. Congratulations on your retirement Jocelyn. Thanks also to Dr Andrew Gorman from Otago University, who was the external thesis examiner. His comments greatly improved this thesis. Dr Phil Tonkin not only provided ever-useful advice and assistance with OSL sampling and loess stratigraphy, but also provided timely sanity checks. Erakine visitors Assoc. Prof. Paul Santi and Dr Jim McKean joined me for valuable late stage geotechnical discussions. Dr. Andy Nicol at GNS enthusiastically introduced me to Professor Richard Norris from Otago, whose MSc student became an important collaborator. I hope that Jeremy Kilner derived as much value from the collaboration as did I.

A great deal of thanks is due to the landowners on whose land I have worked. So thankyou to Gary Ferguson, Ian and Elizabeth Turnbull, Preston Allen, Gareth Renowden and Raymond Herber. Of these, three deserve special mention. Gareth, Ian and Elizabeth went beyond the call of duty. Gareth lent me his ride on tractor mower to mow the RD2 site and took a series of photographs for me that helped me greatly. Ian and Elizabeth were also ever welcoming and among other favours lent me valuable family photographs of the site.

Department staff and students helped greatly. Anekant’s blunt and welcome intervention moved me in the direction that became this thesis. The other technical staff provided assistance along the way, particularly Vanessa Tappenden and Cathy Knight. John Southward was ever patient with my computer frustrations and Nick Etheridge in the Department of Biological Sciences provided timely workshop assistance. Many students assisted me in the field and their help was appreciated. Field assistants included (in chronological order) Keith Machin (hardly a student), Paul Shackleton (him neither!), Yvette Hobbs, Richard and Kirsty Cooksey, Matt Dodson, Vicky Kershaw, Lucille Tatard
and David Hood. My numerous discussions with Florian Buch were invaluable from my
viewpoint and contributed greatly to my understanding of and confidence in my data.
Thank you all, good luck with your own studies and apologies and thanks to any I missed.

David Park and his colleagues at the Geospatial Research Centre helped during my
abortive experimentation with photogrammetric techniques and made it an intriguing and
enjoyable failure. Their expertise is an extremely valuable resource and they will
hopefully be key partners in some of the research directions identified in the concluding
chapter of this document.

Above all, thanks are due to my family. My parents gave me the education that started it
all and supported my decision to return to study. My mother-in-law Gaye-Lynn and her
partner Dave Potter were outstanding with their moral support and timely pep talks. My
father-in-law Graham Boniface kindly arranged the printing of this volume by Xerox
Limited. But most of all, my biggest debt of gratitude goes to my wife, Nicci and my
children, Ciara and Isabella, to whom this volume is dedicated. As well as teaching me to
touch-type, which greatly aided preparation of my manuscript, Nicci inspired me from the
very start of my geology career. Without her support none of this would have been
possible.
# TABLE OF CONTENTS

EXECUTIVE SUMMARY ......................................................................................................................... I

ACKNOWLEDGMENTS ............................................................................................................................ VI

LIST OF EQUATIONS, FIGURES AND TABLES .................................................................................. XIII

ABSTRACT ............................................................................................................................................. XXXI

1 INTRODUCTION ................................................................................................................................. 1

1.1 Motivation: Present day land use planning in New Zealand fault-zones ........................................... 1

1.2 Objectives....................................................................................................................................... 5

1.3 New Zealand's Tectonic Setting ...................................................................................................... 6

1.4 Tectonic setting of Central South Island and study sites ................................................................ 6

1.5 Study Format .................................................................................................................................. 9

2 SHEAR WAVE VELOCITIES: INTERPRETATION AND TECHNIQUES ................................................ 11

2.1 S-wave velocities and fault-zone planning ...................................................................................... 11

2.2 Factors affecting rock mass S-wave velocity .................................................................................. 12

2.3 Potential correlative rock mass classifications ............................................................................... 14

2.3.1 Photogrammetric options ......................................................................................................... 17

2.4 Multi-scale S-wave velocity measurement ...................................................................................... 18

2.4.1 Laboratory scale: Ultrasonic velocity measurement .................................................................. 18

2.4.2 Outcrop scale: crosshole velocity measurements .................................................................... 19

2.4.3 Survey scale: Multichannel analysis of surface waves .............................................................. 20

3 STUDY METHODS ............................................................................................................................. 25

3.1 Mapping ......................................................................................................................................... 25

3.2 Survey scale seismic investigations – MASW ................................................................................. 25

3.2.1 Surface-wave Data Acquisition .................................................................................................. 25

3.2.1.1 Acquisition parameters ......................................................................................................... 25

3.2.2 Equipment and settings .............................................................................................................. 27

3.2.3 Data Pre-processing .................................................................................................................... 27

3.2.4 Dispersion curve extraction ........................................................................................................ 29

3.2.5 Surface wave Inversion and contouring ...................................................................................... 30
3.3 Outcrop scale geotechnical and seismic investigations .......................................................... 31
  3.3.1 Rock mass characterization ............................................................................................... 31
  3.3.2 Crosshole velocity measurements .................................................................................. 31
3.4 Laboratory scale geotechnical and seismic investigations ....................................................... 32
  3.4.1 Laboratory investigations ............................................................................................... 33
3.5 Geomorphological investigations ........................................................................................... 35
3.6 Correlation methods ............................................................................................................... 37

4 Calibration of MASW to the surface expression and engineering properties of faulted Torlesse greywacke at Dalethorpe ....................................................................................... 38
  4.1 Introduction ........................................................................................................................ 38
    4.1.1 Scope and objectives ........................................................................................................ 38
    4.1.2 Previous work .................................................................................................................. 38
    4.1.3 Geomorphology and geology of study area ................................................................... 40
  4.2 MASW profiles .................................................................................................................... 41
    4.2.1 Data collection ............................................................................................................... 41
    4.2.2 Dispersion curve extraction .......................................................................................... 44
    4.2.3 2D S-wave velocity sections ........................................................................................ 47
  4.3 MASW correlations ............................................................................................................. 51
    4.3.1 Geomorphology ............................................................................................................ 51
      4.3.1.1 General correlations ............................................................................................... 51
      4.3.1.2 Detailed correlations ............................................................................................. 54
    4.3.2 Line to line correlations .................................................................................................. 56
    4.3.3 Rock mass character ...................................................................................................... 57
      4.3.3.1 Crosshole velocity testing ...................................................................................... 57
      4.3.3.2 Scanline surveys .................................................................................................... 60
  4.4 Laboratory testing ................................................................................................................. 65
  4.5 Discussion ............................................................................................................................. 70
  4.6 Conclusions ........................................................................................................................... 71

5 Structure, kinematics and paleoseismicity of the Springfield fault, Dalethorpe ......................... 72
5.1 INTRODUCTION ........................................................................................................... 72

5.1.1 Scope and objectives .............................................................................................. 72

5.2 FAULT-ZONE STRUCTURE ..................................................................................... 73

5.3 FAULT-ZONE HISTORY ........................................................................................... 78

5.3.1 Geomorphological principles and observations ...................................................... 78

5.3.2 Coseismic slip estimation ...................................................................................... 82

5.3.3 Timing of paleoearthquakes .................................................................................. 90

5.3.4 Uplift and slip rates .............................................................................................. 93

5.3.5 Recurrence interval ............................................................................................. 94

5.3.6 Paleo-magnitude ................................................................................................ 95

5.4 KINEMATIC OBSERVATIONS AND DISCUSSION .............................................. 96

5.5 CONCLUSION .......................................................................................................... 98

6 MASW CALIBRATION AT THE BOBY’S STREAM FAULT, WAIPARA ......................... 100

6.1 INTRODUCTION ....................................................................................................... 100

6.1.1 Geological setting and previous work .................................................................... 100

6.1.2 Scope and objectives ........................................................................................... 104

6.2 ROCK MASS AND MATERIAL CHARACTERISATION ........................................... 105

6.2.1 Rock property determinations ............................................................................. 105

6.2.2 Ramset gun calibration ....................................................................................... 110

6.3 MASW PROFILES .................................................................................................... 114

6.3.1 Site selection, geology and geomorphology .......................................................... 114

6.3.1.1 The Deans ....................................................................................................... 114

6.3.1.2 Limestone Hills ............................................................................................. 116

6.3.1.3 The Quarry .................................................................................................... 117

6.3.1.4 Randolph Downs .......................................................................................... 117

6.3.2 Data collection .................................................................................................... 120

6.3.3 Processing ............................................................................................................ 121

6.3.4 2D S-wave velocity sections .............................................................................. 125

6.4 GEOMORPHOLOGICAL CORRELATIONS .......................................................... 130
6.5 DISCUSSION ..................................................................................................................... 135
6.6 CONCLUSIONS ............................................................................................................. 137

7 APPLICATION OF MASW AT THE TAIERI RIDGE ......................................................... 138

7.1 INTRODUCTION ........................................................................................................... 138

7.1.1 Scope and objectives ............................................................................................. 138

7.1.2 Site geology and geomorphology ....................................................................... 140

7.1.2.1 High Resolution Reflection Seismic ................................................................. 141

7.1.2.2 Trenching ........................................................................................................ 143

7.2 MASW SURVEY .......................................................................................................... 145

7.2.1 Data collection and processing ............................................................................ 145

7.2.2 Dispersion curve extraction ................................................................................... 145

7.2.3 2D S-wave velocity sections, Interpretations and correlations ......................... 150

7.3 CORRELATIONS ......................................................................................................... 152

7.3.1 Geomorphology .................................................................................................. 152

7.3.2 High Resolution Seismic Reflection Survey ....................................................... 153

7.4 SWEEP FREQUENCY IMAGING ............................................................................... 154

7.5 DISCUSSION .............................................................................................................. 158

8 DISCUSSION, CONCLUSIONS AND RECOMMENDATIONS FOR FURTHER RESEARCH ................................................. 162

8.1 UTILITY OF THE MASW TECHNIQUE ...................................................................... 162

8.1.1 Fault location ....................................................................................................... 162

8.1.2 Geotechnical S-wave variability correlations ....................................................... 163

8.1.2.1 Discrimination of cataclasis ........................................................................... 163

8.1.2.2 Lithological discrimination .............................................................................. 163

8.1.2.3 Transferring between different scales of investigation ................................... 164

8.1.3 S-wave velocity, geomorphology and ground deformation ................................ 165

8.1.4 Topography ......................................................................................................... 166

8.2 FUNDAMENTAL STRENGTHS AND WEAKNESSES OF THE MASW TECHNIQUE .................................................. 166

8.2.1 Strengths ............................................................................................................. 166

8.2.2 Relative strengths: A comparison with other geophysical methods .................... 167
LIST OF EQUATIONS, FIGURES AND TABLES

Figure E1: A northeast facing across site correlation of the MASW profiles with detailed geomorphological surveys and structural interpretations. Significant detail is evident here of shear wave velocity variability within the wider fault zone that can be closely correlated with individual shears and secondary faulting or off-plane deformation. Features to note include the upwarped surface of Terrace B(13) along the fault scarp profiles (7, 10-12), lesser warping of the younger Terrace A strath and subtle gradient changes in the modern river bed coinciding with the high velocity s-wave zone of differentially uplifted bedrock. These features are indicative of persistent deformation for 300-400 metres east of the obvious scarp. The shear zone extends well into the footwall. Note: Fault dips exaggerated by 10x VE. Also, a regional gradient to the NE affects precise superimposition of corresponding surfaces.................................................................ii

Figure E2: Interpreted MASW profile from line RD2, run across a mapped fault at Randolph Downs, Boby's Stream. A strong contrast highlights the faulted contact between higher velocity Amuri Limestone and lower velocity Waipara Greensand. As seen in Table E1 below, the velocity fields of Amuri and Waikari lithologies, however, are almost identical. Fortunately, both Waikari and Amuri lithologies are above the 20 MPa strength threshold and the fault can therefore be picked out as a cataclastic velocity decline.................................................................iii

Table E1: (from Table 8.1) Summary table of predicted shear wave velocities for all lithologies encountered in this study. Greywacke velocities after Perrin (2008, pers. comm) .................................................................iii

Figure E3: A) Elevation corrected conventional MASW survey results for Taieri Ridge test site. The fault is inferred to be located somewhere between stations 580 and 600. B) Swept frequency image of the northwest end of the Taieri Ridge seismic line with interpretation, suggesting a fault outcropping at around station 590, in concurrence with A. The black seismic lines are surface wave events of the frequency scaled on the left of the image. The slope of the line is inversely proportional to the velocity of the event. Red lines are interpreted as faults or shears defined by discontinuities in the stacked lines, whilst the blue line is a velocity-based pick of the location of the basement unconformity. The lateral
location of the first trench is shown, together with the location of major unit boundaries observed in the trench. For discussion see text. ..................v

Figure 1-1: Tectonic setting and main structural features of the New Zealand microcontinent bisected by an obliquely convergent plate boundary between the Australian and Pacific plates. Study sites are shown, numbered 1-3. Arrows show rates of relative convergence in mm/yr. Plate boundary motion and partitioning calculated at Franz Josef by Norris and Cooper (2001) according to NUVEL-1A global model (DeMets et al., 1994). Map modified after Pettinga et al. (2001) and Campbell et al. (2003). ..................................................................................7

Figure 1-2: Tectonic setting and main structural features of North Canterbury. A) Location relative to the subduction/transpression transition zone between the Hikurangi Margin and the Marlborough Fault System (MFS, stippled). Other zones (after Pettinga et al., 2001) are the West Culverden Fault-zone (WCFZ), the Porters Pass Amberley Fault-zone (PPAFZ), and the North Canterbury Fold and Fault Belt (NCFFB). B) Locations of the Dalethorpe (1) and Boby’s Stream (2) study areas at either end of the PPAFB. (modified after Litchfield et al., 2003). ........................................................................................................8

Figure 1-3. Central Otago map with inset South Island location map showing the actively rising northeast trending ranges of Central Otago. The Taieri Ridge is located on the southeastern margin of this system and the study site is marked with a white square. Cross section after Norris and Nicols (2004).........................9

Figure 2-1: Photographs from the San Francisco earthquake of 18 April, 1906 showing the damage associated with low S-wave velocity materials in the subsurface (Soule, 1907). .........................................................................................11

Figure 2-2: Reduction in S-wave velocity attributable to moderate weathering, graphed as a function of the fresh rock S-wave velocity (after Fumal, 1978). ......13

Figure 2-3: Variations of S-wave velocity with fractured rock mass parameters for the model of Boadu (1997). Both parameters exhibit a strong correlation with S-wave velocity. .................................................................................14

Figure 2-4: Geological strength index modified after Habimana et al. (Habimana et al., 2002) .................................................................16
Figure 2-5: Nakagawa et al.'s (2002) experiment was conducted using ultrasonic transducers with waves propagating parallel to fractures in simulated fractured material (steel plates). The lead foil provided acoustic coupling.……..20

Figure 2-6: MASW 2D Vs profile across an unnamed New Jersey fault-zone. The fault-zone and the dipping and weathered bedrock of the fault bounded formations are clearly imaged to 20m depth. High velocities can be seen south of the fault in the interbedded sandstones and shales of the Stockton Formation. This is in contrast with the mid-range velocities of the weathered massive and laminated mudstones of the Lockatong Formation, which outcrops to the north of the fault (Ivanov et al., 2006).…………………………………………………………………………………………………23

Figure 3-1: Basic steps in the MASW method……………………………………..26

Figure 3-2. MASW survey geometry using a land streamer. Key acquisition parameters are illustrated and selection of these is discussed in the text. ……………...26

Figure 3-3. Land streamer carrying 24 channel 4Hz geophone string. The streamer is towed by the vehicle in the photograph, which is switched off between moves. ……………………………………………………………………………………27

Figure 3-4. MASW survey configuration for walkaway construction…………..28

Figure 3-5. Relationship between ease of differentiation of fundamental and higher mode rayleigh waves and the length of the geophone string. The longer the array, the easier to visually identify higher mode and body waves. A short 24 channel array (1) shows very poorly the diverging velocities of fundamental and higher modes compared with a 120 channel array (3) at a similar near offset. A 48 channel array (2) collects significantly more fundamental mode energy, facilitating differentiation and muting……………………………………………………………..29

Figure 3-6: MASW profile map with velocity contrast (A) with no topographic correction and (B) with soundings sheared in Surfer8. Profile B compares favourably with the earth model (C) whilst profile A produces structurally misinterpretable artefacts. ……………………………………………………………………………………30

Figure 3-7: S-wave seismic source manufactured to fit 32 mm drill holes. The rod and steel weight of the slide are insulated from the casing along the entire length and contact is only possible at the two ends. By attaching a cable to both casing and
slide, the impact of the weight is used to trigger data collection. Shear coupling is assumed to be concentrated at the knock on cap. The handle is insulated to prevent static transfer of a 0-5V ‘signal’ triggering data collection. ...32

Figure 3-8: Horizontal crosshole velocity measurement equipment setup. ...32

Figure 3-9: Example of fractured greywacke core. Fractures were introduced from right to left. The core is numbered 1/9/1 using the system adopted for this study (sample 1/core 9/orientation 1). ...34

Equation 3-1 ...34

Figure 3-10: Ramset gun being fired at a Waikari Siltstone outcrop, Waipara River. 36

Figure 4-1 Location map for Dalethorpe study site with inset airphoto detail of the Upper Hawkins Basin. A, B and C refers to the flight of fluvial terraces. The mapped fault’s surface trace is shown. ...39

Figure 4-2: A) Integration and interpretation of GPR and resistivity tomography at Dalethorpe (Corboz, 2004). Corboz delineates boundaries between three subhorizontal units. High resistivity values to the east (Z2) correspond to dry sand and/or gravel seen above outcrop D1, whereas lower resistivities to the west (Z1) indicate higher water content (near the swamp) and/or silt/clay content. Bedrock is invoked to explain both the limited penetration of GPR and lower resistivity values at > ~5 m depth. B) 3D GPR ‘cube’ showing the time to a major GPR reflector. Shorter times are darker indicating uplift to the east. ...40

Figure 4-3: The Dalethorpe surveys covered an area of 726 m by 572 m relative to the strike of the fault. The western end of Line 3 (labelled twice) overran and replaced Line 2. The outside of the meander in the river bend at the SW end of line 6 exposes a shear zone (outcrop D1). No obvious scarp is developed in the terrace crossed by the lines. The orange square indicates the area surveyed by Corboz (2004) using GPR. ...42

Table 4-1. Acquisition parameters for Dalethorpe MASW surveys. ...44

Figure 4-4. Examples of records from each survey line. Surface wave energy is dominant in all records but records are adversely affected by wavefield scattering. This is particularly noticeable for line 4. ...44
Figure 4-5. Contiguous dispersion curves taken from the centre of line 3. A noticeable shift in dispersion occurs at record numbers greater than 170. Some of the lower frequency Rayleigh wave phase velocities are excessive suggesting body wave contamination of the records. Dalethorpe data were the first to be processed during the course of this study and techniques mentioned in Chapter 3 were subsequently adopted to deal with this problem. .................................45

Figure 4-6. A) Comparison of fixed source and fixed receiver walkaway records centred over the same point in the subsurface. Using FRWs is shown to result in repetition of systematic noise, particularly obvious for diffractions. B) Direct comparison of dispersion curves generated using a 24 channel records with both types of 48 channel walkaway records. The 24 channel dispersion curve is generated using the first half of the FRW. The higher mode jump seen in the 24 channel curve is not present in the FSW curve, because the modes were more effectively separable at far offsets. The FRW curve, however, exacerbates the modal jump. Phase velocity differences between the 24 and FSW curves are probably due to bulk averaging over a longer distance that includes higher velocity material outside the 24 channel record. .................................................................46

Figure 4-7. Example record and dispersion curve from Line 4. Note strong backscattered events across traces 1-4 and 7-12. .........................................................47

Figure 4-8. Line 5 was typically weakly coherent and moderately dispersive.47

Figure 4-9. Same scale S-wave velocity profiles for all but line 7. 1D inversions contoured to produce these images are spaced at the shot spacing indicated on the scale bar. .................................................................48

Figure 4-10. Comparison of results over the same ground using different survey parameters and walkaway constructions. A) 12 m offset 24 channel. B) 36 m offset 24 channel. C) Fixed receiver walkaway. D) Fixed source walkaway. Shot/inversion spacing is 5 stations. ........................................49

Table 4-2. Torlesse greywacke data, presenting shallow surface S-wave velocities for Torlesse argillites and sandstones for different degrees of weathering/fracturing, after (Perrin, 2008, pers. comm). .........................................................50

Figure 4-11: Location of topographic profiles and their relationship to MASW profiles. Image rotated 51° to place fault strike (051) in ‘N-S’ orientation. Topographic
profiles 1-24 after Evans (2000). This figure is formatted to fold out for ease of reference whilst reading this chapter and Chapter 5. ..........................52

Figure 4-12: A northeast facing across site correlation of the MASW profiles with detailed geomorphological surveys and structural interpretations. Significant detail is evident here of shear wave velocity variability within the wider fault zone that can be closely correlated with individual shears and secondary faulting or off-plane deformation. Features to note include the upwarped surface of Terrace B(13) along the fault scarp profiles (7, 10-12), lesser warping of the younger Terrace A strath and subtle gradient changes in the modern river bed coinciding with the high velocity s-wave zone of differentially uplifted bedrock. These features are indicative of persistent deformation for 300-400metres east of the obvious scarp. The shear zone extends well into the footwall. Note: Fault dips exaggerated by 10x VE. Also, a regional gradient to the NE affects precise superimposition of corresponding surfaces........................................53

Figure 4-13. Detailed map (A) and MASW pseudo-section view of lines 1 and 3 (B) in the central part of the Dalethorpe structure showing newly mapped faults. For discussion see text. ..............................................................55

Figure 4-14: Along-strike comparison of MASW line 3 alone (A) with line 7 superimposed on line 3 (B) to investigate repeatability of the technique. The vertical exaggeration is x10 and the horizontal lines are at 1 m spacing. Both MASW profiles can be seen to record some major similarities..........56

Table 4-3 Near-surface physical properties and S-wave velocities at outcrop D1, Dalethorpe. .................................................................58

Figure 4-15: Gross features of outcrop D1 (for location see Figure 4-3 or Map 1). A) General view of fault-zone outcrop cut by meander in south branch of the Hawkins River. B) Close up of outcrop. The footwall is intensely brittley deformed by multitudes of low angle thrusts, a few of which are shown along with a major 1 m wide foliated shear zone. The outcrop is more massive and less shocked on the hanging wall where most of the deformation is concentrated in the argillites. The yellow lines pick out the sheared argillite beds, which are oriented approximately 010/65SE. C) Close up of ductile deformation: Isoclinally folded and boudinaged thin argillite layer within footwall forward of shear zone 1.
Elsewhere, the folds are crosscut by low angle thrusts. This is in direct contrast to the continuous but sheared argillite beds on the more intact hanging wall side of the shear zone. D) The hanging wall scree contains significantly larger blocks of sandstone than does the footwall scree (see B). The blocks do not however transport well and typically fall apart as shown when disturbed........59

Figure 4-16: Graph of V_s with distance along scanline for outcrop D1. No obvious velocity asymmetry is apparent to match field observations but this may be skewed by the predominance of argillites sampled at the SE end of the outcrop. ............................................................................................................60

Figure 4-17. Relationship for outcrop D1 between crosshole S-wave velocities and those derived from MASW. (A) Detailed map view showing MASW lines in structural orientation. A GPR data cube imaging three shears, numbered A1-3 (Corboz, 2004 and Figure 4-2B) is shown in position. Shear A3 can be seen to trend towards crosshole 1 at the NW end of outcrop D1, where it outcrops. (B) S-wave velocity profiles for line 1 and 4 with crosshole locations and low velocity zones marked. Line 1 displays significantly lower velocities, probably due to its proximity to an unconfined edge. Both lines display a deep set of low velocity zones, which are numbered as per Corboz (2004). (C) Marked up photo of D1 showing crosshole locations, shear zones and crosshole velocities according to MASW colour scale. The velocities are clearly more representative of Line 1 than of the more distant Line 4. .................................................................61

Figure 4-18: (A) Perspective view of lines 1 and 4-6. (B) The low velocities in line 1 correlate closely with a low velocity zone in line 6 at the intersection of the lines. Line 1 is shown without velocities in B so that line 6 can be seen behind line 1. The lensoidal character of the velocities in line 4 particularly may be partly due to the low angle shears seen in outcrop D1 (Figure 4-15).................62

Figure 4-19. Correlation of geological strength index (GSI) with MASW S-wave velocities. (A) The location of the photographed outcrops is shown on the MASW profile. (B) Geological strength indices derived in the field for the outcrops in A. (C) Plot of GSI against S-wave velocity for the outcrops in A and B. A strong, apparently linear correlation can be observed. ............................................63
Figure 4-20: Integrated scanline (fracture spacing) and velocity data for Dalethorpe. Crosshole data from Outcrops 1 and 2, and MASW velocities for outcrops 3 and 4 (all diamonds) define a trend that agrees with the data in Table 4-2. MASW data from outcrops 2, 5 and 7 (squares) all lie within the area shown by Figure 4-5 to be potentially higher mode contaminated, and this is reflected in their elevated velocities. Magenta markers are velocities for outcrop 2, green markers are for velocities adjacent to shear zone 1.  ........................................64

(Equation 4-1)........................................................................65

Figure 4-21: Intact S-wave velocities for specimens cored from Torlesse greywacke river boulder sourced from the Hawkins River catchment at Dalethorpe. The unconfined S-wave velocity is slightly anisotropic [10% using Okaya et al’s formula (V_{max} - V_{min})/V_{ave} (Okaya et al., 1995)] but this appears to reduce with compression. For the purpose of this study the unconfined intact S-wave velocity is taken as 2750 m/s. ....................................................................66

Figure 4-22. Ultrasonic S-wave velocity plotted as a function of uniaxial stress for an un-fractured specimen (triangles) and the same specimen with saw cut ‘fractures’. The specimen was cored in orientation 3. ........................................66

Figure 4-23: Comparison of sonic and ultrasonic velocity relationships to linear fracture density. A) Both ultrasonic and sonic (outcrop) velocities decline linearly with fracture density over the range tested. Ultrasonic and sonic velocity fields are widely separated. B) Percentage velocity decline from a theoretical maximum plotted against linear fracture density. A better fit can be seen between the sonic velocity field and the lightly loaded ultrasonic data. Maximum sonic velocity was set at 1200 m/s (Table 4-2) and ultrasonic at 2750 m/s (see Figure 4-21). 68

Figure 4-24: Relationship between A) ultrasonic S-wave velocity and porosity, B) ultrasonic S-wave velocity and point load strength and C) Point load strength and porosity for specimens from intact boulders and tectonised outcrops of Torlesse greywacke. ..................................................69

Figure 5-1: Topographic and river thalweg profiles viewed towards 51° showing the detailed and average traces of the main upstream facing thrust. Profile 13 is run on the B terrace, Profiles 15 and 21 on the A surface (see Figure 4-11). This 10x vertically exaggerated elevation shows a fault dip of 15° to the SE. The

XX
dogleg is seen in this elevation as the sudden decrease in the elevation of the riverbed, due to the river flowing along strike. The apparent re-entrant angle in the projected fault profile reflects the gradient on terrace B and consequential drop in the elevation of the base of the fault scarp. ........................................ 74

Figure 5-2: Map (A) and profile (B) views of the thalweg of the south branch of the Hawkins River. The most notable feature of the profile is the 1.3 m high, 312 m long anomaly in the river bed at 5830 m from the divide............... 76

Figure 5-3: Detail from 1963 Ashburton-Kowai River aerial photograph 3706-3, showing lineations (main picture) and inset the trace of the Bell Hill Fault. The meandering of the Hawkins River as it approaches the fault can be clearly seen, as can its increased sinuosity as it approaches the MUT upstream...................... 77

Figure 5-4: Locations and section views showing the valley shape at surveyed sections across the south branch of the Hawkins River (ve x5). The northern bank can be seen to be almost ubiquitously steeper except in the gorge, where it is higher but of roughly equal gradient. Sections 2, 3, 8 and 9 are located very similarly relative to active faults and are all distinctly wider due to meandering. With the exception of the gorge profiles all profiles end on the edge of the A surface...... 80

Table 5-1 Riser heights and calculated characteristics for incision events at Dalethorpe.
............................................................................................................ 87

Figure 5-5: Oblique elevation of topographic profiles (numbers shown) showing coseismic uplift markers at Dalethorpe. Vertical exaggeration x10, all dimensions in metres. B0 and C0 represent the correlatives of the B and C surfaces on the downthrown side. Surfaces B1-4 and B6/A0 record periods of quiescence and varying degrees of lateral planation separating coseismic uplift events that led to abandonment of the previous surface. B3, shown horizontal, is the most extensive and best preserved of these surfaces. For profile locations see Figure 4-11. 89

Figure 5-6. Auger hole log for OSL sample taken from surface B3 south of fault trace. Auger-hole log shows horizons, colours, thicknesses and textures, as well as sampling depths. Both samples are taken from the same auger hole. 91

Figure 5-7: Extrapolated age of B3 surface, assuming fluvial / loess transition at first gravels ................................................................. 92
Figure 5-8. Stress changes and earthquake sequence. (a) Regular sequence. (b) Irregular sequence caused by the changes in loading rate and temporal variations in the strength of crust (after Kanamori and Brodsky, 2004, p.1437) .......... 94

Figure 5-9: Selected fault/earthquake parameter relationships after Wells and Coppersmith (1994). A surface rupture 11 km length on a reverse fault is related to displacements of 1 m (A), and a rupture magnitude of M6.3 (C), not greatly different to the M6.6 magnitude associated with a displacement of 1.5 m (B). On this basis, a likely earthquake magnitude is M6.35 ± 0.15 ............... 95

Figure 5-10: A) A slab of sponge as an analogue for a deforming slab of greywacke. The top two lines on the face represent the progressive deformation of a strath originally derived from a river profile graded across the slab at intervals during translation of the slab up the fault plane. Note the rapid steepening of the upper strath in the compression zone, compared with surfaces cut close to or below the neutral surface. B) particle motion drawing showing the behaviour of incrementally deformed straths close to the neutral surface. The effect is similar to that seen in the sponge analogue. ........................................ 98

Figure 6-1: Location map for the Boby’s Stream study area in North Canterbury. 101

Table 6-1: A summary of the brittle microstructures developed in Oligocene limestones of North Canterbury, their relative sequence of development, and estimated ages (after Nicot, 1992). ................................................................. 102

Figure 6-2: Same-scale GPR profile and trench log across the Boby’s Stream Fault. Faults were inferred over a considerable distance along the line including at the locations marked in A. Two of these are clearly discontinuous sand lenses. ... 104

Table 6-2: Summary of mean laboratory-determined physical properties of soft rock lithologies from the Boby’s Stream Fault, with values from Rewanui mudstone (Campbell, 2008) and Torlesse greywacke (this study) for comparison. The predicted sonic S-wave velocity is calculated by subtracting the ultrasonic velocity percentage below 2750 m/s (Torlesse maximum ultrasonic velocity – this study) from 1100 m/s (fresh Torlesse sonic velocity). This assumes that ultrasonic velocities for intact rocks fall on a common strength-related trend and any decline is reflected by sonic velocity decline. ........................................... 106
Figure 6-3: Correlation of ultrasonic S-wave velocity with UCS for four specimens cored from samples of Amuri Limestone and Waikari Siltstone. The Amuri outlier at 13.5, 1670 failed in UCS testing along a pre-existing fracture. .......................... 107

Figure 6-4: Correlation of ultrasonic S-wave velocity with UCS. The relationship splits into two fields, one dominated by the mortar and Mt Brown sandstone specimens. For discussion see text. ................................................................. 108

Figure 6-5: Observed velocity declines due to incremental increases in fracture density for three soft rock lithologies and comparative data for Torlesse Greywacke. Each trend is accompanied by the UCS of the sample remnant tested after the fracture experiment. ................................................................. 109

Figure 6-6: S-wave velocity decline at fracture density of 50/m, plotted as a function of uniaxial compressive strength. This plot suggests that a threshold exists at around 20 MPa, above which fracture density rapidly exerts a controlling influence on S-wave velocity and below which S-wave velocities are otherwise controlled. 110

Equation 6-1 ...................................................................................... 111

Figure 6-7: Proposed relationship between fresh sonic S-wave velocity and UCS. This graph is plotted using the predicted velocities in Table 6-2. These will be compared later with the MASW survey velocities. ......................... 111

Figure 6-8: A) Nail gun penetration displays an almost linear relationship with Schmidt Hammer rebound number in laboratory testing. B) A strong relationship is also seen between UCS and nail penetration............................................. 112

Figure 6-9: Predicted exponential relationship of S-wave velocity to nail penetration. This relationship does not, however account for threshold effects, which may perturb the velocity decline with nail penetration................................................. 113

Table 6-3: Predicted unconfined S-wave velocity ranges based on laboratory velocity testing and nail gun penetration tests at free outcrop faces............ 113

Figure 6-10: Summary figure for Table 6-3. These velocities straddle the higher alluvium and lower bedrock near surface velocities reported by Fumal (1978) and do not fit comfortably in either field................................................................. 114

Figure 6-12: Photograph looking west along the Boby's Stream Fault at the Randolph Downs site and illustrating the complicated (from an MASW perspective)
topography and structure. The visible section of survey line RD1 is about 30 m

long................................................................. 118

Figure 6-13: Photo towards 90° of sheared Loburn Formation showing sheared relict
bedding planes at <1 cm spacings picked out by limonitic weathering. The whole
outcrop is chopped up by small to medium scale conjugate shear planes that
render it virtually isotropic (white circle shows typical debris), but discrete folded
shear planes such as the white line can be traced through the outcrop. A fault
juxtaposing SE dipping Conway formation against the steeply north-dipping
Loburn Formation is located just outside the right of the main photograph. For
photo location see Figure 6-14............................................. 118

Table 6-4: Acquisition parameters for Boby's Stream MASW surveys (with ref to Fig
3.4). ............................................................................... 121

A Table 6-5: Typical fundamental mode bandwidths for Boby's Stream MASW surveys.
Coherent energy in many cases covered ranges of only 10 Hz within these ranges.
....................................................................................... 122

Figure 6-16: Shot records and associated swept frequency records from D. A neutral
position in the line at shot gather 243 (A) shows little scatter compared with the
fault related scattering of record 253, collected astride the fault (B). The scattering
has little apparent influence on the velocities (event slopes) in the swept frequency
record.............................................................. 122

Figure 6-17: Quarry dispersion curves extracted from records acquired over A) fill and B)
the quarry floor. The response seen in B is comparable to dispersion curve (C)
(after O'Neill and Matsuoka, 2005) modelled using the synthetic layered earth S-
wave velocity model (D) of Tokimatsu et al. (1992). The high velocity cap
between 2 and 5 m in the layer model causes excitation of higher modes with
significant energy over the corresponding wavelengths.............. 124

Figure 6-18: Progression of 5-40Hz bandpass filtered surface wave records from seismic
line RD1. The array moved from left to right and the heavy black lines are an
arbitrarily chosen common point. The red line is the source of the backscatter,
which has greatest amplitude when the array butted up to the red line. Backscatter
and noise quickly disappears after the array passes over the backscattering
contrast, located at station 80........................................... 124
Figure 6-19: Boby's Stream MASW profiles and laboratory-predicted, weathered through intact S-wave velocities for comparison. Countouring is in Surfseis unless otherwise indicated. 3 separately processed profiles are presented for D1, to demonstrate the repeatability of the technique. All profiles contoured between 100 and 600 m/s. S-wave velocity colour scale shown is for Surfseis profiles only. Not all Surfer-corrected profiles are shown............................................. 126

Figure 6-20: Marked up photograph towards 90° of the visible (solid black line) and approximate (dotted black line) base of the Pleistocene gravel cover on The Deans point bar. The Mount Brown/Waikari contact is marked in white and is upthrown to the south by the fault in the centre of the picture, which crosses the northern end of Line D1 (purple line). For location see Figure 6-14. [Photo taken from Limestone Hills homestead by Gareth Renowden, Limestone Hills].129

Figure 6-21: Photo towards 90° of bouldery gravels outcropping against residually weathered Mt Brown Formation in a farm track adjacent to The Deans homestead. Limestone boulders are deposited over greywacke cobbles. The gravels terminate ~2-3 m to the right of the photograph where they give way to sands. A band of limestone boulders can be seen outcropping in the driveway above, which is level with the wagon wheels............................................. 130

Figure 6-22: An east-facing across site correlation of the Boby's Stream MASW profiles with detailed geomorphological surveys and structural interpretations. All dipping features exaggerated in the section view by 5x VE. Regional surface gradient is to the east. A back pocket supplement to this diagram shows both plan and elevation views and the location of topographic profiles............................................. 131

Figure 6-23: Detailed map (A) and profile (B) view of the thalweg of the Waipara River for 2 km south of the gorge. The profile (B) includes both the surveyed river bed (dark blue) and the approximate water level (light blue) Coloured markers are placed along the meander between The Deans and Limestone Hills to allow the reader to correlate between map and profile views.............................. 133

Figure 6-24: Photograph c.1918 looking east across The Deans meander bend directly along the crest of the Onepunga Anticline. Although tree growth precludes taking a comparative photograph, the river can be seen to occupy the full width of its bed as it does today. Judging from the scale of the buildings, of which the main house
in the background and the woolshed in the foreground are still existing, at least a metre or two of downcutting has since occurred along the front of the homestead. Photo by permission of Ian Turnbull, The Deans. .......................... 134

Figure 7-1: (A) Google topographic map showing the location of the Taieri Ridge and study area and (B) geological map (after Kilner et al., 2007) of the Sheehy Road Field area on the southeastern side of the Taieri Ridge.................. 139

Figure 7-2: Oblique aerial photo looking north showing the Taieri Ridge with the parallel Rock and Pillar Range structure in the background. Inset detail shows field area and location of seismic line. The uphill extent of the line stopped in the hanging wall of the inferred fault at the outcropping schist. Main photo by Lloyd Homer (used with permission of GNS Science). Inset photo by Jeremy Kilner.141

Figure 7-3: HRR processing flow leading to production of final stacked section (after Kilner et al. (Kilner et al., 2007)) ......................................................... 142

Figure 7-4: Refraction static model showing (1) surface topography, (2) ‘weathering’ layer thickness and velocity, and (3) second layer velocity......... 142

Figure 7-5: CDP stacks that use a basic velocity model based on the refraction statics in Figure 7-4. A: stack with no muting or f-k filtering. B: stack with surgical mute designed to remove ground roll. C: stack B with f-k filter applied. D: stack with severe mute to remove ground roll but no f-k filtering. ...................... 143

Figure 7-6. Trench photo panorama and log (modified after Kilner (2008)) for Sheehy Road Trench 1. This trench was sited on the basis of interpretation of the HRR survey described in 7.3.2. As predicted by the results of the MASW survey the major Taieri Ridge fault was not located although abundant evidence of minor faulting was seen southeast of the unconformity. ......................... 144

Figure 7-7. Sheehy Road survey line location (yellow) relative to the inferred fault (dotted red). No scarp is developed in the survey area. ......................... 146

Table 7-1. Acquisition parameters for Taieri Ridge HRR survey and effective parameters for MASW surveys................................................................. 146

Figure 7-8: Dispersion curves for the A) start [east], B) middle and C) end [west] of the seismic line. A systematic shift can be observed, that relates to changes in the character of the subsurface.............................. 148

xxvi
Figure 7-9: Raw unprocessed dispersion images either side of record 46 (outlined), showing the change in character between record 40 (top left) and record 58 (bottom right). Record 41 (top centre) is located at the western end of the trench where schist was uncovered and shows higher mode completely dominating fundamental mode dispersion. Aside from that, the fundamental mode is progressively swamped by higher mode dispersion before re-emerging with a much more strongly dispersive character. ................................. 149

Figure 7-10. Comparison of MASW profiles derived from the same seismic data but using single shot and walkaway configurations. A) 5m offset 38 channel single shot survey with higher mode contamination. B) 5 m offset 86 channel FSW survey. C) 5 m offset 38 channel reprocessed in the light of B and contoured in Surfer8 to produce a crudely ‘elevation corrected’ profile. .................. 151

Figure 7-11: S-wave velocities indicating the potential for S-wave splitting in Alpine Schist (after Okaya et al., 1995). The anisotropy and velocities of the NZ1 Alpine Schist mylonite are very similar to the NZ 3 and 4 Haast Schist. Even at low pressures, propagation parallel to foliation is approximately 1 km/s faster than propagation perpendicular to foliation. This has implications for the planning, execution and interpretation of MASW in a schistose setting. .......... 152

Figure 7-12. Sheehy road seismic line station-to-station gradients indexed against the end-to-end gradient of the line. ...................... 153

Figure 7-13: Correlation of the elevation corrected MASW profile with the original HRR profile. .............................................................. 154

Figure 7-14: Swept frequency image of the northwest end of the Taieri Ridge seismic line A) without and B) with interpretation. The black seismic lines are surface wave events of the frequency scaled on the left of the image. The slope of the line is inversely proportional to the velocity of the event. Red lines are faults or shears, whilst the blue line is a velocity-based pick of the location of the unconformity. The lateral location of the trench is shown, together with the location of major unit boundaries observed in the trench. For discussion see text. .......... 156

Figure 7-15: Dynamic Linear Moveout (DLMO) stacked section of the Taieri Ridge seismic line. Rectangle is trench location and line indicates possible imaging of faulting. ...................................................................... 159

xxvii
Table 8-1: Summary table of predicted S-wave velocities for all lithologies encountered in this study..........................................................164

Figure A-1: The effect on Rayleigh-wave phase velocities of 25% change in density (squares), P-wave velocity (diamonds), S-wave velocity (circles) and layer thickness (triangles) in a layered earth model. S-wave velocity can be seen to dominate the dispersion curve, whilst P-wave velocity has little effect (after Xia et al., 1999)...................................................................................188

Figure A-2. An example from Xia et al. (2003) illustrating the effect of fundamental dispersion curve error on the inverted model. 'Measured' curves are extracted from the shot record, whilst 'Final' curves are calculated from the final S-wave velocity model. A) Accurate fundamental mode dispersion curve. B) Erroneous fundamental mode dispersion curve with higher mode and body wave contamination deliberately introduced between 13-19 Hz. C) As B but with the inclusion of higher mode data from 20-30Hz. D) The erroneous fundamental mode data produces an irrational profile (diamonds) except where it is supplemented by the higher mode dispersion curve (triangles). In the latter case, the S-wave velocity model closely follows that of the accurate fundamental mode curve (squares)................................................................................189

Figure A-3: Synthetic shot gathers with a) fundamental mode only and b) fundamental and first higher modes. The higher mode is assigned overall energy of approximately 2 x fundamental mode (after Park et al., 2002). The low frequency fundamental mode events overlap with the higher mode events, whilst the higher frequencies of fundamental mode inhabit a lower velocity field in the x-t domain. ........................................................................................................191

Figure A-4: Muting of higher modes in the x-t domain. 'Records' 1 and 2 have been cut from record 3 (Red and blue boxes). The longer the array, the easier to visually identify and mute the velocity fields dominated by higher mode and body waves. Note, however, that the higher mode dominated velocity field will still contain the low frequency (high velocity) component of fundamental mode Rayleigh-waves. This problem is addressed by the dispersion curve extraction technique of Ivanov et al. (Ivanov et al., 2005). ........................................................................192
Figure A-5. Application of velocity (pie slice) and bow-slice (polygonal) filters in the f-k domain. The amplitude spectrum for fundamental mode and higher mode (A) (inside the green slice) is that of Figure A-3 (b). Application of a velocity filter in situation (A) will remove all but the higher frequency components (hatched) of the fundamental mode wavefield, due to the overlapping velocities of fundamental and higher modes at low frequencies. Although the velocity filter works better in situation (B), the same result in both situations could be more quickly achieved by muting in the x-t domain. The bow slice filter, however, is defined by a band around a rejection zone that follows a trajectory in f-k space, thus avoiding the fundamental mode entirely. It should work well in both situations (Modified from Park et al., 2002). ................................................................. 193

Figure A-6: a) A vertical fault model with a source on the left side of the fault. b) A synthetic shot gather due to the fault model. c) FK filtered data with a diffraction travelt ime curve. c = 190 m/s and tx = 0.237 s are the average phase velocity and the travel time at x = 14 m, respectively (after (Xia et al., 2007)) .... 195

Figure A-7: A) Dispersion curve generated forwardly from a model containing a low velocity layer at a depth of 15-25 m. B) The low sample density (5Hz) dispersion curve fails to produce the low velocity layer whilst the higher sample density (1Hz) is in good agreement with the model (Zhang and Chan, 2003).195

Figure A-8: MASW survey geometry using a land streamer. Key acquisition parameters are illustrated and selection of these is discussed in the text. .......... 196

Figure A-9: Rayleigh wave propagation away from a generation point. Near to the source the wavefront is cylindrical, whilst further away it becomes planar (after Park and Miller, (Park and Miller, 2006)) ................................................................. 198

Figure A-10. A 5 s long swept frequency display of a shot gather collected using an 8 kg sledgehammer at Dalethorpe. Traces can be seen to be strongly coherent from about 10 Hz through to about 30 Hz where far field effects are established. The dispersion should therefore be imaged from 10-30 Hz. ............... 198

Figure A-11: Construction of 48 channel walkaway records by concatenation of A) files with a common mid-point and differing offsets (Fixed Receiver Walkaway) and B) files with a common shot point (Fixed Source Walkaway). ........ 199
Map 1: Tectonic Geomorphology and Structure of the Upper Hawkins Basin, Dalethorpe (1:5000)..........................201

Map 2: Geology and Geomorphology of the Boby’s Creek Study area (1:5000)201

Figure 6-22 supplement: An east-facing across site correlation of the Boby’s Stream MASW profiles with detailed geomorphological surveys and structural interpretations. All dipping features exaggerated in the section view by 5x VE. Regional surface gradient is to the east..........................201
ABSTRACT

Bulk rock strength is greatly dependent on fracture density, so that reductions in rock strength associated with faulting and fracturing should be reflected by reduced shear coupling and hence S-wave velocity. This study is carried out along the Canterbury rangefront and in Otago. Both lie within the broader plate boundary deformation zone in the South Island of New Zealand. Therefore built structures are often, located in areas where there are undetected or poorly defined faults with associated rock strength reduction. Where structures are sited near to, or across, such faults or fault-zones, they may sustain both shaking and ground deformation damage during an earthquake. Within this zone, management of seismic hazards needs to be based on accurate identification of the potential fault damage zone including the likely width of off-plane deformation. Lateral S-wave velocity variability provides one method of imaging and locating damage zones and off-plane deformation.

This research demonstrates the utility of Multi-Channel Analysis of Surface Waves (MASW) to aid land-use planning in such fault-prone settings. Fundamentally, MASW uses surface wave dispersive characteristics to model a near surface profile of S-wave velocity variability as a proxy for bulk rock strength. The technique can aid fault-zone planning not only by locating and defining the extent of fault-zones, but also by defining within-zone variability that is readily correlated with measurable rock properties applicable to both foundation design and the distribution of surface deformation. The calibration sites presented here have well defined field relationships and known fault-zone exposure close to potential MASW survey sites. They were selected to represent a range of progressively softer lithologies from intact and fractured Torlesse Group basement hard rock (Dalethorpe) through softer Tertiary cover sediments (Boby's Creek) and Quaternary gravels. This facilitated initial calibration of fracture intensity at a high-velocity-contrast site followed by exploration of the limits of shear zone resolution at lower velocity contrasts.

Site models were constructed in AutoCAD in order to demonstrate spatial correlations between S-wave velocity and fault zone features. Site geology was incorporated in the models, along with geomorphology, river profiles, scanline locations and crosshole velocity measurement locations. Spatial data were recorded using a total-station survey. The interpreted MASW survey results are presented as two dimensional snapshot cross-
sections of the three dimensional calibration-site models. These show strong correlations between MASW survey velocities and site geology, geomorphology, fluvial profiles and geotechnical parameters and observations. Correlations are particularly pronounced where high velocity contrasts exist, whilst weaker correlations are demonstrated in softer lithologies. Geomorphic correlations suggest that off-plane deformation can be imaged and interpreted in the presence of suitable topographic survey data. A promising new approach to in situ and laboratory soft-rock material and mass characterisation is also presented using a Ramset nail gun.

Geotechnical investigations typically involve outcrop and laboratory scale determination of rock mass and material properties such as fracture density and unconfined compressive strength (UCS). This multi-scale approach is espoused by this study, with geotechnical and S-wave velocity data presented at multiple scales, from survey scale sonic velocity measurements, through outcrop scale scanline and crosshole sonic velocity measurements to laboratory scale property determination and sonic velocity measurements. S-wave velocities invariably increased with decreasing scale. These scaling relationships and strategies for dealing with them are investigated and presented.

Finally, the MASW technique is applied to a concealed fault on the Taieri Ridge in Macraes Flat, Central Otago. Here, high velocity Otago Schist is faulted against low velocity sheared Tertiary and Quaternary sediments. This site highlights the structural sensitivity of the technique by apparently constraining the location of the principal fault, which had been ambiguous after standard processing of the seismic reflection data. Processing of the Taieri Ridge dataset has further led to the proposal of a novel surface wave imaging technique termed Swept Frequency Imaging (SFI). This incoherent technique apparently images the detailed structure of the fault-zone, and is in agreement with the conventionally-determined fault location and an existing partial trench. Overall, the results are promising and are expected to be supported by further trenching in the near future.
1 INTRODUCTION

New Zealand has a growing population. It is expected to grow by up to 1.3 million people by 2026, which represents a 35% increase on 2001 figures (Statistics-New-Zealand, 2004). This increased population must be provided with safe housing. However, New Zealand's tectonic setting, outlined in 1.3 below, poses challenges to planners attempting to zone land for residential development. Basins such as Canterbury are often afflicted with poor outcrop, indistinct marker horizons and a lack of subsurface data. This adversely affects the resolution at which fault-zones can be characterized using conventional geomorphic and structural mapping (Litchfield et al., 2003). This study will investigate and put forth a method of improving planning efficacy by enhancing the reliability and resolution with which active faults are located in New Zealand's active landscapes.

1.1 Motivation: Present day land use planning in New Zealand fault-zones.

Land use planning within active tectonic settings is fraught with difficulty. Buildings sited across faults that rupture will invariably sustain the greatest damage during an earthquake, closely followed by those buildings in close proximity to the fault (eg Kelson et al., 2001). A 2001 report, entitled 'Building on the Edge: The use and development of land on or close to fault lines' (Williams, 2001, BOE report) and released by the office of New Zealand's Parliamentary Commissioner for the Environment raised concerns about the ability of legislation such as the Building Act 1991 (since replaced by the Building Act 2004) and the Resource Management Act 1991 (RMA) adequately to manage the development of seismically hazardous land. The report was initially motivated by concerns regarding the safety of the Winara Village retirement complex on the Kapiti Coast, where inappropriate provisions in the Kapiti Coast District Plan coupled with scientific uncertainty in determining the location of the fault placed the development in a potentially precarious situation.

Parties consulted in the preparation of the BOE report considered earthquake hazard planning to be a two stage process involving firstly the mapping of active faults and secondly site-specific, pre-development, geotechnical investigations. The latter stage is typically expensive but nevertheless essential. Whilst these parties agreed on a need to prevent construction over active faults, they also expressed concern regarding a lack of consistency of approach in site-specific investigations due to lack of guidance,
appropriate criteria and expertise. The BOE report identified a need for clear guidance on what a site-specific investigation should entail and concluded that it was important for local authorities to base their assessment and management of seismic hazards on accurate and relevant scientific information.

Following on the publication of the BOE report, a working party of relevant professionals developed a set of guidelines to assist New Zealand’s resource management planners and these were published as a report entitled ‘Planning for Development of Land on or Close to Active Faults’ (Kerr et al., 2003, The Guideline). The Guideline sets out a risk-based planning approach based on the three elements of recurrence interval, fault complexity and building importance category.

The second step of this approach after fault identification is creation of fault avoidance zones around active faults. Section 6 of The Guideline defines a fault avoidance zone as a 20 m buffer either side of the active fault trace or likely fault rupture zone, with provision to increase this based on an assessment of fault complexity. Fault complexity in this instance is geomorphically defined but difficulties may arise when attempting to quantify the extent of a fault-zone and its associated deformation, which may be considerably degraded or buried by young sediments (Hart and Bryant, 1997; Similox-Tohon et al., 2006). This problem may directly impact uptake of The Guideline as shown by Becker et al. (2005). A key issue raised by their questionnaire returns in a study of uptake of The Guideline was that of justification. An interviewee responded that local authorities had difficulty justifying 'heavy-handed' implementation of the structured guideline approach where fault scarps were not well defined, especially without additional [subsurface] information.

The simple fault avoidance zone specified in The Guideline encompasses a zone 20 m either side of the fault trace or likely fault rupture zone. Oglesby et al. (1998) however, in a widely cited study, report an asymmetry in expected ground motion either side of dip-slip faults. For instance, they state that, given identical stress magnitudes, the intensity of motion depends on both fault sense (thrust fault ground motions surpass those of normal faults by 100%) and on location relative to the fault (hanging wall motions are larger than footwall motions for both normal and thrust faults). Duan and Oglesby (2005) further report that irrespective of dip sense, horizontal ground motion dominates on the footwall whilst vertical ground motion dominates on the hanging wall. This suggests that fault
avoidance should take greater account of fault geometry, which in the case of a buried fault may be less than clear and requiring information derived by subsurface investigation.

Although the Guideline is limited by the somewhat arbitrary 20 m setback criteria, the problems associated with active fault ground rupture and deformation do at least come within its scope. The Guideline, however, fails to address related issues such as strong ground shaking and liquefaction. The differential high intensity shaking mentioned above will extend beyond the immediate range of a few tens or even hundreds of metres and therefore cannot be completely mitigated by land zoning as such, although strategies reviewed in Chapter 2 have shown significant success in this area by classifying the shaking potential of sites. No such program is suggested by the Guideline and no further guidance is to be found in the Building Code (DBHNZ, 2007), set out in the first schedule to the Building Regulations (1992). The Building Code requires simply that “Buildings, building elements and sitework shall have a low probability of rupturing, becoming unstable, losing equilibrium, or collapsing during construction or alteration and throughout their lives.” (Clause B1.3.1). It further states that “Account shall be taken of all physical conditions likely to affect the stability of buildings, building elements and sitework including: ... earthquake” (Clause B1.3.3(f)) with due allowance required to be made for site characteristics and inaccuracy in performance prediction (Clause B1.3.4(d) and (e)). These performance requirements, for building elements imparting structural stability, extend to the life of the building, which unless specified is not less than 50 years (Clause B2.3.1(a)(i)). The Building Code does not however contain any specification of site classifications such as are found in the United States’ National Earthquake Hazard Reduction Program (NEHRP) building provisions (BSSC, 2006).

A consideration which deserves more immediate planning attention within a fault-zone is the extent of ongoing and coseismic secondary off plane deformation, particularly in the hanging wall of thrusts and transpressional faults. The internal structure of faults is not limited to a simple planar surface but incorporates a damage zone that may extend to several hundreds of metres within which complex distributed deformation may take place. Kelson et al. (2001) carried out detailed surveying of styles of deformation and related structural damage following the 1999 Chi-Chi earthquake in Taiwan. They recorded a deformation zone with a width that was typically 10-20 times the net vertical displacement of up to 4 m. In some instances they observed distributed secondary faulting and folding well into the hanging wall up to 350 m from the scarp. Such well
documented observations demonstrate that the extent of folding, tilting and secondary faulting in a variety of configurations may extend well beyond a 20 m corridor relative to the main rupture plane. A clear planning advantage would be gained by not only identifying the projected position of the fault trace but also identifying the width of the zone of damaged rock caused by off plane deformation.

Control over property development in New Zealand is devolved to the regional councils and the RMA requires regional councils to prepare a regional policy statement on issues including natural hazards. The current Canterbury Regional Policy Statement (ECAN, 1998) predates publication of the BOE report and The Guideline. However, a scoping report released in 2006 (ECAN, 2006) recommends a comprehensive review of the relevant chapter to include more precise policies to guide risk management. In addition, Environment Canterbury commissioned Opus International Consultants Limited to develop a draft risk assessment methodology for Christchurch (Brabhaharan et al., 2005). Although this document only addresses the risk in the Christchurch area from fault sources lying predominantly outside Christchurch, the proposed methodology requires the acquisition of ground shaking and liquefaction data. Both of these factors have been related to S-wave velocities (Anderson et al., 1996; Benjumea et al., 2003; Yunmin et al., 2005) (see section 2.4.3).

Despite the weaknesses of The Guideline, the first of its four planning principles calls for the gathering of accurate active-fault hazard information. The Guideline’s authors acknowledge that gathering such information will require specialized scientific knowledge and surveys but note that, given the major effects of such information on any property development decisions, the information must be as accurate as technology and resources permit. A successful outcome of this study has practical application to the identification of obscured fault-zones and the siting of buildings, as well as contributing to fundamental research into the nature of high-level fault processes. It will validate Multichannel Analysis of Surface Waves (referred to throughout the remainder of this document as MASW) as a rapid, low-cost, scientific basis for fault-zone planning that is complementary to, and if necessary independent of, geomorphological interpretation, thus making a valuable contribution to New Zealand’s fault risk management practices. Additionally, data generated by shallow seismic characterization of rock mass properties will be attractive in subsequent applications such as building foundation design, waste isolation or water containment in fractured rock.
1.2 Objectives

The primary objective of this study is to prove the utility of MASW to aid fault-zone planning by locating and defining the extent of fault-zones in several typical New Zealand settings, and defining within-zone variability that is readily correlated with measurable rock properties and with the distribution of surface deformation. Fundamentally, the method derives a near surface profile of S-wave velocity variability that is taken as a proxy for bulk rock strength.

This goal is attacked by addressing three secondary aims, including:

1. Investigation of the influence of shearing and cataclasis on the physical properties of a relatively uniform, high compressive strength rock type (Torlesse greywacke), by:
   - Initial definition of a multi-scale velocity structure of the Springfield Fault at Dalethorpe (SFD), Hawkins River Valley, South Canterbury and relation of this structure to fractured Torlesse sandstone engineering rock properties, and
   - Correlation of the interpreted (in terms of structure and rock properties) MASW profiles with the geomorphic expression at the outcrop of the SFD.

2. Exploring the limits of resolution of the method when applied in soft rock and discriminating variance due to primary lithologies from the effects of cataclasis, by:
   - Testing and recalibration of MASW at Boby’s Stream Fault, Middle Waipara, where at different locations there is juxtaposition of a range of weaker lithologies including limestone, siltstone and greensand. This culminates in a test of the resolution at a complex site where two different siltstones are separated by greensand between two fault strands, with anticlinal bulging of the terrace between the strands.

In both 1 and 2 existing river exposures are used to correlate MASW data with both in-situ and laboratory testing, and with the detailed structure of the fault-zones.

3. Blind application of the MASW technique is used predictively to characterise a further fault-zone at Sheehy Road, Macraes Flat, Otago.
The sites have been chosen because they have been subject to previous investigation. Their field relationships are thus known and they fulfill the requirements of the stated objectives.

1.3 New Zealand’s tectonic setting

New Zealand’s location astride a plate boundary is well documented (Figure 1-1). The Pacific plate rotates anticlockwise relative to the Australian plate causing systematic variation in the orientation and partitioning of convergence along the plate boundary system. This results in along strike changes in deformation styles, from oblique subduction at the Hikurangi Margin and back-arc rifting in the Taupo Rift in the North Island, through an almost purely strike slip transfer zone in the faults of the Marlborough fault system (MFS) of the northern South Island. The MFS links the Hikurangi Margin to the Alpine Fault, where oblique convergence is building the Southern Alps in the central South Island. The onshore New Zealand plate boundary finishes with an opposite-sense subduction of the Australian plate under Fiordland at the northern termination of the Puysegur trench (Pettinga et al., 2001; Wallace et al., 2007).

Active faulting due to accommodation of plate boundary motion poses hazards countrywide, particularly where fault-zones approach or transect inhabited or subdividable areas. This study is based in the central South Island, south of the MFS, north of Dunedin and east of the Alpine Fault (see site localities Figure 1-1). The area is particularly suitable for this research, as faults that offset and/or juxtapose a variety of lithologies, and that display late Quaternary and Holocene activity have been well documented in Canterbury (Cowan, 1992; Howard et al., 2005; Litchfield et al., 2003; Pettinga et al., 2001) and Otago (Litchfield and Norris, 2000; Norris and Nichols, 2004).

1.4 Tectonic setting of central South Island and study sites

According to the NUVEL-1A global plate motion model of DeMets et al. (1994), the central South Island accommodates 37 mm/yr of convergence on an azimuth of 071° (calculated at Franz Josef by Norris and Cooper (2001)). This convergence is expressed by varying styles of faulting in Canterbury and Otago. In North Canterbury, the southern boundary of the MFS has migrated south over time and is currently dominated by the Hope Fault. An incipient new southern boundary to the MFS is, however, deforming the northern margin of the Canterbury Plains along a zone of dextral oblique shear that has
been termed the Porters Pass Amberley Fault-zone (PPAFZ - Figure 1-1, Figure 1-2) (Campbell et al., 2003; Cowan, 1992).

Figure 1-1: Tectonic setting and main structural features of the New Zealand microcontinent bisected by an obliquely convergent plate boundary between the Australian and Pacific plates. Study sites are shown, numbered 1-3. Arrows show rates of relative convergence in mm/yr. Plate boundary motion and partitioning calculated at Franz Josef by Norris and Cooper (2001) according to NUVEL-1A global model (DeMets et al., 1994). Map modified after Pettinga et al. (2001) and Campbell et al. (2003).

Two of the sites investigated in this study lie at opposite ends of the east-northeast trending PPAFZ, a hybrid system of strike-slip and thrust/reverse faults (Pettinga et al., 2001). Study area 1 is a strand of the Springfield Fault outcropping in the Upper Hawkins Basin at Dalethorpe, 7 km south west of Springfield. The mapped fault is a northwest
facing thrust with no evidence of strike slip offsets that falls within the PPAFZ (Figure 1-2 B). Its structure, however, is influenced by its location on a very close domain frontier between the PPAFZ, the Mt Hutt – Mt Peel [thrust] Fault-zone and the South Canterbury Zone as defined in Pettinga et al. (2001). The faults in the latter two zones are typically east facing. Lithologically, the Springfield fault juxtaposes different crustal levels of Torlesse greywacke, within the limited area covered in this study.

![Figure 1-2: Tectonic setting and main structural features of North Canterbury. A) Location relative to the subduction/transpression transition zone between the Hikurangi Margin and the Marlborough Fault System (MFS, stippled). Other zones (after Pettinga et al., 2001) are the West Culverden Fault-zone (WCFZ), the Porters Pass Amberley Fault-zone (PPAFZ), and the North Canterbury Fold and Fault Belt (NCFFB). B) Locations of the Dalethorpe (1) and Boby's Stream (2) study areas at either end of the PPAFB. (modified after Litchfield et al., 2003).](image)

The second study area is the Boby's Stream Fault in North Canterbury. Again part of the PPAFZ but located at the opposite end, this fault is more stereotypical of the domain and is structurally and lithologically dissimilar to the Dalethorpe site. The Boby's Stream Fault is dominantly strike slip with a significant transpressional component of northward thrusting. At various locations along its length it juxtaposes many of the typical lithologies of the South Island Tertiary succession, all of which are significantly softer than intact Torlesse Greywacke.

The third and final study area is located well south of the first two at Sheehy Road, along the Taieri Ridge in the Macraes Flat area (Figure 1-3). This area, which is shortening at ~1.5 mm/year, is characterized by basement schist thrust over Quaternary alluvium and
colluvium and patchy Tertiary remnants (Kilner et al., 2007; Norris and Nichola, 2004). The base of the southeastern slope of the Taieri Ridge structure harbours a concealed fault-zone, with no clear indicators of its position. As such it provides an admirable final site on which to test MASW.

![Map of Central Otago with inset South Island location map showing the actively rising northeast trending ranges of Central Otago. The Taieri Ridge is located on the southeastern margin of this system and the study site is marked with a white square. Cross section after Norris and Nicol (2004)](image)

1.5 Study Format

A literature review is presented in Chapter 2 that looks first at the relevance of S-wave velocity data to fault-zone planning. It then outlines the geotechnical applications of S-wave velocities, including a review of the technical bases for correlation of S-wave velocities with geotechnical parameters. This section also encompasses a review of rock mass classification systems for tectonized rock masses and briefly considers photogrammetric characterization techniques. Having established the relevance of S-wave velocity data, the review turns to the techniques by which the data are acquired, concentrating on those relevant to this study.

Chapter 3 outlines the geomorphic, geotechnical and seismic methods employed in the various aspects of this study.
Chapters 4 and 6 present self-contained accounts of the MASW calibrations at Dalethorpe and Boby’s Stream respectively. These chapters detail the site context and site-specific geology, review previous work on the sites, present details of the data and results obtained and report the correlations observed at these sites. Chapter 5 presents a structural and paleoseismic analysis of the Springfield Fault at Dalethorpe in the light of the work presented in Chapter 4.

Chapter 7 details a blind-test of MASW at the site at Sheehy Road, Macraes Flat in Otago. Although the test site was originally intended to be in similar lithologies to those encountered in Chapters 4 and 5, the Sheehy Road site presented an unexpected opportunity to expand the range of lithologies whilst fine tuning and blind-testing the techniques developed at the previous sites. It also offered the opportunity to reprocess an existing dataset over a concealed fault that had already been processed as a high resolution reflection survey. Chapter 7, therefore, includes a comparison of the survey methods in terms of data density (hence acquisition time), processing time, interpretability and results. Finally, it presents a first look at a technique developed in this study and not previously reported in the literature that promises to reduce shallow seismic survey times over an active fault by an order of magnitude, whilst providing exceptional lateral resolution.

Chapter 8 presents a general summary of the study and a discussion of the similarities and differences between sites. In addition to discussing points arising from the test sites, it also discusses the relative utility of MASW compared with other geophysical techniques and makes mention of promising photogrammetric geomorphic analysis techniques being developed at the University of Canterbury’s Geospatial Research Centre. It draws conclusions and makes recommendations for further work.
2 SHEAR WAVE VELOCITIES: INTERPRETATION AND TECHNIQUES

A wealth of literature relates S-wave velocities to fault-zones, whether for use as a planning tool, geotechnical interpretation or the application of S-wave imaging techniques in tectonically active areas. This chapter reviews this literature as it applies to this study. Limited methodological details are provided here only where necessary to establish a point or where a particular technique is important in the literature but not in this study.

2.1 S-wave velocities and fault-zone planning

Soule (1907) and Branner (1911) noted the relationship between surficial geology and site earthquake response a century ago (Figure 2-1). Soule (1907) wrote:

“The destruction wrought by earthquake in its severe effects was proportional in a way to the nearness of the locality of the fault trace, but varied greatly according to the character of the rock and soil formation throughout the disturbed area. [It] amounted to little or nothing in well built structures resting upon solid rock and, all other things being equal, increased in proportion to the depth and incoherent quality of the foundation soil. [Santa Rosa], built upon a deep alluvial soil, was more severely shaken and suffered greater damage, in proportion to its size, than any other town in the state”

Figure 2-1: Photographs from the San Francisco earthquake of 18 April, 1906 showing the damage associated with low S-wave velocity materials in the subsurface (Soule, 1907).

An understanding of these observations has grown with research. Anderson et al. (1996), for instance, found that the ground motion responsible for such destruction is influenced to the greatest degree by the S-wave velocity properties of the upper 30 m. This influence
is out of keeping with that zone's thickness as a percentage of the source-to-surface distance. They investigated the relative importance of the surficial, intermediate and deeper velocity structure on vertically propagating S-waves and concluded that similar influence on ground motion is attributable to surficial (<30 m) velocities and to deep (>5 km) attenuation properties. They found little influence attributable to the velocities of the intermediate layers even in an attenuating medium. This makes estimation of shallow shear velocity structure an important component of ground motion estimation and site response characterization (Anderson et al., 1996; Louie, 2001) and therefore of comprehensive earthquake preparedness.

Boore (2004a) cautioned that the question of whether site response can be predicted depends on what kind of site response is being predicted and the accuracy required. He showed that there is a large degree of ground motion variability between sites for a given earthquake and, more importantly, between earthquakes for a given site. Nevertheless, shallow shear velocity structure forms the basis of site hazard classification under the United States' NEHRP building provisions (BSSC, 2006). This classification was shown by Boore (2004b) to have been effective in the case of the 2002 Denali Fault Earthquake in Anchorage, Alaska.

2.2 Factors affecting rock mass S-wave velocity

Land zonation using S-wave velocities must be underpinned by sound geology and geotechnical research. The geometry and the material properties of fractures in a medium dictate its seismic parameters (e.g. Boadu, 1997; Leucci and De Giorgi, 2006; Rasolofosaon et al., 2000) so achieving the ultimate objective of this study requires definition of within-zone S-wave velocity variability that is readily correlated with measurable rock properties. An early correlation exercise, although not fault-specific, was undertaken by the United States Geological Survey and reported by Fumal (1978). At each site they drilled a hole to 30 m in a wide range of geologic materials to record the site stratigraphy, physical properties and downhole P and S-wave velocities. The study reported strong correlations of S-wave velocity with hardness and fracture spacing. Fracturing dramatically reduces the elastic moduli, often without significantly changing the porosity so that, in well-cemented lithologies, fracture spacing exerts a greater influence on S-wave velocity than does lithology, hardness or primary porosity (Fumal, 1978; Rasolofosaon et al., 2000). Fumal (1978) found hardness to be a proxy for weathering in bedrock material, in which fracturing dominated S-wave velocities. In soft
rock and alluvium, however, he reported that hardness dominated the S-wave velocities. He further reported less strong correlations between these physical properties and p-wave velocities.

Fumal (1978) typically observed three weathering-related velocity zones in his well logs. With increasing depth, these were a layer of unconsolidated residual material, an underlying zone of moderately to deeply weathered bedrock, and relatively fresh bedrock. His results show a close correlation on almost every well log between these weathering zones and the corresponding interval velocities. More significantly, from the point of view of discriminating lithologies based on S-wave velocities, the thickness of individual weathering zones varied between geological units in a given area. Fumal compared the effect of moderate weathering on S-wave velocity for a variety of lithologies. By plotting the velocity reduction for weathered rock against the fresh rock velocity at the same site over a range of lithologies, he showed that the velocity drop due to weathering is a direct function of the strength of the fresh rock. Weak rocks will exhibit a greater absolute velocity drop for an equal degree of weathering than will equivalently weathered strong rocks. This suggests the potential to discriminate fault-zones juxtaposing different soft lithologies based on changing vertical velocity gradients.

![Graph showing reduction in S-wave velocity attributable to moderate weathering](image)

*Figure 2-2: Reduction in S-wave velocity attributable to moderate weathering, graphed as a function of the fresh rock S-wave velocity (after Fumal, 1978).*

Geotechnical interpretation of S-wave behaviour in a geological medium must be guided by an understanding of the physical relationships between S-wave propagation and rock
mass character. A waveform propagating through a fractured medium is delayed and the high frequencies are filtered, so that the peak frequency in the received signal is shifted toward the lower frequencies. This has been observed in theoretical models (Boadu, 1997), and in laboratory models and the field (Leucci and De Giorgi, 2006). As a wave propagates, an apparent attenuation accumulates because reflections from individual fractures interfere with the propagating wave (Boadu and Long, 1996). This seismic response to fracturing provides a means of inferring fracture density from variations in seismic velocity in and otherwise homogenous medium Boadu and Long (1996) incorporate linear fracture density into their equations for estimating velocity and attenuation in a fractured medium.

2.3 Potential correlative rock mass classifications

Many parameters have been used to attempt to quantify the condition and mechanical state of a fractured rock mass. Volumetric crack density has been correlated with seismic velocities (Crampin et al., 1980; Leary and Henyey, 1985) but is not easily measurable in field conditions. In recognition of this, Boadu (1997), modelling the effect of fractures on seismic waves, correlated S-wave velocity strongly with linear fracture density (fractures per unit length) and rock quality designation (RQD - % of core with fracture spacing >10 cm) among other parameters (Figure 2-3). His correlation, although significant for this study, takes no account of the effects of weathering.

![Figure 2-3: Variations of S-wave velocity with fractured rock mass parameters for the model of Boadu (1997). Both parameters exhibit a strong correlation with S-wave velocity.](image)

Rock masses are typically characterised based on intact rock strength and the condition of the rock mass discontinuities (Gokceoglu and Aksoy, 2000; Hoek et al., 1998). Mineralogy of the intact material is only accounted for where it affects intact rock
strength or, as in the work of Gokceoglu and Aksoy (2000), renders the rock mass vulnerable to the damaging effects of water. This characterisation approach implicitly assumes that discrete discontinuities separate intact rock, a fair assumption only if core can be recovered from the rock-mass in question to determine the intact material properties of the rock mass itself (rather than the parent material), and if the properties of the fracture fill do not dominate the rock mass.

The commonly-used Hoek-Brown failure criterion was developed by Hoek and Brown (1980) to estimate rock mass properties from intact rock properties and discontinuity characteristics. It was originally dependent on a simplified rock mass rating (RMR) system. Quantitative characterisation and classification of cataclastic rocks using RQD-based systems such as RMR is troublesome due to difficulties in acquiring undisturbed samples for laboratory test procedures such as unconfined compressive strength (UCS). These difficulties persist during sample preparation and testing leading to a very low experiment success rate (e.g. Habimana et al., 2002). When specimens are successfully tested they are typically not at all representative of the rock mass.

The difficulty obtaining meaningful RQD values in weak rock masses led to the abandonment of RMR in favour of the specially developed and introduced Geological Strength Index (GSI) (Hoek et al., 1992). GSI now forms the basis of parameter selection for the Hoek-Brown failure criterion. It is a classification system for qualitatively estimating the reduction in rock mass strength due to geological conditions. Index values are assigned based on visual assessment of lithology, the degree of interlocking of rock mass components and the quality of the discontinuities indicated by surface roughness and alteration. It therefore takes account of the two major factors affecting S-wave velocity, fracturing and weathering. The original GSI and Hoek-Brown failure criterion system inherently assumed rock mass isotropy and was thus inappropriate where a preferred fabric orientation such as schistosity dominated the mass. Regular encounters with anisotropy have led to the extension of the GSI to allow assigning of values in anisotropic masses where the difference between the strength of the rock and that of the discontinuities is small, such as closely spaced weak schistosity or shear planes (Hoek et al., 1998). GSI has also since been further extended to flysch (Marinos and Hoek, 2001) and modified for cataclastic rocks (Habimana et al., 2002), but is still inappropriate where a single set of discontinuities dominate an otherwise sound rock mass.
In addition to modifying the GSI classification system, Habimana et al. (2002) improved the reliability of the Hoek-Brown failure criterion for cataclastic rocks. Hoek and Brown (1997) presented formulae relating the Hoek-Brown failure criterion constants $m_b$, $s$ and $a$ to the estimated GSI values. Habimana et al. (2002), however, found that the failure criterion for cataclastic rocks varies between pure Hoek-Brown for intact rock as the upper bound (un- or lightly tectonised) and Mohr-Coulomb for soil as the lower bound (extremely tectonised). They related these tectonisation degrees to the GSI (Figure 2-4) and substantially modified the formulae of Hoek et al. (1998) to account for this transitional behaviour.

![Geological Strength Index Diagram](image)

**Figure 2-4: Geological strength index modified after Habimana et al. (Habimana et al., 2002)**

Based on the observations of Fumal (1978), GSI shows promise as a correlative classification against which field survey S-wave velocities may be interpreted. Such a correlation would feed directly into the geotechnical design via the modified Hoek Brown failure criterion of Habimana et al. (2002). However, GSI does not lend itself to laboratory characterization and, as the objective of this study is to relate the laboratory experiments to field data, the most transferable correlative may be linear fracture density as observed by Boadu (1997).
2.3.1 Photogrammetric options

A significant amount of the time invested in site investigations is spent collecting scanline and other geotechnical data at outcrop, and carrying out laboratory work in an attempt to effectively classify the rock mass. This study is no exception and the result is costly in terms of time and money, and almost certainly subject to the skill and experience of the geologist. An early objective of this study was to develop a cheap, rapid and objective photogrammetric technique for rock mass classification. Although such a technique was never developed, considerable time was invested prior to its abandonment, during which several existing packages were investigated for inspiration. One of these packages in particular stood out as a potential tool for fault-zone classification.

Fragalys is a software package developed by the Central Mining Research Institute in India for explosive excavation optimisation. It is based on a fundamental assumption that rock mass fracturing displays a fractal distribution. A fractal is defined by Hobbs (1993) as 'a shape made of parts similar to the whole in some way'. Fractured rock masses have been shown to display fractal geometries (Chelidze and Gueguen, 1990; Xie, 1993; Xie and Chen, 1988) and rock mass characterization on the basis of this fractal geometry is well established in the literature (Babaie et al., 1995; Bagde et al., 2002; Dor et al., 2006; Gao et al., 1999; Gao et al., 2004; Ghosh and Daemen, 1993; Hamdi and du Mouza, 2005; Lu et al., 2005; Tu et al., 2005). Boadu's (1997) computational analysis, along with his subsequent modelling work (Boadu, 2000) and laboratory work by Leucci and De Giorgi (2006), suggests that simple and valid empirical relationships can easily be developed correlating seismic velocities with fractal rock mass parameters for any given lithology. Leucci and De Giorgi (2006) tested their derived relationships in the field and found good agreement between geological, geomorphological and seismic results.

The Fragalyst system consists of capturing video photographs of a scree pile and its open face using a calibrated video camera. The footage is downloaded and enhanced, calibrated, processed and analyzed to determine the area, size, shape, and sphericity of the fragments on the basis of greyscale difference. The fractal indices are determined from the shape factor (ratio of area to perimeter squared) of fragments for both the scree pile and the in situ rock blocks. Bagde et al. (2002) tested the system and concluded that whilst most in-situ rock produced fractal indices lying in the RMR range of 30-40 and beyond, the ratio of the scree fragment dimension to the in situ dimension decreased with increasing RMR. This reflects the rapid disaggregation of an apparently strong rock often
seen with tectonized rock masses. The strong correlation of the dimension ratio with an important rock mass parameter suggests that a single photogrammetrically-obtainable fractal-based number should be sufficient to classify weak rock masses.

2.4 Multi-scale S-wave velocity measurement

Characterization of the within-zone spatial variability in lithology, fracturing, fabric development and cataclasis is necessary to understand past and likely future fault behaviour. These are characteristics that vary at multiple scales and also have been shown to correlate with wave propagation phenomena such as low seismic velocities (Gettemy et al., 2004; Zinszner et al., 2002).

Unfortunately the correlation is scale dependent. Material heterogeneity across a fault-zone may induce apparent dispersion (frequency dependent velocity) due to the significant lithological, fluid and deformation variations that occur at scale lengths from <5mm to >5 m (Gettemy et al., 2004; Mukerji et al., 1995). The dispersion is controlled by the ratio of the wavelength of investigation (\(h\)) to the scale length of heterogeneity (\(a\)). In general the velocities increase with decreasing values of \(h/a\) so that ultrasonic velocities measured at core scale in the laboratory (\(h\) is small) are higher than those measured at crosshole or seismic scale and require corrections when changing scales (Mukerji et al., 1995).

This dispersion can be hard to predict without recourse to numerical modelling (eg Gettemy et al., 2004) and this may affect the reliability of interpretations and predictions made based on velocities obtained at seismic survey scale. Nevertheless, this study is conducted at three scales, from ultrasonic laboratory testing to sonic crosshole and seismic surveys averaging velocities over metres to tens of metres and simple scaling relationships will be sought. Each of the methods used is a transmission technique, and therefore suitable for investigating fractured media (Boadu, 1997). The techniques will be reviewed in scale order from smallest to largest.

2.4.1 Laboratory scale: Ultrasonic velocity measurement

Published data on the ultrasonic S-wave velocities of this study’s target lithologies will be introduced site by site. However, the objectives of this study demand more than simple laboratory S-wave velocity determination. Fracturing clearly dominates S-wave velocities in competent bedrock (Fumal, 1978), and in fact fracturing dominates other causes of
seismic anisotropy by <100% at confining pressures lower than a few tens of megapascals (Rasolofosaon et al., 2000). What is required for this study is a method of quantifying the effect of fracturing and/or weathering, and the commensurate reduction in rock mass strength, on the characteristic S-wave velocity of a lithology.

Previous studies have approached this question using different techniques but with the same general pattern of results. Shakeel and King (1998) and King et al. (1997) investigated seismic wave propagation in sandstone with a system of aligned cracks, which they introduced during measurements using a polyaxial stress loading system. They increased \( \sigma_1 \) and \( \sigma_2 \) to near failure and demonstrated a strong correlation between permeability in the plane of cracks (related to aperture) and S-wave velocities propagating, not only in the direction normal to the plane of the cracks, but also in the plane of the cracks and polarized normal to the cracks. In a similar experiment Wulff et al. (2000) investigated the mechanisms involved in the attenuation of seismic waves so as to relate attenuation to fracturing. As with Shakeel and King (1998) Wulff et al. measured ultrasonic wave propagation whilst increasing fracturing. They observed similar results and report relationships demonstrating that microcracking due to uniaxial unconfined compression strongly influences velocity and attenuation of ultrasonic waves in dry sandstones.

Both of these studies quantified relationships between cracks and acoustic velocities. Although these types of experiments seem ideal, difficulties arise in trying to reproduce them using simple and limited rock laboratory equipment. Particular difficulties arise in quantifying the incremental intensity of fracturing, hence Wulff et al. (2000) used axial strain as a proxy for fracture intensity. A different approach is taken by Nakagawa et al. (2002), who modelled elastic wave propagation parallel to an infinite number of plane parallel fractures with equal fracture spacing and fracture stiffnesses. Their results are of limited relevance to this study but, in validating their model data, they used a steel analogue with artificial ‘fractures’ (Figure 2-5). This concept seems promising for simple laboratory work, with the substitution for steel of parent material ‘fractured’ normal to the S-wave propagation direction.

2.4.2 Outcrop scale: crosshole velocity measurements

As has occurred in laboratory studies, correlations based on crosshole transmission of seismic energy have been demonstrated between decreasing S-wave velocities and
increases in fracture frequency or density (e.g. Worthington, 1984). Zhu and Toksooz (2003) used modelling to develop a novel seismo-electric approach to investigate fractures between boreholes. Their technique, unfortunately beyond the scope of this study, records arrival times for an electromagnetic wave induced by a guide wave at a fracture between an acoustic source and receiver boreholes.

![Diagram of experimental setup](image)

Figure 2-5: Nakagawa et al.'s (2002) experiment was conducted using ultrasonic transducers with waves propagating parallel to fractures in simulated fractured material (steel plates). The lead foil provided acoustic coupling.

Propagation through a fracture causes substantially greater amplitude reduction in a reflected wave than a transmitted wave, so transmission techniques are best for imaging fractured media (Boadu, 1997). Crosshole geometry also allows study of raypaths normal to the strike of fractures. King et al. (1986) carried out horizontal crosshole measurements in blast-damaged columnar basalt. Their study however, whilst validating horizontal crossholes for rock mass investigation, used drilling and seismic equipment appropriate to a mining budget and beyond the reach of this study. They did, however, demonstrate a strong relationship in 3 dimensions between S-wave velocity and the intensity of fracturing due to blast damage at the face.

2.4.3 Survey scale: Multichannel analysis of surface waves

Detailed discussion of the method and geophysical principles of Multichannel Analysis of Surface waves (MASW) are reserved for coverage in Chapter 3 and Appendix A respectively. This section will briefly introduce those aspects as necessary for clarity but will mainly be concerned with reviewing the several advantages and applications of the MASW technique, as well as validation studies undertaken and previous uses of MASW in active tectonic environments.

S-wave velocity determination by surface wave inversion has been applied at widely varying scales, from regional full-crustal scale (Surface wave tomography e.g. Behr et al.,
2007), through multichannel near-surface surveys such as Refraction Microtremor (Louie, 2001) and MASW (Park et al., 1999b) to Spectral Analysis of Surface Waves (SASW) soundings (Stokoe et al., 1994), MASW's small-scale predecessor. Surface wave inversion assumes that S-wave velocity dominates layer thickness, P-wave velocity and density in its influence on Rayleigh wave (the direct surface parallel transmission of S-waves) phase velocity and dispersion characteristics (e.g. Xia et al., 1999).

In a seismic survey, more than two thirds of the seismic energy generated is partitioned into ground-roll-generating Rayleigh waves. Propagation through a fracture causes substantially greater amplitude reduction in a reflected wave than a transmitted wave, so transmission techniques are best for imaging fractured media (Boadu, 1997). Transmitted Rayleigh wave noise is exploited by the MASW technique, developed by Park et al. (1999b) at Kansas Geological Survey, to yield a 2D S-wave velocity profile. The procedure consists of acquiring multichannel records of surface wave events, extracting the fundamental-mode dispersion curves from each record and inverting these curves to obtain 1-D (depth) Vs profiles for each curve. The interpolation of the inverted profiles by contouring software then produces a 2D Vs profile for the length of the survey. The use of multichannel records permits identification and removal of noise prior to dispersion curve extraction. The nature of surface wave surveys favours the use of land streamers because optimal receiver coupling is generally obtainable by pressure contact with the surface rather than labour-intensive planting and recovery (Miller et al., 1999b). The receiver spread is necessarily short to accommodate lateral velocity variation (Park et al., 1999b) and thus easily handled by a minimal team. Geophones are vertically oriented and of the type used for near surface seismic reflection and refraction surveys. This versatile and towable multichannel setup enhances both speed and redundancy in data acquisition, whilst reducing labour requirements.

A fundamental assumption of MASW acquisition and processing is that the subsurface conforms to a laterally homogeneous layered earth model. This assumption can be validated by examination of trace to trace linear coherency, and is best satisfied if receiver spread is kept as short as possible (Park et al., 1999b). This can create issues with dispersion curve extraction, which is optimized by having a long receiver spread to separate phase velocities of the multiple modes (see Appendix A). However, given a suitably short receiver spread and successful dispersion curve extraction, the inherent sensitivity of surface wave velocities to lateral seismic velocity changes renders them
particularly suitable for detection of structural features that are characterized by lateral velocity changes in the upper 50 m such as faults (Ivanov et al., 2006).

The MASW method has recently become increasingly popular for a range of applications including the detection of subsurface features such as bedrock topography. Shallow seismic characterization of rock mass fracturing and bedrock topography is attractive in applications such as waste isolation or water containment in fractured rock. Bedrock topography and character has a major influence on subsurface hydrology and may exert significant control over the movement or otherwise of contaminants in the subsurface (Miller et al., 1999b). Miller et al. observed large lateral velocity-gradients within their inverted Vs profile that were consistent with drill-confirmed bedrock topography. They also observed localized lateral velocity reductions consistent with low velocity fracture zones and erosional channels. Miller et al.'s study was carried out in a manufacturing area with significant cultural and electrical noise but MASW proved largely insensitive to this. Ivanov et al. (2006) discriminated two formations of dipping weathered and intact bedrock, the overlying regolith and a concealed fault-zone. Their success at delineating the dip was due to compositionally preferential weathering of the bedrock surface, whilst they attributed their overall success to a half-day of pre-survey on-site calibration to optimize the source offset and shot spacing. The importance of field parameter selection is discussed in Appendix A.

The primary objective of Ivanov et al. (2006) was to use MASW to investigate a fault-zone. They imaged an 80 m wide zone with seismic velocities ranging from 400-800 m/s, which coincided with a previously mapped thrust fault. They attributed the reduction in Vs of the fault-zone to a reduction in shear modulus related to fracturing and in-situ weathering. This conclusion is supported by drill logs in the fault-zone, which record blocks of indurated bedrock floating in clay gouge at depths of up to 60 m. A secondary but very important capability demonstrated by Ivanov et al. (2006) is discrimination between primary lithological variation and faulting. They were able to distinguish the two juxtaposed formations both from each other and from the intervening fault based on weathering velocities (Figure 2-6). This supports the conclusions of Fumal (1978) (Section 2.2).

Besides bedrock and faults, other features detected include buried pits and trenches (Miller et al., 2000), sinkholes (Miller et al., 1999a) and cavities (Nasser-Moghaddam et al., 2005; Xu and Butt, 2006). Liquefaction hazards are correlated with S-wave velocity in
the laboratory (Andrus and Stokoe, 2000; Yunmin et al., 2005), and Lin et al. (2004) have favourably assessed the utility of MASW for estimating the extent of liquefaction hazards. Geotechnical applications of MASW are particularly promising. Shallow S-wave velocity structure is an indication of stiffness and is used when estimating site response to shaking (section 2.1). By facilitating construction of an S-wave velocity profile MASW can provide critical information for geotechnical site characterization. Specific applications have included non-destructive testing of asphalt or concrete slabs (Rhazi et al., 2002; Ryden and Park, 2005), ground stiffness assessment (Joh et al., 2006; Ryden and Park, 2005; Tomeh et al., 2006) and marine sediment stiffness assessment (Park et al., 2005c). Crice (2005) suggests that the popularity of surface wave methods in general, including MASW, will continue to grow due to their ease of acquisition, processing and interpretation.

![Figure 2-6: MASW 2D Vs profile across an unnamed New Jersey fault-zone. The fault-zone and the dipping and weathered bedrock of the fault bounded formations are clearly imaged to 20m depth. High velocities can be seen south of the fault in the interbedded sandstones and shales of the Stockton Formation. This is in contrast with the mid-range velocities of the weathered massive and laminated mudstones of the Lockatong Formation, which outcrops to the north of the fault (Ivanov et al., 2006).](image)

Despite this popularity and the successful fault study by Ivanov et al. (2006 9) little else has been written on the application of MASW to the study of faults. Only three other examples appear in a review of the literature as at June 2008. Seshunarayana et al. (2008) used MASW to map shear zones in fractured granite basement. Of the remaining two, one is a preliminary study (Parrales et al., 2003) on which nothing further has been published to date and the other (Karastathis et al., 2007) made only very limited use of MASW, using it as a tool to provide velocities for other shallow seismic techniques. Neither provide details or discussion of the results of the MASW surveys, leaving Ivanov et al. (2006) as the only comprehensive published literature on application of MASW to active faults.
Several studies have been published independently validating the S-wave velocities produced by inversion of surface wave data using the MASW and other surface wave techniques (Joh et al., 2006; Stephenson et al., 2005; Xia et al., 2002; Zhang and Chan, 2003). Of these studies, Stephenson et al.'s (2005) blind study achieved matches of S-wave velocity averaged to 30 m (V_{30}) within 15% on all four boreholes and within 3% on two of the four. Xia et al. (2002) similarly achieved less than 15% variance from borehole data on their 8 well study, although only one well was matched blind. That well achieved a calculated overall difference of only 9%. Stephenson et al.'s (2005) study also included a comparative evaluation with the refraction microtremor (ReMi) method of Louie (2001). They concluded that neither method was consistently better and suggested that both MASW and ReMi were appropriate tools for hazard assessment. These positive results, however, require well constrained fundamental mode dispersion curves and deteriorate rapidly with poor data and/or processing (Zhang and Chan, 2003). Such issues surrounding data acquisition and processing and are covered further in Chapter 3 and Appendix A. A proposed quality index for classifying dispersion curves is appended at Appendix B.
3 STUDY METHODS

The stated objective of this study can be broken down into three major components. The first of these is to prove and optimize the utility of the MASW technique to locate and define the extent of fault-zones. The second component is to correlate the MASW data with standard rock mechanics measures of fracture intensity. The third and final component is to correlate MASW and rock mechanics data with the distribution and character of ground deformation as a predictive tool. Following a brief outline of mapping methods, the structure of the remainder of this chapter reflects the order of those objectives. A variety of techniques were used repeatedly during this study, each selected to best achieve the study objectives. The use of these techniques is reviewed and the methods employed are outlined in the necessary detail here. Methodological details are then omitted in subsequent chapters.

3.1 Mapping

The two selected calibration areas were mapped in detail to establish the geological and geomorphic relationships. The general site geomorphology, emphasising tectonically significant features, was mapped in the field using a combination of the 1963 Aerial Mapping Limited air photos and the 1995 orthophotos from Land Information New Zealand. The older photographs were used because the study areas were significantly less vegetated at that time. A2 sized maps were prepared at scales of 1:5000. No field work was carried out for the blind test at the Taieri Ridge site.

3.2 Survey scale seismic investigations – MASW

The primary objective, outlined in section 1.2, was served by carrying out extensive MASW surveying. The MASW procedure is summarized in Figure 3-1 and its theory expounded in Park et al. (1999b), Xia et al. (1999) and Appendix A. Stages in the technique include data acquisition, pre-processing, dispersion curve extraction, and surface wave inversion and contouring.

3.2.1 Surface-wave Data Acquisition

3.2.1.1 Acquisition parameters

Data-acquisition parameters were chosen to optimize recording of ground roll signals. The choice of near offset (x1 – Figure 3-2) is normally critical, as it is best to record both
high frequency and low frequency planar (as opposed to cylindrical – Appendix A) surface waves. This allows the best combination of near surface resolution (high frequencies) and penetration (low frequencies). Another critical parameter in this regard is geophone spacing. Reducing geophone spacing can increase the amount of high frequency energy that is collected at far offsets. Surveys in this study were acquired initially with spacings of 1 m and later surveys with 0.5 m spacing. The relative merits of the two geometries will be discussed in Chapter 8.

![Diagram](image)

**Figure 3-1: Basic steps in the MASW method.**

![Diagram](image)

**Figure 3-2. MASW survey geometry using a land streamer. Key acquisition parameters are illustrated and selection of these is discussed in the text.**

Some sites are amenable to optimization of acquisition parameters by examining a swept frequency record from a full wavefield profile. In this study setting, however, this was found to be of limited use and collection of data for 48 channel walkaways proved more effective and versatile. Where this was possible, greater freedom could be taken in
selection of the near offset. Literature is available on optimal parameters for soil sites (Park et al., 2005a) but is probably of limited use for bedrock floored sites. This study’s conclusions will include recommended near offsets for the lithologies encountered.

3.2.2 Equipment and settings

Multichannel shot records were collected in a CDP roll-along type survey using an end-on string of 24 8Hz geophones mounted on steel ground-contact sleds connected together in a land-streamer (Figure 3-3). As a minimum, the start and finish position of the array centre were recorded by a theodolite survey relative to the local site survey pegs. Array centres were surveyed where topographic correction was needed. The array was towed by a 4wd vehicle, which was switched off between shots. Additionally, the vehicle was rolled back slightly at each position to avoid transmission of noise from the vehicle to the geophone string along the tow cable. An 8 kg sledgehammer was used to provide an impulse source and an accelerometre on the hammer handle triggered data collection. Shots were recorded using the first 24 channels of a Geometrics Strataveiw 48 channel seismograph. They were then transferred to a laptop computer for processing using Surfseis software from Kansas Geological Survey. Recording time and sampling intervals varied between surveys. No filters were used and a variable number of shots were vertically stacked at each location. A table of acquisition parameters is included for each site in the relevant chapter.

![Image](image.jpg)

*Figure 3-3. Land streamer carrying 24 channel 4Hz geophone string. The streamer is towed by the vehicle in the photograph, which is switched off between moves.*

3.2.3 Data Pre-processing

The data were preprocessed within the Surfseis setup and display utilities as described in the manual (KGS, 2003). Steps included
- Converting the data from SEG-2 to KGS format. This was done as a batch process for standard surveys. In addition, files to be used as part of a walkaway record were individually converted.

- Walkaway construction where necessary. Both fixed receiver (FRW) and fixed source walkaways (FSW) (Vincent et al., 2006) were used at various locations. Walkaway pairs were selected as shown in Figure 3-4.

![Fixed Receiver Walkaway and Fixed Source Walkaway Diagram]

**Figure 3-4. MASW survey configuration for walkaway construction.**

- Application of field geometry to the converted files, based on field survey observer records of source and receiver locations. FRWs were assigned a field geometry with the source 'relocated' such that the '48 channel' array centre was located at the true array centre.

The following three steps are essentially preprocessing but were carried out record by record during the data analysis process, rather than batch processed prior to analysis, because of the variability between records.

- Swept frequency analysis to determine frequency ranges over which velocities are coherent and to determine velocities of body wave and higher mode contamination.

- F-K filtering to remove backscattered and/or body wave contamination.

- Muting of the first arrival to improve surface wave signal to noise ratio.

If large sections of a survey are expected to be internally laterally homogenous, steps 4, 5 and 6 can be batch processed prior to analysis.
3.2.4 Dispersion curve extraction

This stage of MASW includes the following steps.

1. Overtone analysis to determine the frequency and velocity ranges of surface waves and to examine the dispersion image for the presence of modes.

2. Where higher modes were observed existing alongside fundamental modes, modal separation before dispersion curve extraction using a combination of
   - F-K filtering if the modal velocity fields are not critically overlapping or
   - Offset-time domain muting of higher modes (Ivanov et al., 2005). Muting higher modes also mutes low frequency fundamental mode energy so this method requires separate extraction of low and high frequency dispersion and combination of the curves. Fundamental and higher modes propagating at slightly different velocities are more easily visually distinguished at far offsets, so this technique is more easily carried out on records with a greater number of channels (Figure 3-5). This is the main reason for the use of walkaways in this study.

![Figure 3-5](image)

Figure 3-5. Relationship between ease of differentiation of fundamental and higher mode rayleigh waves and the length of the geophone string. The longer the array, the easier to visually identify higher mode and body waves. A short 24 channel array (1) shows very poorly the diverging velocities of fundamental and higher modes compared with a 120 channel array (3) at a similar near offset. A 48 channel array (2) collects significantly more fundamental mode energy, facilitating differentiation and muting.

3. Dispersion curve extraction from the overtone image. Fundamental mode surface wave velocities were picked by user guided automatic picking at 0.5 Hz point to point intervals over the full fundamental mode range present in the record, followed by manual editing of the picked curve.

Removal of higher modes was not always, or even often, possible, which typically limited the highest frequencies for which a dispersion curve could be picked and thus the near surface resolution (see Appendix A).
3.2.5 *Surface wave inversion and contouring*

Dispersion curves were batch inverted in Surfiseis to generate a map of 8-wave velocity with depth (Xia et al., 1999). The profile consists of interpolated soundings, each located in the middle of the receiver spread. Default initial models defined by Xia *et al.*'s algorithm were used as they generally converged to geologically acceptable models with an acceptable dispersion curve fit, typically 5-15%. The number of layers chosen was 10 for all models. The thickness of each layer varies based on the depth of investigation and the recorded bandwidth (see Appendix A).

Where significant topography was encountered during a survey, the velocity text file was imported into Surfer8 contouring software to allow topography to be incorporated in the contoured profile. This was done by first altering the text file in Excel to correct the depth to each layer relative to the surveyed elevation of the array centre. The Excel file was then gridded and contoured in Surfer8. Surfer produces rectangular maps so an extra 0 m/s layer was added at the surface elevation to highlight the topography and to prevent extrapolation of surface wave velocities into the air. Overall, this technique has the effect of shearing the 1D soundings but should produce a much closer fit with the real earth model than no correction at all (Figure 3-6).

![Figure 3-6: MASW profile map with velocity contrast (A) with no topographic correction and (B) with soundings sheared in Surfer8. Profile B compares favourably with the earth model (C) whilst profile A produces structurally misinterpretable artefacts.](image)

30
3.3 Outcrop scale geotechnical and seismic investigations

3.3.1 Rock mass characterization

Standard scanline surveys, mainly consisting of fracture counts, were done along scanlines at all outcrops, whether hard or soft rock. Schmidt hammer testing was carried out at the Boby’s Stream site only. In addition, a novel nail penetration technique was trialled at Boby’s Stream for in-situ determination of rock strength for soft rocks. Description of the nail penetration test is reserved for the laboratory methodologies (Section 3.4.1).

3.3.2 Crosshole velocity measurements

Horizontal holes, 32 mm in diameter and ranging up to 0.7 m deep, were drilled in outcrops for the purposes of making crosshole velocity measurements. The holes were located so as to sample as closely as possible tectonic or stratigraphic units that were internally geotechnically homogenous. Each hole was drilled into the face and sleeved using 1 m lengths of 32 mm electrical conduit. Where possible the holes were drilled parallel with each other, but where the face had significant topography this was often not possible. In these cases the holes were drilled normal to the face and the orientation of the conduit sleeve was recorded in terms of trend and plunge. The location of the protruding ends of the conduits, was recorded relative to each other and to the site using EDM equipment, allowing the drill holes to be incorporated into a 3D model of the site and the source-receiver distance to be calculated.

Beginning at one end of the outcrop, a specially developed slide hammer (Figure 3-7) was inserted in one hole to provide a seismic source whilst a geophone was inserted in the adjacent hole. The slide hammer was set up as both source and trigger. For each hole-pair the source and receiver hole numbers, the depth of penetration of the slide hammer and the depth of the geophone were recorded. The slide hammer was operated both into and out of the face, twice in each direction, and the signals recorded using the equipment illustrated in Figure 3-8. Each measurement was recorded as a text file. These files were imported into a Microsoft Excel spreadsheet and signal voltage was graphed as a function of time. S-wave arrival times were picked visually and crosshole velocities computed using the calculated source-receiver distance.
3.4 Laboratory scale geotechnical and seismic investigations

Field classification of tectonized rock masses is difficult and may be somewhat subjective. In order to soundly underpin any apparent correlations, a laboratory programme was undertaken to explore calibration of S-wave velocities to measurable rock properties. As no standard method is available for such a programme, development of the programme is important, both as a stand alone methodology and as a tool to ensure the success and validity of this study as a whole.
This study produced seismic data at core scale in the lab, drill-hole scale at the outcrop and survey line scale on the terraces and/or roads. A key objective is to undertake laboratory calibration and upscale the data to crosshole and survey scale. The upscaling process needs to account for any lack of weathering in lab samples and correct for scale-dependent apparent dispersion due to the different input signal wavelengths of core-scale and survey-scale work. Simple scaling techniques were developed site by site and are discussed in the relevant chapters.

3.4.1 Laboratory investigations

Laboratory experiments were carried out to investigate the effect of fracturing on S-wave velocity for relatively intact specimens and to relate the S-wave velocity to the physical properties of tectonized specimens. The ‘intact’ specimens were simply the largest blocks that could be found on site representative of each lithology. Where locally sourced large blocks from which core could be recovered were unavailable, the best alternative was used. Examples include use of a river boulder for Torlesse greywacke and use of a quarry block taken from further off the fault plane for Amuri and Weka Pass Limestone. At least three 50 mm diameter cylindrical core specimens were recovered from each specimen, preferably sampling in three mutually perpendicular directions, and the ends were ground square to the core axis. The specimens were then oven dried for at least 24 hours. The axial S-wave velocities of the intact cores were measured using the GCTS Testing Systems ULT-100 ultrasonic velocity measurement system. The full waveforms are recorded by CATS ultrasonics software and velocity is determined by pulse travel time. Coupling was improved using either superglue or petroleum jelly and the measurements were taken at increasing uniaxial stress. The three groups of measurements for each lithology were plotted together to look for evidence of S-wave splitting.

The individual cores were then incrementally ‘fractured’ (sawn) at regular intervals normal to the core axis and S-wave velocity was measured after introduction of each fracture with correction for length lost from the saw cut thickness (Figure 3-9). These velocities were all measured at a nominal axial stress determined from the initial testing to be the minimum required for coupling to be achieved. The thickness of the sawn disc was recorded to allow calculation of a running tally of fracture spacing as well as fracture density. Fractures were introduced until S-waves were completely attenuated.
Figure 3-9: Example of fractured greywacke core. Fractures were introduced from right to left. The core is numbered 1/9/1 using the system adopted for this study (sample 1/core 9/orientation 1).

Where possible, physical properties were determined using core, for which volume is simply calculated. In many instances however, such as for Torlesse Greywacke, even the best site samples were too internally fractured to yield any core at all. In this case irregular blocks, whose size was limited by the disaggregation of the sample along preexisting fractures, were sawn and their volume determined by the immersion method. Volume by this method is given by:

\[ V = \frac{(M_{sd} - M_s)}{\rho_w} \]

Equation 3-1

Where \( V \) is the volume in m\(^3\), \( M_{sd} \) is the surface dried saturated mass and \( M_s \) is the submerged saturated mass, both in kg, and \( \rho_w \) is the density of water in kg/m\(^3\). Density and porosity were determined for all specimens, whether core or irregular.

Uniaxial Compressive Strength (UCS) was determined for any cores not required for the fracturing experiment following the ISRM guidelines (ISRM, 1979). The ISRM requirement for 2:1 length to diameter ratio was prohibitive for the purpose of characterizing weak and/or tectonized rocks, which typically broke into short lengths during core drilling, so short cores were often used. Short cores are stronger than long cores for a given diameter so the measured UCS was corrected using the relationship proposed by Turk and Dearman (1986), which averages three very similar relationships (ASTM, 1980; Protodyakonov, 1969; Slaevin, 1974). To avoid unacceptable scatter of results, no core was tested with a length to diameter ratio > 1. This approach has been shown to keep all results within ±5% of values obtained using cores with L:D of 2:1 (Turk and Dearman, 1986).

Point load strength (Ghosh and Srivastava, 1991; ISRM, 1985) was determined for all irregular blocks. Where possible, S-wave velocity was also determined for these irregular
block samples, although it was noted that factors such as non-parallel faces and incomplete fit of the transducer on the contact face would probably lead to systematic errors in the data.

A late change to the laboratory and field program was the inclusion of hardness determination because of problems characterizing soft rock on the basis of fracturing. Hardness was measured for the Boby’s Stream test site using both a Schmidt hammer and the nail penetration technique mentioned earlier. The latter was trialled with the objective of developing a field test, more reliable than the Schmidt hammer, that would be useful for low strength lithologies to bridge the range from soft to harder rock, and that could be readily correlated to a standard estimate of rock strength such as UCS. Indentation testing is already used to good effect to determine hardness in metals, alloys and even rocks (Szwedzicki, 1998) but the published standard testing procedure includes application of a standard indenter, specification of a standardized loading rate, criteria for test termination, specification for the properties of the cementing agent and application of continuous data logging (Brown, 1981).

These are difficult conditions to comply with whilst testing in-situ rock masses in inaccessible locations. Nail penetration testing was carried out in both laboratory and field using a Ramset nail gun, firing 220 green, .22 calibre x 0.825 low power cartridges and 75 mm nails. The gun was loaded and placed against the test surface to release the safety mechanism (Figure 3-10). The gun was fired and the nail’s protrusion in mm was recorded and converted to a penetration figure.

### 3.5 Geomorphological investigations

Quantitative analysis of geomorphology is widely used to derive structural and tectonic data. Downcutting streams in an uplifting catchment reflect an interaction between variables that include climate, sediment load, tributary contributions, lithological variations and tectonics. River gradient, pattern, valley morphology and downcutting history are particularly sensitive to faulting and deformation within the floodplain (Local examples include Bull, 1996; Campbell et al., 2003; Litchfield et al., 2003; Nicol and Campbell, 2001). It follows that interpretation of the tectonic history of a landscape must be underpinned by an understanding of how tectonic perturbations act on a fluvial system, and how the tectonic signal can be separated from other factors. For complete coverage of
most tectonic geomorphologic concepts the interested reader is referred to Bull (2007),

Figure 3-10: Ramset gun being fired at a Walkari Siltstone outcrop, Waipara River.

Detailed topographic surveys were carried out using a Trimble Geodimeter 5600 semi-
robotic total station. Base stations were surveyed in as necessary to provide line-of-sight
coverage and the positions of at least two stations per site were fixed using differential
GPS. These peg locations, together with previous data where possible, were integrated
with the survey data in Trimble’s Terramodel 10.41 survey software. South Island contour
shapefiles were clipped in ArcMap for the various study areas and the clipped files
imported into Terramodel. The completed dataset was exported to AutoCAD in the New
Zealand Map Grid coordinate system. Where survey data were sufficiently dense,
localised small interval contour maps were generated in Surfer8 and exported to
AutoCAD for incorporation in the site model.

As this study is primarily intended to prove the utility of MASW, 3D topographic profiles
were created that showed lateral geomorphological variation in spatial context and could
be easily correlated along-strike with 2D MASW profiles. Although 2D longitudinal
profiles of river channels are presented, their spatial distortion made them of limited use
in this study. Surveyed features include river thalwegs and topographic profiles. The latter
included valley cross sections, fault scarp and terrace edge profiles, terrace surfaces and
straths, although not for all sites. As far as possible, profiles were surveyed approximately normal to the strike of the local structures. This strategy was to avoid incorporating complex non-tectonic gradients. In AutoCAD, the integrated survey data were georeferenced to the site orthophoto and used to render accurate maps of the geomorphic features of interest. A new coordinate system (the structural UCS) was defined in AutoCAD for each site, which placed the fault structure contours as close as possible to parallel with the Y axis. This allowed the surveys to be viewed orthographically from the top, front and side relative to the structure.

3.6 Correlation methods

Site layout factors such as topography, ground conditions and access generally constrained the location and orientation of the MASW survey lines. The lines thus varied between strike normal and variably oblique to the structure being investigated. In order to locate and define the extent of fault-zones using MASW and to spatially correlate the profiles with each other and with geomorphological and geotechnical data, the MASW profiles were interpreted in 3D using AutoCAD. The profiles were imported into the geomorphology drawing previously created from survey data and hung in a vertical plane from the geophysical lines defined by EDM surveying of their starts, ends and intermediate positions. These profiles could then be viewed along the structure contours using the structural UCS defined in 3.2. In order to enhance visual identification of changes in surface gradients the entire dataset was exported as an AutoCAD block and inserted into a new drawing with vertical exaggeration of up to x10.
4 CALIBRATION OF MASW TO THE SURFACE EXPRESSION AND ENGINEERING PROPERTIES OF FAULTED TORLESSE GREYWACKE AT DALETHORPE

4.1 Introduction

The presence of a significant upstream-facing fault scarp on a strand of the Springfield Fault at Dalethorpe, Canterbury (the SFD) has hazard implications for Canterbury as a whole and particularly for the nearby rangefront town of Springfield, 70 km west of Christchurch on State Highway 73 (Figure 4-1). The scarp strikes to the northeast across and displaces a flight of glaciofluvial terraces in the Upper Hawkins Basin, labelled from A to C with increasing elevation and age. Its along-strike projection passes almost directly under Springfield, some 7 km distant. In keeping with the basin’s rangefront location the views from the displaced terraces are outstanding and the area would undoubtedly attract significant attention if the landowner were to subdivide. On initial inspection the fault-zone appears well defined with the fault only obscured for a short distance under the lowest terrace.

4.1.1 Scope and objectives

This chapter reports on the field and laboratory exercise to correlate laboratory, survey and outcrop scale S-wave velocity structure with the underlying rock-mass properties of the SFD and to relate these properties to detailed geomorphic indicators of faulting at this locality. A number of MASW surveys are used to locate and define the extent of the SFD, and to define within-zone variability. The S-wave velocities obtained from the surveys are compared and correlated with geotechnical and sonic/ultrasonic seismic data derived from outcrops and in the laboratory. The MASW surveys are also compared and correlated with detailed topographic surveying of ground deformation.

4.1.2 Previous work

The fault-zone of the SFD has been the subject of several unpublished previous investigations. Speight (1928) identified and mapped the fault. Evans (2000) investigated
the paleoseismicity of the SFD. A further review of his work is reserved for Chapter 5, but of relevance to this chapter is that he was unable to positively identify any additional faults in the area other than the Bell Hill Fault, for which good outcrop evidence was available (Figure 4-1).

A field party from the Swiss Federal Institute of Technology (ETH) in Zurich carried out a seismic reflection survey at Dalethorpe. Limited processing of the survey revealed little of interest and the data were shelved (Green, 2004). Corboz (2004), also from ETH, used ground penetrating radar (GPR) and resistivity tomography (Figure 4-2) to survey the terrace above the projection of the Springfield Fault through Terrace A (Figure 4-1). The survey resolved three discrete shear zones in the overlying, visually undisturbed, gravels but does not resolve shears below the bedrock interface at ~5 m. Paleoseismic implications of the Corboz study are covered in Chapter 5. The easternmost of these shears is thought to outcrop just west (left) of outcrop D1 (Map 1, Figure 4-1) where heavily sheared Torlesse sandstones are exposed (see Figure 4-15). The size of the survey was, therefore, limited both in extent and in penetration by the use of GPR and covers only a fraction of the area covered in this study (See Figure 4-3).
4.1.3 Geomorphology and geology of study area

The SFD transects the Upper Hawkins Basin, which is surrounded by the bedrock hills of the Canterbury rangefront (Figure 4-1). The underfit Hawkins River ('the south branch') and minor tributaries feed the Upper Hawkins basin from the west and northwest, hugging the western and southern sides of the basin. A larger tributary (the 'north branch') flows southward into the basin along its eastern boundary. Only minor drainage is found in the central basin. The basin drains through a gap in its southeast corner, from whence the Hawkins River eventually joins the local baselevel of the Selwyn River, perched on the outwash terraces above the incised Waimakariri River. The south branch shows some interesting gross characteristics in the vicinity of the fault trace, most notably a series of northeast trending doglegs separated by eastward flowing reaches (Map 1 and inset-Figure 4-1).

Figure 4-2: A) Integration and interpretation of GPR and resistivity tomography at Dalethorpe (Corbox, 2004). Corbox delineates boundaries between three subhorizontal units. High resistivity values to the east (Z2) correspond to dry sand and/or gravel seen above outcrop D1, whereas lower resistivities to the west (Z1) indicate higher water content (near the swamp) and/or silt/clay content. Bedrock is invoked to explain both the limited penetration of GPR and lower resistivity values at > 5 m depth. B) 3D GPR 'cube' showing the time to a major GPR reflector. Shorter times are darker indicating uplift to the east.
A narrow deeply incised reach is located just downstream of the projection of the obvious fault trace. The bedrock geology of the area is dominated by the Rakaia subgroup of the Jurassic to Lower Cretaceous Torlesse Supergroup (Andrews et al., 1976). In this area the unit comprises the typical turbidite sequence of interbedded massive quartzofeldspathic sandstones and argillites. The beds are steeply inclined as is typical of Cantabrian Torlesse, and strike approximately north-northeast, subparallel to the strike of the SFD. An outlier of Cretaceous Broken River Coal Measures and View Hill Volcanics (Speight, 1928) is found in fault contact with Torlesse in the NE of the study area, but beyond providing evidence of faulting it is not significant in terms of this study. Clear bedrock exposure of the SFD is limited to a single outcrop in the south branch, where the river meanders across the fault and cuts in hard against the valley side. At this location the fault-zone comprises at least two sub-vertical foliated shear zones separated by intensely fractured zones that display low angle thrusts. These thrusts can be seen in several places to offset attenuated, boudinaged and folded argillites (Figure 4-15, section 4.3.3).

Above the river, particularly to the north, glacio-fluvial terraces comprising variable thicknesses of gravels on bedrock strath surfaces occupy the majority of the central basin. The terraces are labelled by decreasing age from C through A (Map 1 and inset Figure 4-1) and step down southward toward the modern south branch. The intersection of the fault with these terraces forms a fault scarp that increases in elevation and offset away from the modern river channel. Details of this geomorphology are discussed further in the next chapter. Most importantly for this chapter, the youngest and lowest of these terraces (Terrace A) does not have an apparent scarp, other than ponding due to impeded drainage that has created a swamp on the terrace tread above the outcrop of the fault in the south branch on terrace A, although the GPR shows shearing in the gravels. This terrace conceals the fault-zone that is evident both to the north and south and is the surface used for this chapter’s geophysical investigations, to test the efficacy of the technique to define hidden fault-zones.

4.2 MASW profiles

4.2.1 Data collection

Seven MASW surveys were carried out across and approaching the SFD over a period between December 2006 and July 2007. The lines were run to investigate the previously-mapped fault and fold, the apparent lack of footwall deformation and to look for within-
fault-zone variability. Together they cover an interpretable area of approximately 42 hectares relative to the strike of the fault (Figure 4-3).

Lines 1 and 2 were acquired as a pilot trial in April 2006 by a class of graduate students and overlap across what was then considered to be the projection of the SFD through outcrop D1. Line 1, run on terrace A immediately north of the south branch of the Hawkins River, approaches and crosses the SFD from the footwall side. It intersects the projection of a major shear zone that outcrops in the river bed (outcrop D1 – see section 4.3.3.1 Crosshole velocity testing) and is roughly aligned with the projection of the fault trace. Line 2, run along the Dalethorpe road (a metalled single lane farm road on the south side of the river), approaches the SFD from the hanging wall oblique to Line 1, and ends 145 m away.

![Image](image.png)

*Figure 4-3: The Dalethorpe surveys covered an area of 726 m by 572 m relative to the strike of the fault. The western end of Line 3 (labelled twice) overran and replaced Line 2. The outside of the meander in the river bend at the SW end of line 6 exposes a shear zone (outcrop D1). No obvious scarp is developed in the terrace crossed by the lines. The orange square indicates the area surveyed by Corbaz (2004) using GPR.*

Lines 3-6 were surveyed in December 2006. Line 3 was run from east to west. Almost 700 m long, it took less than 6 hours to survey. It replicates line 2 and extends it further east along the Dalethorpe road into the hanging wall to look for evidence of folding and/or faulting. Lines 4, 5 and 6 were acquired with a further 8 hours work in the vicinity of line 1 on a grassy, sometimes boggy or tussocky, surface comprising predominantly Holocene fluvial gravel with little soil development. The terrace is mapped as being underlain by Torlesse greywacke bedrock, but this is uncertain because of the possibility
of Cretaceous outliers in the footwall. Surface waves behave unpredictably where the
topography is of a shorter scale length than the far offset so these lines were surveyed
with shorter near offsets in order to survey as close to the bounding terrace risers as
possible without needing to change the source position mid-survey. Lines 4 and 6 for
instance, though relatively short, closely approach the terrace boundary at both ends. Line
4 was run from SE to NW, orthogonal to the strike of the fault trace so as to sample
normal to the fractures. At the NW end of Line 4 the streamer was rotated around its
centre point to form the start of Line 5, which continues westward parallel with the terrace
edge and with line 1. Line 6 was sited to investigate within-zone variability in S-wave
velocity and was run parallel to the strike of the fault from NE to SW at right angles to
line 4, finishing directly above the shear zone outcrop.

The final line, line 7, was surveyed from east to west on a windy day in July 2007.
Although inconvenient, the wind provided an opportunity to assess the utility in common
Canterbury conditions. The wind was westerly, gusting to 55 km/h, and blew strongly
around and through the steel geophone sleds. The line was run on terrace A on the south
side of the river, beginning parallel to and 8 m north of line 3 and then diverging from it.
Whereas line 3 had followed the bend in the road, line 7 ran due west in a straight line.
This line had multiple objectives. Firstly, it was run to investigate whether the near
surface high velocities seen in Line 3 were influenced by the coincident change in
direction of the line. Essentially, was there directional influence in the velocities due to factors
such as a preferred fracture orientation? Secondly, processing the earlier lines had
highlighted the difficulty in separating the fundamental mode when surveying with a short
geophone string (see section on Dispersion Curve Extraction). Line 7 provided an
opportunity to experiment with larger near offsets whilst also acquiring walkaway
records. Two shot points, one at 12 m and one at 36 m, were used at each array position.
This yielded four separate datasets including fixed receiver (FRW) and fixed source
(FSW) walkaways. Line 7 also forms the basis of an assessment of the repeatability of the
technique.

All surveys were acquired following the roll along methodology outlined in Chapter 3. No
site calibration was carried out and the near-offset was the primary parameter that varied
between surveys (Table 4-1). Cultural noise was limited to mostly-distant livestock
movement and the main environmental noise source was wind.
4.2.2 Dispersion curve extraction

The seismic records were preprocessed, walkaway records were constructed and dispersion curves were extracted following the methodology outlined in Chapter 3. Received bandwidths varied widely but were typically between 10 and 50 Hz. Within this range, coherent dispersion in some cases covered ranges of only 10 Hz. Although coherence was typically low, the surface wave energy generally dominated the records (Figure 4-4). The quality, dispersiveness and phase velocity of the seismic records were variable but often noticeably systematic. For example, dispersion curves evaluated from records acquired at the eastern end of line 3 were typically moderately dispersive and displayed low phase velocities and a high signal to noise ratio. These records were collected in the distal hanging wall relative to the mapped fault.

Table 4-1. Acquisition parameters for Dalethorpe MASW surveys.

<table>
<thead>
<tr>
<th>Survey line</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7a</th>
<th>7b</th>
<th>7c</th>
<th>7d</th>
</tr>
</thead>
<tbody>
<tr>
<td>Acquisition</td>
<td>24 ch</td>
<td>24 ch</td>
<td>24 ch</td>
<td>24 ch</td>
<td>24 ch</td>
<td>24 ch</td>
<td>24 ch</td>
<td>24 ch</td>
<td>48 ch</td>
<td>48 ch</td>
</tr>
<tr>
<td>Walkaway</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>FSW</td>
</tr>
<tr>
<td>construction</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>PRW</td>
<td></td>
</tr>
<tr>
<td>Aperture/ Array</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>23 m</td>
<td></td>
</tr>
<tr>
<td>dimension (D)</td>
<td></td>
<td>23 m</td>
<td>23 m</td>
<td>23 m</td>
<td>23 m</td>
<td>23 m</td>
<td>23 m</td>
<td>23 m</td>
<td>23 m</td>
<td>47 m</td>
</tr>
<tr>
<td>Near offset (x1)</td>
<td>8 m</td>
<td>8 m</td>
<td>12 m</td>
<td>10</td>
<td>10</td>
<td>10</td>
<td>12 m</td>
<td>36 m</td>
<td>12 m</td>
<td>12 m</td>
</tr>
<tr>
<td>Shot spacing</td>
<td>5 dx</td>
<td>5 dx</td>
<td>4 dx</td>
<td>4 dx</td>
<td>4 dx</td>
<td>4 dx</td>
<td>4 dx</td>
<td>4 dx</td>
<td>4 dx</td>
<td>4 dx</td>
</tr>
<tr>
<td>(m)</td>
<td>(5 m)</td>
<td>(5 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
<td>(4 m)</td>
</tr>
<tr>
<td>Sampling interval (ms)</td>
<td>0.25</td>
<td>0.25</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>Recording time (ms)</td>
<td>1500</td>
<td>1500</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
</tr>
<tr>
<td>Number of records</td>
<td>59</td>
<td>41</td>
<td>157</td>
<td>21</td>
<td>48</td>
<td>13</td>
<td>81</td>
<td>81</td>
<td>81</td>
<td>81</td>
</tr>
<tr>
<td>Survey length</td>
<td>295</td>
<td>205</td>
<td>624</td>
<td>84</td>
<td>192</td>
<td>52</td>
<td>324</td>
<td>324</td>
<td>324</td>
<td>324</td>
</tr>
</tbody>
</table>

Figure 4-4. Examples of records from each survey line. Surface wave energy is dominant in all records but records are adversely affected by wavefield scattering. This is particularly noticeable for line 4.

An abrupt change in record character occurred at shot records 70+ (Dispersion curves 170+, Figure 4-5), which were acquired in closer proximity to the mapped fault. These records proved to be strongly dispersive but had poor signal to noise ratio. They displayed higher phase velocities, greater wavefield scattering and reduced coherence together with
a change in recorded surface wave bandwidth. They also ubiquitously exhibited a bimodal amplitude spectra with a second peak at frequencies of 80-120Hz that was absent from the lower phase velocity records. Noise and scatter in these fault-proximal records often took the form of surface wave diffractions, amplitude attenuation at some frequencies or within-record backscatter. Backscatter occurs where surface waves propagate after reflection off a velocity contrast. This results in reverse dipping events in the shot gathers, which together with other noise suggests significant lateral velocity variations.

Figure 4.5. Contiguous dispersion curves taken from the centre of line 3. A noticeable shift in dispersion occurs at record numbers greater than 170. Some of the lower frequency Rayleigh wave phase velocities are excessive suggesting body wave contamination of the records. Dalethorpe data were the first to be processed during the course of this study and techniques mentioned in Chapter 3 were subsequently adopted to deal with this problem.

A further record character change occurred over the final 100 m of the line. This was marked by a second shift in the average position of the dispersion curve. The SNR improved and the curves flattened out to become only slightly dispersive. This character was then maintained for the remainder of the line.

Line 7 was also acquired on the hanging wall. Most notable in processing the Line 7 records were the relative ease of processing 24 channel and corresponding 48 channel walkaway records. Modal separation was difficult on the 12 m offset: records and unremediated higher mode contamination led to further difficulties with processing and exaggerated phase velocities on the dispersion curves. The 36 m offset records proved useless as a standalone dataset but were invaluable for modal separation using walkaway records. Both FRWs and FSWs were constructed, with the latter proving to be the better option for this site (Figure 4.6). The FRW records were badly affected by repetition of systematic noise and probably also by static errors introduced by walkaway construction.
(see Appendix A). The FSW records, however, facilitated modal separation and provided a stable result. Line 7 followed a very similar progression of record and dispersion characteristics to Line 3.

Figure 4-6. A) Comparison of fixed source and fixed receiver walkaway records centred over the same point in the subsurface. Using FRWs is shown to result in repetition of systematic noise, particularly obvious for diffractions. B) Direct comparison of dispersion curves generated using a 24 channel records with both types of 48 channel walkaway records. The 24 channel dispersion curve is generated using the first half of the FRW. The higher mode jump seen in the 24 channel curve is not present in the FSW curve, because the modes were more effectively separable at far offsets. The FRW curve, however, exacerbates the modal jump. Phase velocity differences between the 24 and FSW curves are probably due to bulk averaging over a longer distance that includes higher velocity material outside the 24 channel record.

The third hanging wall line, Line 4, overran the projection of the sheared outcrop in the river, and it predictably produced the least coherent records. Even the best records were significantly backscattered (Figure 4-7) and dispersion curves were difficult to extract with poor signal to noise ratio. They generally displayed relatively high phase velocities, but were extremely variable. The fault parallel survey Line 6 was similar in most respects to Line 4.
Lines 1 and 5 were dominantly on the footwall of the mapped fault. Their records were typically moderately dispersive and only weakly coherent (Figure 4-8).

4.2.3 2D S-wave velocity sections

For each survey line, the dispersion curves were inverted and 2D S-wave velocity sections were produced. As depth of investigation is a function of the longest wavelength (Appendix A), variability in bandwidth caused variability in the depth of Vs profiles. Results are presented first for the initial surveys (Lines 1, 3 4, 5 and 6, Figure 4-9) and then for the subsequent survey (Line 7, Figure 4-10).
Figure 4-9. Some scale S-wave velocity profiles for all but line 7. 1D inversions contoured to produce these images are spaced at the shot spacing indicated on the scale bar.
Figure 4-10. Comparison of results over the same ground using different survey parameters and walkaway constructions. A) 12 m offset 24 channel. B) 36 m offset 24 channel. C) Fixed receiver walkaway. D) Fixed source walkaway. Shot/inversion spacing is 5 stations.

The transition from cover material into completely weathered Torlesse can be inferred at S-wave velocities of around 500 m/s (Perrin, 2008 #260 Perrin). Weathering in this case refers to the processes of in ground alteration due to fluid migration through fractured
rocks. In the sections presented here this boundary typically occurs at slightly less than 5 m depth, which agrees with observations on site and with the geophysical interpretations of Corboz (Corboz, 2004). Velocities in excess of 1200 m/s or 1300 m/s, however, are unlikely in the upper 100 m of Torlesse greywacke (Buch, 2008; Perrin, 2008, pers. comm) (Table 4-2). This suggests that some of the MASW data presented here are contaminated with body waves. This problem generally only exaggerates the lower frequencies and hence the velocities close to and within the halfspace. Dispersion curves with low frequency Rayleigh wave phase velocities higher than the expected S-wave velocity, such as are seen in Figure 4-5, are almost certainly contaminated by body waves.

**Table 4-2. Torlesse greywacke data, presenting shallow surface S-wave velocities for Torlesse argillites and sandstones for different degrees of weathering/fracturing, after (Perrin, 2008, pers. comm).**

<table>
<thead>
<tr>
<th>Degree of weathering</th>
<th>S-wave velocity v_s</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>CW</strong> (completely weathered)</td>
<td>500-550 m/s</td>
</tr>
<tr>
<td><strong>HW</strong> (highly weathered)</td>
<td>550-750 m/s</td>
</tr>
<tr>
<td><strong>MW</strong> (moderately weathered)</td>
<td>750-1000 m/s</td>
</tr>
<tr>
<td><strong>SW-UW</strong> (slightly to unweathered)</td>
<td>1000-1200 m/s</td>
</tr>
</tbody>
</table>

Notwithstanding the high velocities in the halfspace, the velocity gradient at depths >5 m vary significantly across the site even though the entire survey area is underlain by Torlesse greywacke. This can be caused by spatial variations in either one or a combination of weathering, effective stress, pore space and fracturing (Fumal, 1978). On a site such as this a combined effect is most probable.

Above the high and mid velocity strata interpreted as bedrock is a low velocity layer extending from the surface to depths of 5 m. The velocity in this layer is typically higher in line 3 than in line 7, probably due to running line 3 on a well-used metalled road. These low velocities are interpreted as being the weathered greywacke gravels that are immediately encountered when attempting to auger the soil profile on terrace A. No loess profile was encountered (Phil Tonkin, 2007 pers comm.). Mid-velocities are notable at the surface in both of lines 3 (Figure 4-9) and 7 (Figure 4-10D), and these relate to an area of the terrace surface littered with boulders, that are probably the products of mass wasting of the hill to the south.
4.3 **MASW correlations**

The inverted MASW profiles were correlated with each other and with geomorphological, geotechnical and smaller scale seismic data. This section outlines the results of those correlations.

4.3.1 **Geomorphology**

A wealth of surveyed topographic profiles has been amassed at Dalethorpe, both during this study and by Evans (2000) and their locations are shown in Figure 4-11. This section details the strong and often surprising correlations that were observed throughout this study area between these profiles and the 2D MASW profiles. Correlations were made using a structural UCS defined by structure contours on the mapped fault. The contours align with the repeated doglegs in the river.

4.3.1.1 **General correlations**

A site-wide along-strike correlation of the MASW profiles with topographic profiles is presented that agrees strongly with much of the structure predicted by field and structure contour mapping of the fault trace and associated geomorphology (Figure 4-12). Additional structures can, however, be identified. The structure of the fault-zone will be covered in detail in Chapter 5, but several points relating to structural interpretation of depth to bedrock on the S-wave velocity profiles must be made here. The approximate strath profile is drawn in assuming a transition into completely weathered Torlesse at 500 m/s (Section 4.2.3, Table 4-2). Much of this surface would originally have been a continuous, evenly graded bedrock strath below the A surface. The uplift rates make it improbable that any ancestral channels exist below the visible strath.

The velocity-based depth to the strath is indeed relatively flat across a great deal of the site, at an average depth of about 5 m, but several significant departures can be seen. Three areas of deep low velocity can be seen below the strath and are interpreted as fault damage zones (Figure 4-12). The central damage zone is located in the hanging wall of the main mapped thrust. Although the surveys could not cover the projection of the mapped fault through the dogleg in the river, the width of the damage zone appears to be greater in the hanging wall.
Figure 4-11: Location of topographic profiles and their relationship to MASW profiles. Image rotated 90° to place fault strike (051) in "N-S" orientation. Topographic profiles 1-24 after Evans (2000). This figure is formatted to fold out for ease of reference whilst reading this chapter and Chapter 5.
The western footwall damage zone underlies much of MASW line 5, between eastings 2415918 and 2416059 (Map 1). It is of similar width, depth and velocity to the mapped fault’s (central) damage zone, suggesting that the causative faults are of a similar scale. It is interpreted as being related to a previously unmapped imbricate thrust in the footwall of the main mapped fault. Assuming the imbrication damage to be similarly concentrated in the hanging wall, the projection of that thrust passes through the dogleg west of outcrop D1, beyond which there is a steady rise in velocity. There is no other geomorphic expression. Although the fault was not previously mapped the river dogleg would certainly suggest a mapped-fault-parallel structure at that location.

The eastern damage zone indicated in Figure 4-12, however, is unlikely to have been noticed without MASW. It lies some 290 m southeast of the previously mapped fault scarp. The major westward shift in the dispersion curves in lines 3 and 7 (Figure 4-5), towards higher phase velocity for a given frequency, occurs at this location. This shift is modelled in the inverted profiles as a large and sudden eastward reduction in velocity or
depth to bedrock. The beginning of the velocity reduction correlates with a knick-point in the thalweg profiles of both the north and south branches of the river. The step is more obvious in the north branch, which is a smaller 1st order stream and where it can be measured at 0.72 m in height. Once again assuming the damage to be concentrated in the hanging wall, a fault is mapped here passing through the base of the terrace riser, and through the river just downstream of the knickpoint. The central and eastern damage zones bound a zone of elevated velocity where the surveyed location of the strath correlates closely with the interpreted depth and both indicate that the strath is clearly warped up by ~2.5 m.

The eastern damage zone is mapped as a major downstream facing thrust fault (termed the Main Downstream Thrust – MDT) based on the sense of motion indicated by the strath warping. The fault is not visible in outcrop, but its presence has a strong effect on the local geomorphological development. The location of the fault on MASW line 7 coincides exactly with the location at which the array crossed a small structurally-controlled gulley. The gulley approaches the survey line from the road, intersects the line and disappears into the river. The MASW inferred damage zone and the gulley are correlated along strike with a matching gulley on the opposite side of the river and with the southeastern B – A terrace riser, which is significantly higher than the same terrace riser upstream on the footwall side of the previously mapped fault (see profiles 7 and 20, Figure 4-12, Figure 4-13). On this basis, the terrace riser is remapped as a fault-controlled scarp, modified by river erosion during synchronous deformation and downcutting of the river from terrace B-A as the river flowed diagonally across the valley floor to the northeast. The nature of the processes operating here are discussed more fully in the next chapter.

In addition to low velocities below the inferred original strath position, up-warping of the strath can be observed between the central and eastern damage zones.

4.3.1.2 Detailed correlations

A particular surprise was the correlation observed between lines 3 and 7 and detailed topographic surveys. This correlation supports the identification of at least 2 and possibly a third previously unmapped faults in addition to the MDT, none of which display any immediately obvious surface displacement. Figure 4-13 shows the location of the mapped fault (Evans, 2000) (northwestern-most fault Figure 4-13 A) and the locations of the four newly mapped faults including the MDT. The new faults are numbered on the elevation
view (Figure 4-13B) from 1-3, northwest to southeast. Fault 1 shows up as a deep velocity low in the MASW profile, which correlates closely with a deep pool in the riverbed. It also correlates with an eastward step up in the strath profile. This indication of upthrow to the southeast is confirmed by the river profile. East of this fault the river grades to a higher upstream level than the upstream reach of the river, confirming that the fault is upthrown to the east. This fault is visible in river bank outcrop but its sense is not obvious.

Figure 4-13. Detailed map (A) and MASW pseudo-section view of lines 1 and 3 (B) in the central part of the Dalethorpe structure showing newly mapped faults. For discussion see text.

This fault also correlates with an embayment in the terrace edge on the south side of the river. Fault 2 is picked out by a significant eastward velocity increase that correlates closely with an eastward flattening of the river profile observed in both the north and south branches of the river. Again, this suggests up-throw to the southeast. This is borne out by a correlative up-warping of about 0.5 m in a very young inner terrace profile (profile 26, Figure 4-13), and an eastward step up in the strath profile. The inner terrace appears to be an isolated degradational feature coinciding with this uplifted reach. A northwestward dipping shear zone is visible in outcrop on the north side of the river, but again its sense is unclear.

Identification of the third fault is rather more tenuous. The fault is inferred based on an eastward velocity reduction that correlates with a knickpoint in the river profile. Although not mapped as such here, there is some suggestion of a similar structure approximately
20 m further west. Neither structure is visible in outcrop but outcrop is typically scarce and highly degraded in this section.

The high velocities recorded above the strath surface in Figure 4-13B are probably related to a bouldery deposit, which litters the present terrace surface at this location only. These relatively intact and unweathered boulders have higher velocities than the underlying weathered and fractured bedrock strath.

4.3.2 Line to line correlations

A significant question to address when first applying a technique such as this is repeatability. Can a survey be repeated with confidence? The proximity and orientation of line 3 relative to line 7 facilitated a simple investigation of the repeatability of the technique (Figure 4-14). The lines were run on different days with different teams, survey parameters, locations, orientations and ground and weather conditions. Their commonality was that they both investigated the same part of the structure. Along strike lateral correlation between lines 3 and 7 is very strong, which is especially surprising in view of the differences in orientation and location between the lines (see Figure 4-11). The location of the eastward rise in velocities at Figure 4-14 [3] coincides clearly, as does the drop in velocity into the MDT (Figure 4-14 [2]) and the velocity recovery in the footwall of the MDT (Figure 4-14 [4]). The strong lateral correlation of diverging lines testifies to the along strike linearity of the structure.

![Profiles vertically exaggerated (x10). Lines at 1 m spacing](image)

**Figure 4-14:** Along-strike comparison of MASW line 3 alone (A) with line 7 superimposed on line 3 (B) to investigate repeatability of the technique. The vertical exaggeration is x10 and the horizontal lines are at 1 m spacing. Both MASW profiles can be seen to record some major similarities.
Depth correlation is also strong between lines 3 and 7. The elevation of the low velocity contours across the top of the uplift at Figure 4-14 [1] varies by only \( \sim 1 \) m between lines, and even this is likely to be due to either natural variance or southwestward plunge of the anticline structure between faults 1 and 4. The elevation of velocity contours >1000 m/s coincide almost exactly across the uplift. A significant depth variance is, however, seen at (3) where line 3 extends the low velocity zone to an unrealistic depth compared with line 7. This is probably a processing artefact, caused by the use of a 10 m near offset on line 3 with no farther offset shot to allow walkaway construction.

4.3.3 Rock mass character

4.3.3.1 Crosshole velocity testing

In addition to MASW profiles, horizontal crosshole velocity data were acquired at an outcrop (D1) close to the main thrust, at the southwestern end of line 6. Table 4-3 presents a summary of the physical properties and associated S-wave velocities at the outcrop. The velocities vary from 63 to 450 m/s and generally reflect the structural element being sampled (see Figure 4-15 to Figure 4-17). Large reductions in \( V_s \) are apparent in the two major shear zones relative to surrounding rock, and in the sheared argillite. The highest velocities were recorded on either side of shear zone 1 (holes 8-9). The extremely low velocity between holes 11 and 12 may be due to hole 12 being drilled at a low angle to the face to follow the strike of the argillite. This may have caused poor coupling of the geophone with the base of the hole without noticeably impairing the performance of the slide-hammer.

As shown in Figure 4-15, outcrop D1 exhibits a strong damage asymmetry. The visual appearance of the media either side of shear zone 1 appears different and one would expect to see this reflected in the velocity structure (e.g. Dor et al., 2006). Unfortunately no strong trend emerged, possibly due to the amount of sheared argillite sampled on the SE end of the outcrop.

When the crosshole velocities obtained at outcrop D1 are correlated with lines 1 and 4, several well defined low velocity zones can be seen within the strike-perpendicular range of outcrop D1 (Figure 4-17B). The spacings and locations of these velocity lows are very similar to and closely associated with shear zones seen in outcrop. This appears to emphasise the structural sensitivity of the technique but the results of Corboz (Corboz, 2004) show that the major shears in this vicinity are not linear over the distance between
the outcrop and Line 4, at least in the gravels (Figure 4-17A). The structural correlation with line 1 is, however, solid. Two well developed velocity lows can be observed that correlate exactly with the foliated shear zones observed in outcrop between holes 1-2 and 8-9 (Figure 4-17C). The incipient shear zone between holes 4-5, which is highlighted by a low crosshole velocity, is not obvious in the MASW profile but is also not foliated in outcrop.

Table 4-3 Near-surface physical properties and S-wave velocities at outcrop D1, Dalethorpe.

<table>
<thead>
<tr>
<th>Hole pair</th>
<th>Structural element</th>
<th>Distance (cm)</th>
<th>Number of fractures</th>
<th>Minimum fractures/m (density)</th>
<th>Maximum spacing (mm)</th>
<th>V, (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-2</td>
<td>Foliated gouge</td>
<td>58</td>
<td>&gt;106</td>
<td>185</td>
<td>5.4</td>
<td>81</td>
</tr>
<tr>
<td>2-3</td>
<td>Fractured and faulted</td>
<td>168</td>
<td>&gt;186</td>
<td>110</td>
<td>9</td>
<td>253</td>
</tr>
<tr>
<td>3-4</td>
<td>Fractured and faulted</td>
<td>205</td>
<td>&gt;175</td>
<td>86</td>
<td>11.7</td>
<td>259</td>
</tr>
<tr>
<td>4-5</td>
<td>Incipient shear zone</td>
<td>220</td>
<td>&gt;250</td>
<td>114</td>
<td>8.8</td>
<td>184</td>
</tr>
<tr>
<td>5-6</td>
<td>Fractured and faulted</td>
<td>265</td>
<td>&gt;145</td>
<td>55</td>
<td>18.3</td>
<td>255</td>
</tr>
<tr>
<td>6-7</td>
<td>Fractured and faulted</td>
<td>245</td>
<td>&gt;130</td>
<td>54</td>
<td>18.9</td>
<td>255</td>
</tr>
<tr>
<td>7-8</td>
<td>Fractured and faulted</td>
<td>340</td>
<td>&gt;224</td>
<td>65</td>
<td>15</td>
<td>450</td>
</tr>
<tr>
<td>8-9</td>
<td>Foliated gouge</td>
<td>190</td>
<td>&gt;212</td>
<td>111</td>
<td>9</td>
<td>174</td>
</tr>
<tr>
<td>9-10</td>
<td>Massive sandstone</td>
<td>260</td>
<td>&gt;198</td>
<td>76</td>
<td>13</td>
<td>434</td>
</tr>
<tr>
<td>10-11</td>
<td>Sheared argillite</td>
<td>105</td>
<td>&gt;175</td>
<td>167</td>
<td>6</td>
<td>157</td>
</tr>
<tr>
<td>11-12</td>
<td>Sheared argillite</td>
<td>105</td>
<td>&gt;180</td>
<td>167</td>
<td>6</td>
<td>63</td>
</tr>
<tr>
<td>12-13</td>
<td>Sheared argillite and sandstone</td>
<td>190</td>
<td>&gt;160</td>
<td>84</td>
<td>12</td>
<td>272</td>
</tr>
</tbody>
</table>

Velocity value correlation between line 1 and outcrop D1 is reasonable although the maximum values in outcrop are generally some 200-300 m/s slower than those observed in the survey. This discrepancy is almost certainly due to a combination of enhanced fracturing and greater fracture aperture at the unconfined surface of the outcrop, and the shallow depth of the crossholes. Shakeel and King (Shakeel and King, 1998) found that blast damage loosened material to a significant depth, which resulted in a noticeable decline in velocity towards the free face. The properties of the D1 rock mass show significant fault-related damage (Figure 4-15), and the foliated zones are noticeably spongy in outcrop.
Figure 4-15: Gross features of outcrop D1 (for location see Figure 4-3 or Map 1). A) General view of fault-zone outcrop cut by meander in south branch of the Hawkins River. B) Close up of outcrop. The footwall is intensely brittle deformed by multitudes of low angle thrusts, a few of which are shown along with a major 1 m wide foliated shear zone. The outcrop is more massive and less shocked on the hanging wall where most of the deformation is concentrated in the argillites. The yellow lines pick out the sheared argillite beds, which are oriented approximately 010/65SE. C) Close up of ductile deformation: isoclinally folded and boudinaged thin argillite layer within footwall forward of shear zone 1. Elsewhere, the folds are crosscut by low angle thrusts. This is in direct contrast to the continuous but sheared argillite beds on the more intact hanging wall side of the shear zone. D) The hanging wall scree contains significantly larger blocks of sandstone than does the footwall scree (see B). The blocks do not however transport well and typically fall apart as shown when disturbed.
Figure 4-16: Graph of $V_s$ with distance along scanline for outcrop D1. No obvious velocity asymmetry is apparent to match field observations but this may be skewed by the predominance of argillites sampled at the SE end of the outcrop.

Section 4.3.2 demonstrated strong MASW line-to-line correlations whilst lines 1 and 4 appear to belie that claim, in terms of velocity if not structure. This may be due to processing artefacts, the location of line 1 close to an unconfined, actively eroding terrace edge, S-wave splitting or along-strike velocity variability. Leaving aside processing artefacts, unconfined edge effects should not result in low velocities below ground so can probably be discounted. At a given pressure, S-wave splitting in foliated media results in lowest velocities perpendicular to foliation (Okaya et al., 1995). Line 4 is closest to orthogonal to the fault and displays the highest velocities, so S-wave splitting is unlikely to be responsible. That leaves along-strike variability, which is investigated in line 6. The intersection of lines 1 and 6 certainly appears to occur at a low velocity zone that is common to both lines (Figure 4-18).

4.3.3.2 Scanline surveys

Apart from outcrop D1, downcutting of the Hawkins River exposes Torlesse greywacke basement rocks in many places along the river’s course. Several of these outcrops, marked on Map 1 from D2 through D7 are also investigated and classified in this study and their properties are correlated with S-wave velocities inverted in the MASW studies (Figure 4-19A). Outcrops D3 and D4, which are outside the fault-zone as defined in
Figure 4-13. are characterized by wider fracture spacings and are moderately to completely weathered.

Figure 4-17. Relationship for outcrop D1 between crosshole S-wave velocities and those derived from MASW. (A) Detailed map view showing MASW lines in structural orientation. A GPR data cube imaging three shears, numbered A1-3 (Carboz, 2004 and Figure 4-28) is shown in position. Shear A3 can be seen to trend towards crosshole 1 at the NW end of outcrop D1, where it outcrops. (B) S-wave velocity profiles for line 1 and 4 with crosshole locations and low velocity zones marked. Line 1 displays significantly lower velocities, probably due to its proximity to an unconfined edge. Both lines display a deep set of low velocity zones, which are numbered as per Carboz (2004). (C) Marked up photo of U1 showing crosshole locations, shear zones and crosshole velocities according to MASW colour scale. The velocities are clearly more representative of Line 1 than of the more distant Line 4.
Outcrops D2 and D5, which are inside the fault-zone, are more closely fractured but the material is only moderately weathered. Outcrop D1, which is close to the main trace of the mapped fault, is completely to residually weathered, often strongly foliated and everywhere intensely fractured. The outcrops were characterized by Geological Strength Index (GSI) (Figure 4-19B) using the scheme of Hoek et al. (Hoek et al., 1998). Geological strength indexing provides an estimate of the reduction in rock mass strength based on degree of interlocking and joint surface quality. The indices were then plotted against the S-wave velocities obtained in the MASW survey, yielding a linear relationship for the range of GSI and S-wave velocities covered (Figure 4-19C).

A linear relationship such as this seems unlikely and this conjecture is borne out by integrating all crosshole, MASW and scanline data for the Dalethorpe area. The crosshole data from D1 and D2, together with the MASW velocities correlated with outcrops D3 and 4 (outside the fault-zone), plot linearly on a semi-log plot (Figure 4-20). MASW
Figure 4.19. Correlation of geological strength index (GSI) with MASW S-wave velocities. (A) The location of the photographed outcrops is shown on the MASW profile. (B) Geological strength indices derived in the field for the outcrops in A. (C) Plot of GSI against S-wave velocity for the outcrops in A and B. A strong, apparently linear correlation can be observed.
velocities for outcrops D2, 5 and 7 define a similar trend at higher velocity values but these outcrops fall within a part of the MASW survey considered likely to be dominated by higher mode surface waves (Figure 4-20). It is thus likely that use of velocity data from both contaminated and uncontaminated sectors accounts for the ‘linearity’ seen in the GSI trend (Figure 4-19).

Figure 4-20: Integrated scanline (fracture spacing) and velocity data for Dalethorpe. Crosshole data from Outcrops 1 and 2, and MASW velocities for outcrops 3 and 4 (all diamonds) define a trend that agrees with the data in Table 4-2. MASW data from outcrops 2, 5 and 7 (squares) all lie within the area shown by Figure 4-5 to be potentially higher mode contaminated, and this is reflected in their elevated velocities. Magenta markers are velocities for outcrop 2, green markers are for velocities adjacent to shear zone 1.

Fracture spacing is known to exert a greater influence on S-wave velocity than does lithology, hardness or primary porosity in well-cemented lithologies (e.g. Fumal, 1978; Rasolofosaon et al., 2000). Fracture aperture is also significant. The crosshole velocity from D2 is suggestive of a linear fracture density that is double that observed at outcrop but this anomaly is influenced by the wide fracture apertures of between 2-3 mm, with little or no fill. Conversely, the velocities measured adjacent to the major shear zone is suggestive of a fracture density half of that observed, probably due to fracture closure due to local stress regimes.
Based on the data presented, Torlesse S-wave velocities, including those inverted from fundamental mode Rayleigh waves, can be confidently related to fracture spacing such that

$$\Gamma = 283.78 e^{-0.006v_s}$$

(Equation 4.1)

where $\Gamma$ is the linear fracture density in the subsurface. This relationship even appears to hold for the sheared argillites sampled in outcrop D1 suggesting that fracture density does indeed dominate S-wave velocities in Torlesse sediments.

4.4 Laboratory testing

This section presents the results of an ultrasonic analysis of artificially-fractured parent material. The objective of this laboratory work was to supplement the outcrop data and potentially to provide an alternative method of calculating an empirical relationship between fractured rock-mass parameters and S-wave velocities. Samples taken from outcrops D1 and D2, as well as from an intact boulder from the Hawkins River, were tested in the laboratory to determine a range of physical properties including density, porosity, unconfined compressive strength, point load strength and S-wave velocity.

The river boulder had been selected for apparent homogeneity but, in the first instance, several cores were drilled in three mutually perpendicular directions and tested to determine their intact S-wave velocities. 10% unconfined S-wave velocity anisotropy was shown that reduced slightly with increasing uniaxial stress (Figure 4-21).

Still using the intact boulder, an artificially fractured core was tested to determine the effect of fracturing on the ultrasonic S-wave velocities of Torlesse greywacke. This experiment predictably produced the strongest relationship observed in the laboratory (Figure 4-22). S-wave energy was completely attenuated across sawn fractures until coupling was improved with uniaxial stress. This dry attenuation was due to frictional sliding of the fracture surfaces (Wulff et al., 2000). Once uniaxial stress was applied the S-wave velocity for a given specimen was seen to decline dramatically with increased fracturing, especially at lower uniaxial stresses.

Microcracking due to uniaxial unconfined compression strongly influences velocity of ultrasonic waves in dry sandstones. The experiment depicted in Figure 4-22 began with 12 fractures and the number of fractures was reduced at the start of each run.
Figure 4-21: Intact S-wave velocities for specimens cored from Torlesse greywacke river boulder sourced from the Hawkins River catchment at Dalethorpe. The unconfined S-wave velocity is slightly anisotropic [10% using Okaya et al.’s formula \( V_{\text{max}} - V_{\text{mean}} / V_{\text{mean}} \) (Okaya et al., 1993)] but this appears to reduce with compression. For the purpose of this study the unconfined intact S-wave velocity is taken as 2750 m/s.

Figure 4-22. Ultrasonic S-wave velocity plotted as a function of uniaxial stress for an un-fractured specimen (triangles) and the same specimen with saw cut ‘fractures’. The specimen was cored in orientation 3.

When the fracture number was reduced to 8 following the 10 fracture run the velocities suddenly dropped to unrealistically low values. These velocity changes are explainable by
the creation of cracks located in the grain contact areas during the preceding test cycles (Wulff et al., 2000).

Following this experiment, all subsequent testing was done using only minimal uniaxial stresses of <5 MPa to avoid creating unaccounted for fractures. One specimen from each orientation was fractured at regular intervals from a point part way along its length and S-wave velocities measured with increasing linear fracture density. Plotting the results for this ultrasonic data alongside the sonic data revealed no obvious relationship (Figure 4-23A), a problem probably related to the differing ratios of investigation wavelength to heterogeneity scale length (Gettemy et al., 2004; Muerj et al., 1995). An improved fit of the two datasets was achieved by plotting linear fracture density against the velocity decline relative to the theoretical maximum for the wavelength of investigation (Figure 4-23B). This approach yielded a reasonable fit (~20% misfit) between the lightly loaded ultrasonic data and the sonic velocity data.

The suboptimal fit can be significantly improved if the maximum theoretical velocity is increased to 3500 m/s, a figure that would agree with the work of Okaya et al. (1995). Either alternatively or in conjunction with this, the misfit may reflect the influence of weathering on S-wave velocity. The boulder was relatively unweathered, and it could be conjectured that weathering has a greater influence on S-wave velocity at low fracture densities, whilst fracturing dominates the velocities at higher fracture densities. This might explain the convergence of the data at higher fracture densities.

No cores were obtainable from outcrop samples so the physical properties of irregular blocks were determined. Some reasonably strong trends emerged from this testing. S-wave velocity in Torlesse greywacke material is shown to be related to specimen porosity (Figure 4-24A). This suggests that the porosity of Torlesse hand specimens may be a useful indication of volumetric fracture intensity because intact boulder porosities are typically less than 0.5%. A less strong relationship was observed between S-wave velocity and point load strength (Figure 4-24B), although the relationship between porosity and point load strength was very strong (Figure 4-24C).

In all cases, the velocities and physical properties measured for irregular blocks yielded a larger data spread than those determined using core. This scatter can be rationalized. The difference in velocity spread is caused by measuring velocities across non-parallel faces of roughly rectangular blocks rather than along cores. The scatter in physical property estimates is probably due to errors in volume determination using the immersion method.
Comparisons of core and block data in the plots of $V_s$ against porosity and point load strength suggest that the irregular block methods consistently over-state S-wave velocities and other physical properties. Use of the block method is unavoidable given the degraded nature of the outcrops, from which no core is recoverable. It should, however, be possible to account for overestimation during data analysis. If the relative core to block

![Graph A](image1)

Figure 4-23: Comparison of sonic and ultrasonic velocity relationships to linear fracture density. A) Both ultrasonic and sonic (outcrop) velocities decline linearly with fracture density over the range tested. Ultrasonic and sonic velocity fields are widely separated. B) Percentage velocity decline from a theoretical maximum plotted against linear fracture density. A better fit can be seen between the sonic velocity field and the lightly loaded ultrasonic data. Maximum sonic velocity was set at 1200 m/s (Table 4-2) and ultrasonic at 2750 m/s (see Figure 4-21).
Figure 4-26: Relationship between A) ultrasonic S-wave velocity and porosity, B) ultrasonic S-wave velocity and point load strength and C) Point load strength and porosity for specimens from intact boulders and tectonised outcrops of Torlesse greywacke.
errors for intact rock reflect the likely errors in the data from the tectonized specimens, derivation of any empirical relationships should be guided based on the relative spreads of block and core data. Many of the observed relationships would also be improved if the block cutting was more accurate and if it were possible to cut larger blocks with parallel faces. In this instance, however, quantity has superseded quality due to equipment limitations.

4.5 Discussion

The close correlations of geomorphological, geotechnical and seismic techniques exceeded expectations for this site. Despite the smearing effect of MASW and the non-uniqueness of 1D inversions (Xia et al., 2005), structural and geotechnical boundaries within the Dalethorpe Fault-zone correlate closely with changes in S-wave velocity measured at both tomographic (MASW) and crosshole scale. The MASW velocity variations are damped by the range of S-wave velocities within the structural element below the array. However, the magnitude of velocity contrasts in faulted Torlesse creates sufficient perturbation in the average velocity that even subtle features can be detected. Gross features can definitely be clearly detected.

Field parameter selection for future surveys over Torlesse-floorled sites should be simple. Examination of hundreds of records suggests that the optimal near offset, beyond which near-field effects are absent, appears to vary between 12 and 14 m. As near field effects do not adversely affect dispersion curve extraction, the 12 m option should prove acceptable in most cases, thus limiting the potential for higher mode contamination of high frequencies at far offsets. In this study such contamination was contributed to by the use of a 23 m array at 1 m geophone spacings. The strategy employed to best effect was filtering by muting (Ivanov et al., 2005) using 47 m, 48 channel walkaway records to highlight higher modes. Recent commercial application of MASW in Torlesse, however, demonstrated the remarkable improvements obtainable using a 48 channel array with a reduced geophone spacing of 0.5 m (Duffy et al., 2008). Such records can be cut to remove both near and far field effects following examination of a swept frequency record. This is not possible with 1 m spacings without significantly reducing the surface wave energy in the cut record.

The higher mode contamination in lines 3 and 7 is probably due to the shallow depth to bedrock (O'Neill and Matsuoka, 2005) and as such is not a significant hindrance to using
MASW to locate the fault-zone. It does mean that care should be taken during acquisition and processing if engineering use is to be made of the inverted velocities. A subjective scale and other strategies for grading survey quality are appended that should be used in engineering applications of MASW data.

The integration of field and laboratory seismo-technical relationships in Figure 4-23 provides a strong basis on which to infer geotechnical parameters from MASW S-wave velocities. The major issue that remains is that of determining whether velocity changes relate to fracturing, weathering or both. In an area with differential uplift it is not reasonable to assume that vertical velocity gradients are weathering related whilst horizontal gradients are fracture related.

4.6 Conclusions

Spatial variability in the wavefields, the dispersive nature of the MASW data and the magnitude of inverted velocities correlates with the geological and geomorphic context of the survey lines. Overall, a strong correlation is observed between the MASW data and topographic profiles. The transition into and out of the fault-zone is strongly correlated with quantitative assessments of ground deformation, with the river thalweg proving the most sensitive measure of deformation. On the basis of these correlations the extent of the SFD has been re-defined and significantly extended. The within-zone variability that is seen in the MASW images is correlated with crosshole velocity measurements and measurable rock properties particularly linear fracture density. This correlation is refined by laboratory testing and a scheme is proposed to relate both laboratory and field derived velocities with the same measure of S-wave velocities. The laboratory testing also correlates the S-wave velocity of a variety of Torlesse material with secondary porosity in addition to visible fracture density. This additional parameter provides a measure of volumetric fracture density and with refinement could potentially be correlated with bedrock aquifer porosity and permeability.
5 STRUCTURE, KINEMATICS AND
PALEOSEISMICITY OF THE SPRINGFIELD FAULT,
DALETHORPE

5.1 Introduction
The broad geological and geomorphological setting of the Springfield Fault Dalethorpe (SFD) was detailed in the introductory chapter and in Chapter 4. The MASW data presented in Chapter 4 established the existence of a second major strand of the SFD that lies between the Main Upstream-facing Thrust (MUT) and the Bell Hill Fault studied by Evans (2000). This provides a key starting point for quantifying the fault-zone’s hazard potential using structural, kinematic and paleoseismic analyses. This chapter integrates the MASW profiles from Chapter 4 with the detailed geomorphology of the SFD and should be read in conjunction with Map 1.

5.1.1 Scope and objectives
The detailed geomorphology of the Upper Hawkins Basin was mapped with the objective of elucidating the sequence of channel occupation, deformation and abandonment in the uplifted terraces, and placing them in the context of local structural development. Particular attention was paid to delineating and characterizing:

- the fault scarp.
- the extents, gradients and relative ages of the major glacio-fluvial surfaces and subsequent degradational terraces.
- the evolving relationship between SFD and the past and present watercourses occupied by the Hawkins River.
- the strath surface underlying terrace A.
- the river bed elevation as a sensitive indicator of bedrock faulting that may or may not be expressed by the mapped fault trace.

The survey detail is drawn together in Map 1.

This chapter begins with a description of the structure of the fault-zone (Section 5.2) that defends the cross section (Section A-A’, Map 1), followed by a review of the
geomorphological character, evolution and paleoseismic history of the fault-zone (section 5.3). This includes a general description of the tectonic geomorphology of the Upper Hawkins Basin. 3D analyses are used to isolate individual event traces and determine their relative timing and the magnitude of their coseismic displacements. Samples were collected for optically stimulated luminescence (OSL) dating and the results are presented. Long term slip rates and recurrence interval are calculated based on the OSL dating and number and magnitude of events recorded. Paleoseismic analysis is followed by a geomorphological investigation of the kinematics of the fault-zone (section 5.4), discussion and conclusions.

Reconstructing the paleoseismic history of the site is outside the primary objectives of this study, since MASW is not directly useful as a tool for this purpose. However, it is an important step in the reconstruction of the geometry and history of the uplift of the hanging wall structure. This is relevant to the question of predicting the likely distribution and style of off-plane deformation within a fault zone, and as such is an important objective of this study in the context of land use planning considerations.

5.2 Fault-zone structure

The fault trace geometry and structure of the SFD was integrated 3 dimensionally with surface mapping and topographic data in AutoCAD (Map 1 and cross section A-A'). The MUT has a clear-trace length of at least 3.8 km in the study area, over which distance it is obscured for only ~0.5 km. The fault plane was defined by drawing structure contours on the fault trace, which exactly parallels the dogleg in the river between MASW lines 1 and 7. Based on the elevation view in Figure 5.1, the fault trace is relocated from the work of Evans (2000) to pass along the dogleg rather than simply projecting under Terrace A and through outcrop D1. Although Figure 5.1 shows a large apparent deviation in the fault trace, this is probably largely due to burial of the fault trace to depths of <2 m by fan debris and swamp deposit accumulation. This conjecture is supported by soil augering, which encountered an indeterminate thickness of fan debris in the area. Evans (2000) also suggests that even the clearest scarps are significantly modified. The modification adversely affected his diffusion modelling, which failed to produce reliable ages. The MUT has a dip of only 15°, upthrown to the southeast. Despite the regular sinistral sense doglegs in the river, there is no unequivocal field evidence of strike slip offset.
The minor relocation of fault 1 leaves the steeply dipping foliated shears observed in outcrop D1, and in the GPR imaging of Corboz (2004), reassigned as part of the footwall structure. The presence and orientation of these shears must be accounted for in a fault model. Sheared and truncated argillite bedding in outcrop D1 (Figure 4-15) is oriented at approximately 010/65E, in line with the average orientation of most local Torlesse bedding. The steeply dipping shears that truncate the bedding are strike-parallel. These two orientations are both reflected in the GPR survey of Corboz (2004) (shown on Map 1), where his south-eastern shear zone (A3) is similarly oriented to the truncated argillite beds, A1 is strike parallel and A2 appears to represent a diffuse interaction of the two. The shearing observed in the argillites is interpreted as flexural slip thrusting on favourably oriented bedding planes, whilst the strike-parallel shears represent the evolution of the internal structure of the footwall into a set of through-going shears. Subhorizontal shears were also noted (Figure 4-15B), often offsetting the steeper shears and accommodating the forward component of thrust motion.

Figure 5-1: Topographic and river thalweg profiles viewed towards 51° showing the detailed and average traces of the main upstream facing thrust. Profile 13 is run on the B terrace, Profiles 15 and 21 on the A surface (see Figure 4-11). This 10x vertically exaggerated elevation shows a fault dip of 15° to the SE. The dogleg is seen in this elevation as the sudden decrease in the elevation of the riverbed, due to the river flowing along strike. The apparent re-entrant angle in the projected fault profile reflects the gradient on terrace B and consequential drop in the elevation of the base of the fault scarp.
The western damage zone in Figure 4-12 was interpreted in section 4.3.1.1 as a thrust that breaks out through the river dogleg west of the MUT. It is not directly related to the MUT since outcrop D1 is forward of that thrust and has higher velocities. The flexural slip shearing discussed in the previous paragraph, together with the adjacent meander in the Hawkins River, suggest a significant amount of strain is partitioned in growth of a footwall syncline and this conclusion is supported by the rate and shape of the velocity increase in the MASW profiles in Figure 4-12. With its lack of topographic expression, the fault could be an out of syncline thrust with not much movement on it. Such a fault is unlikely to account for the degree of damage suggested by the velocities so an alternative and favoured interpretation is as a previously active splay forward of the main thrust.

The location and apparent sense of faults 1 and 2 are intriguing. Evans (2000) carried out a detailed analysis of slickensides in shear zones of the Bell Hill Fault on the NW face of Bell Hill. He recorded data that appeared suggestive of a component of normal faulting and conjectured that the slickensides may represent normal faulting on the oversteepened hanging wall of the fault. Both faults 1 and 2 are upthrown to the southeast, the same as the main upstream facing fault. The warping of the young terrace profile (profile 26) over fault 2 suggests thrusting, but no such warping is observed at fault 1. For this reason, fault 1 is interpreted as a normal fault, supporting the conjecture of Evans (2000).

The Main Downstream-facing Thrust (MDT) is the largest newly identified fault in this study. As outlined in the previous chapter a scarp previously mapped as the terrace riser between B and A picks out the fault as it strikes to the northeast. Trimming by the Hawkins River during downcutting to terrace A and fans exiting the antecedent stream channels modify the scarp morphology. Other indicators are structurally controlled gullies to the north and south of the south branch and knickpoints in the beds of both the north and south branches of the Hawkins River. The MDT has a significantly shorter topographic expression than the MUT (only 540 m), with a further 300 m confidently inferred based on thalweg-to-thalweg and thalweg to MASW correlations of both the north and south branches of the Hawkins River. The fault is upthrown to the northwest and appears to dip steeply in that direction, which together support interpretation as a backthrust off the MUT. Together the MUT and MDT bound an uplifted block (the pop-up width).

At the southeastern limit of the MASW surveys the profiles of both rivers show a marked increase in gradient that correlate along strike with each other and with a same-sense,
though less obvious, gradient change in profile 21 (Figure 5-1). These features also correlate with an eastward decrease in velocity in MASW profile 3 (Figure 4-13). When seen in a long river profile (Figure 5-2) this gradient increase is preceded by a gradient reduction that creates a 1.3 m high anomaly relative to the average elevation of the riverbed.

Figure 5-2: Map (A) and profile (B) views of the thalweg of the south branch of the Hawkins River. The most notable feature of the profile is the 1.3 m high, 312 m long anomaly in the river bed at 5830 m from the divide.

The vertical expression of the riverbed anomaly is paradoxically greater in the active river bed than on the abandoned A terrace immediately above. One probable explanation is to invoke folding rather than faulting. Local river and basin characteristics, discussed later in sections 5.3.1 and 5.4 respectively, suggest that a footwall syncline is actively growing in
association with the Bell Hill fault scarp further downstream. Although the river bed anomaly may represent a fault, it is more probably a very small parasitic anticline formed by compression along the inner free surface of the syncline developing in the greywacke slab. A fault with 1.6 m of expression in the river bed is unlikely to leave little trace on the abandoned terrace above. The gravel cover on a fold, however, would deform considerably differently to the underlying bedrock and the strain would be more easily disseminated. For this reason, a syncline is mapped at this location.

The Bell Hill Fault forms a well preserved, upstream facing scarp that cuts across the Dalethorpe Road in the easternmost section of the study area. Selective erosion picks out the fault's plane on the hillside on the northeastern side of the Hawkins River, at the confluence of the north and south branches. This allows a reliable projection of the strike and dip of the fault and its southwestward trace across the north face of Bell Hill using structure contours. The constructed trace coincides with the projected position of the fault located by Evans (2000) outcropping in forestry roads high up on Bell Hill. It also seems to account for the roughly west-east trending lineations observed on the 163 (pre-forestry) aerial photographs of the study area (Figure 5-3).

Figure 5-3: Detail from 1963 Ashburton-Kowal River aerial photograph 3706-3, showing lineations (main picture) and inset the trace of the Bell Hill Fault. The meandering of the Hawkins River as it approaches the fault can be clearly seen, as can its increased sinuosity as it approaches the MUT upstream.
5.3 Fault-zone history

5.3.1 Geomorphological principles and observations

Before continuing to a detailed geomorphological description of the study area, it is necessary to outline several fundamental concepts about the genetic nature of terraces. Three types of terraces are understood (Bull, 1990) and all three are present in this area. Firstly, climatic terraces form due to climatically controlled aggradation followed by degradation leaving a fill terrace known as an aggradation surface. An aggradation surface is entirely climate controlled because base level of erosion cannot be attained during a climatic aggradation event. Conversely, tectonic terraces are major straths bevelled at the base level of erosion by streams in dynamic equilibrium with tectonic uplift. Climate and tectonically-controlled responses are not, however, independent of each other and flights of complex response terraces are found below or between major aggradation or strath surfaces. These terraces form in the vertical space provided by uplift of the channel, whether by aggradation or by tectonic uplift. Differentiation between tectonic (major) and complex response (minor) straths is difficult as tectonic straths may be completely backfilled leaving only exposure of the complex strath below (Bull, 1990).

As outlined in the previous chapter, the SFD fault-zone transects the Upper Hawkins Basin. Within this fault-zone, the wedge bounded by the MUT and MDT vertically offsets a flight of glaciofluvial terraces that step up in elevation northward with increasing age. These terraces, designated A through C with increasing age and elevation by Evans (2000), record a history of southeastward migration of the Hawkins River through distinct climatic cycles. Although referred to here as aggradation surfaces, the accumulated gravels are only of moderate thicknesses of a few metres and rest on eroded bedrock straths. Even so, the modern Hawkins River and its catchment are underfit to represent the source of the gravels. The source of the gravels is speculatively related to the Rakaia glaciations via glaciation of the Selwyn catchment that lies beyond the saddle southwest of Bell Hill. This area, however, merits investigation well beyond the scope of this study because there is presently a 200 m elevation difference between the top of the saddle and the headwaters of the Selwyn River below.

The highest, C, surface is best preserved in the west of the basin, where it wraps around the south of the bedrock ridge that occupies the northwest of the study area. Further east toward the MUT it becomes thoroughly degraded and dissected by small streams. The
streams issue from the bedrock ridge and deposit fans on the B surface below. A correlative surface is mapped in the pop-up wedge. Directly north of the uplifted C surface, the downthrown surface is obscured by a debris flow deposit, although a surface remnant remains that, based on its elevation, is an intermediate surface from the incision down to the B surface.

An extensive original B surface (B0) is also well preserved in the west of the basin on the north side of the modern river. This surface is correlated with multiple displacements in the pop-up wedge, each of which is recorded by remanant surfaces incised into the wedge. These surfaces record a waning flow, from a period during which the river had sufficient excess stream power to plane a wide surface across most of the basin. During later uplift events the pop-up wedge was dissected by a much smaller flow than that which planed the B3 surface. At that stage the river was still slipping southward around the nose of the propagating Eastern Russell Range Anticline and had probably begun cutting down to the A surface. The modern downthrown B0 surface in the centre of the study area is extensively buried by the fan and debris flow deposits issuing from the bedrock ridge through a network of streams incised into the degraded C surface. The limits of this fan are unclear as it is considerably degraded.

A well defined terrace riser separates the B and A surfaces in the west. The A surface is the lowest of the surfaces and the only surface preserved on the south side of the Hawkins River. It appears undisturbed but several tectonic signals are apparent. Firstly, drainage appears trapped at the back of the A terrace and flows east parallel to the river, toward the major structural dogleg that defines the trace of the MUT. This trapping of drainage, along with the trimming of a steep A terrace riser north of the river (Figure 5-4), seems to be caused by recently reversed tendency for the river to slip northward since abandonment of the A terrace. Prior to abandonment, the river was trapped on the south side of the basin against Bell Hill, to where it had migrated over the three aggradation cycles preserved in the terrace flight. Secondly, a fault-strike oriented impediment visible on the air photo has further confined the already-trapped drainage just short of the dogleg and led to the creation of a swamp. This feature was mapped as a fault displacement in the gravels by Corboz (2004) (Figure 4-2B, located on Map 1). Thirdly, the aforementioned dogleg is repeated about 300 m upstream of D1 but without the marked increase in sinuosity.
These observations lead naturally to a discussion of the site river channel morphology, which is of considerable interest. At the western end of the basin the valley width of the south branch abruptly increases, along with the river sinuosity. The river cuts a wide strath with a sharp double meander bend as it approaches and passes outcrop D1. This behaviour has been observed elsewhere including on large rivers such as the Rakaia (Yousif, 1987) and on smaller rivers such as the middle Waipara (Campbell et al., 2003; Nicol and Campbell, 2001). Campbell and Nicol (2003), reporting very similar circumstances, make a distinction between the response of the Waipara River to coseismic uplift and to long-term deformation of the bed (see also Chapter 6).

![Map of river channel morphology](image)

*Figure 5-4: Locations and section views showing the valley shape at surveyed sections across the south branch of the Hawkins River (ve x5). The northern bank can be seen to be almost ubiquitously steeper except in the gorge, where it is higher but of roughly equal gradient. Sections 2, 3, 8 and 9 are located very similarly relative to active faults and are all distinctly wider due to meandering. With the exception of the gorge profiles all profiles end on the edge of the A surface.*

Meandering is not simply a response to coseismic uplift. Modern drainage configurations can be interpreted in terms a competition between stream power and upstream deposition rates on one hand, and bedrock resistance and rates of crestal uplift and fold widening on the other (Burbank et al., 1996) The spatial and temporal interplay of these variables controls geomorphic thresholds, across which a river must change its channel form in order to remain in dynamic equilibrium with the uplift. A channel's form is an independent variable controlling water and sediment discharge over short time scales (10<10² years) (Moseley and Schumm, 2001). Only over longer timescales does the
channel form become a dependent variable that must respond to and keep pace with the growth of emergent structures. Where uplift and folding weigh heavier, an antecedent stream will be defeated. The timescales required for channel form adjustment suggest that the meander bend in the Hawkins River is forming in response to ongoing deformation over the width of the meander. Just downstream of the double meander bend the channel morphology changes again to accommodate greater uplift rates. This significantly straighter reach incises a deep gorge in the hanging wall of the MUT (Figure 5-4). The long river profile (Figure 5-2B) suggests that the river in this rapidly incising reach, and in the western half of the basin, is presently in dynamic equilibrium with the uplift.

Downstream of the MDT the river channel widens and the southern bank returns to the slip off type of morphology seen upstream of the MUT. It is in this reach that the anomaly occurs that is interpreted as a parasitic anticline in the Bell Hill Fault footwall syncline. This morphology must be of recent origin. It has certainly developed since the formation of a wide bedrock strath underlying the extensive surface of Terrace A, since which period there is no evidence of significant displacement on the major fault strands. Downcutting of 11 m through Torlesse bedrock has since occurred along the whole length of the river, which is a combination of local effects superimposed on regional uplift or base level fall. Any older events affecting the river should by now have been completely eliminated from the thalweg profile. A small meander has begun to develop across the uplift but no significant incision has occurred at that location, suggesting that the anomaly may be of the order of only very few hundred years old. The river may take some time to adjust. For example, the Mississippi River has yet to adjust fully to the uplift of 190 years ago in the New Madrid Seismic Zone and is still wider and shallower across the uplift (Guccione et al., 2002).

A very similar progression of channel forms can be seen still further downstream in the southeast of the study area and these are similarly located with respect to structure. The broad slip off surfaces are present south of the river as is a steep riser to terrace A immediately north. The river, which has maintained a relatively straight path since entering the gorge, begins to meander again within 2-300 m of the Bell Hill Fault, as it obliquely crosses the rising limb of the footwall syncline. The river has incised through the 2.5 m high upstream facing scarp of the Bell Hill Fault, which displaces the A surface at the exit of the river from the Upper Hawkins Basin.
Just as the modern river reflects modern activity on the faults and folds, the drainage history of the uplifted wedge between the MUT and MDT records a long history of uplift. The height of the MUT fault scarp increases to the northeast with the age of the terraces and the emergent scarp is cut in at least 3 places by significant incised drainage channels other than the modern north or south branches. These channels, labelled from east to west as ch1, ch1a and ch2 (Map 1), record a history of episodic downcutting and drainage disruption that is not seen in the only slightly perturbed, mainly aggradational, terraces on the footwall. The uplift and incision of the wedge, particularly during and post the river occupation of the B surface, therefore directly records a local tectonic signal governed by the magnitude and frequency of local co-seismic ground-rupture events. This coseismic signal is almost certainly overprinted by folding on the anticline but, if the MDT has been operating over the whole period of the terrace record in the uplifted wedge, the influence of folding on the relative terrace elevations is probably limited because, even at modern low flows, the river rapidly trims down to grade equilibrium. The spatial and temporal relationships between the various surfaces on the wedge, and between the wedge and its footwalls, therefore provide an opportunity to establish the paleoseismic and kinematic history of this fault by direct measurement of coseismic slip and by constraining the ages of the various surfaces and by direct measurement of rotations.

Whilst the influence of folding on relative terrace elevations may be limited, it has clearly had a marked effect on the channel morphology, particularly in the gorge. Given the south branch's present dynamic equilibrium and its 11 m of post-terrace A downcutting, the narrowing and steepening of the gorge in response to only 2.5 m of additional local folding in the pop-up wedge is perhaps a little surprising. It may be that the combined uplift, coupled with the decreased erodibility in the area of the gorge shown by the higher S-wave velocities, pushes the river across a critical geomorphic threshold above which planation is not possible at the present stream power. The north branch of the river is responding to the same base level and uplift signals and is not in dynamic equilibrium, having developed a marked convexity (see for example Figure 5-1) that is exaggerated by its sudden fall to join the local base level of the south branch.

5.3.2 Coseismic slip estimation

The reliable identification of individual events and measurement of their coseismic slip is an important factor in determining paleomagnitudes, slip rates and recurrence intervals.
The major surfaces upstream of the wedge were progressively abandoned following the climate driven aggradation events to which they were responding. Their abandonment is a complex downcutting response to glacio-climatic and tectonic lowering of regional baselevel. They are unlikely to have been abandoned due to local neotectonics because such activity would have dammed the river and is more likely to have caused revisitation of previously abandoned surfaces. They are also free of intermediate complex-response degradational events, quite the opposite of what is observed on the hanging wall. This situation allows one to reliably infer from the footwall the inter-surface elevation difference attributable to out-of-basin forcings. The remainder of any elevation difference seen in the wedge must have accumulated by tectonic displacement.

The reader is now advised to open Map 1 and pull out Figure 4-11 (profile locations – p. 52) and Figure 5-5 (p. 89) for ease of reference whilst reading this section.

The influence of original gradient and apparent rotation of the wedge block made measurement of the individual downcutting events difficult. To simplify matters a plane was constructed on the most extensive surface (surface B3, Figure 5-5) and rotated to place the surface horizontal. The elevation differences between each surface were then measured perpendicular to the B3 plane. Datum surfaces are numbered alphanumerically, beginning with the parent terrace identifier and followed by a number where 0 is the peak-aggradation parent surface and n represents the nth increment of downcutting during occupation of the downthrown parent surface. As it is the risers that record events, the base of any channel is also recorded as a surface.

An early challenge in this exercise was to identify the original correlative surfaces within the wedge for the footwall aggradation terraces (B0 and C0). This exercise is complicated by the similarity of the loess sheet on the C-B slip-off surface with the loess profiles on the B3 surface and the downthrown B0 surface. The loess on these surfaces is all of similar stratigraphy and thickness, suggesting no great age difference since the onset of loess accumulation. The downthrown slip off surface is clearly higher than any projection of the downthrown B0 surface but not as high as the C0 surface. Its gradient suggests it is a relic slip off surface between C and B and the measured slip off is within a metre of the C-B downcutting measured in the west of the basin. The upthrown C-B slip off surface has a very similar gradient but records a smaller amount of slip off. The back of the terrace is only 5 m above the highest point on the downthrown side, well less than the total uplift recorded in the B surfaces. It can therefore only be loosely correlated within
the time period represented by the downthrown slip off surface. As shown on Figure 5-5, there is also almost certainly uplift recorded within the C-B interval so the exact elevation of the upthrown B0 surface cannot be calculated based on the data. It seems likely, however, that it is at a higher elevation than the mapped B1 surface.

Accurate determination of coseismic slip is further complicated at Dalethorpe by what appears to be concurrent activity on two bounding faults. It was not possible to distinguish events on the wedge-bounding faults from each other so uplift increments presented here refer to wedge uplift on either one or both of the bounding faults. Some uncertainty is also introduced by the preservation of only one extensive surface (designated B3) and several smaller scraps of surfaces within the wedge. Although at first sight the extent of the B3 surface might be interpreted in terms of recurrence interval as reflecting a period of quiescence, the broad swath over which the surface is planed off suggests that the full flow of the Hawkins River was operating across the area, whilst the small channels (ch1a, ch2) were cut by a much lower flow. Because the younger A surface is similar in extent to the B surface, this can only mean that between events 3 and 4 the river had relocated southward to a lower elevation, probably during the process of incising down to the strath below the aggradational surface A. This would have left the smaller channels draining only the small hills to the north and northwest, much like the present day, and controlled at the downstream end by the progressive downcutting of the main river to the level of surface A. The reduction in streampower across the wedge that accompanied this shift has contributed to improved preservation of the record of individual events.

This observation opens the door to inferring further, unrecorded events prior to the cutting of the B3 surface. The likely response of the aggrading ancestral Hawkins River to episodic scarp emergence across its alluvial plain would have been to pond behind the scarp, depositing sediment and quickly aggrading to the level of the scarp. This would result in the river being perched significantly above the correlative downstream surface, which together with the abrupt upstream aggradation would increase the streampower of the flow across the MDT, the downstream boundary of the wedge. In this situation, the reconfigured low-gradient meandering river crossing the wedge would immediately begin mining its banks and depositing the sediment downstream in order to re-attain dynamic equilibrium with the regional aggradation event. Modelling suggests that dynamic equilibrium could rapidly be regained for a river encountering a fault of this type in its alluvial plain (Ouchi, 1985), leaving the river free to bevel a cut-fill terrace and remove
all traces of the higher cut-fill surface(s). The time to achieve equilibrium would, however, have been greatly increased once bedrock emerged in the wedge. Auger investigations of the B3 surface suggest that it is draped in loess over gravel or weathered Torlesse, whilst augering of the B4 surface shows that it is probably underlain by weathered Torlesse bedrock. On this basis it seems likely that the uplift event that led to the cutting of the B3 surface was the last event before the southwestward propagation of the anticline and the consequent exhumation of bedrock pushed the river further south. There it began incising down to the strath underlying terrace A in response to regional base level lowering.

The ultimate relevance of this discussion lies in the choice of event numbers for calculating likely recurrence interval. The relative elevation changes within the B surfaces are all less than the elevation of the A-B riser on the footwall, so the surfaces can be safely inferred to be coseismic degradational event markers. Only six individual surface remnants are preserved below the C-B slip off surface, each of which probably records an increment of coseismic slip. The first of these (B1) is some 7 m below the C-B slip off surface (Figure 5-5). Surface B1 is seen as a small knob bounded to the northwest by the fault scarp and to the south and east by the broad planar surface of B3. To the west it is bounded by an aigap tributary to ch2, designated ch2a. Ch2a is now well above the level of the swamp and channels ch2 and ch2a are separated by a planar surface (B2) which is itself at a lower level than B1 but above the extensive B3 surface. Given the measured elevation differences between surfaces B1, B2 and B3, it is extremely improbable that the 7 m interval above surface B1 represents a single event. It more probably includes an unknown series of several unpreserved events, all traces of which have been removed due to the efficiency of the river in recycling alluvium from the wedge prior to exhumation of bedrock. The mean increment of coseismic uplift recorded by the five subsequent events is 2.234 m. If this is characteristic, the 7 m interval would yield almost exactly 3 events rather than 1, bringing the total number of events since occupation of the preserved section of C-B slip off surface to eight.

Channel ch2, which occasionally still drains the swamp, is incised below surface B3 and its sides record further increments of downcutting. These are best preserved at the mouth of the channel as remanant surfaces B4 and B5. Both of these surfaces are also preserved in the triangle defined by the MUT fault scarp and channels ch1 and ch1a. Within this triangle the B4 surface can be clearly seen to have been planed off at a lower level than
B3. As seen in Figure 5-5, surface B5 is an airgap partly defined by the highest point of ch2 where it warps over the MUT fault scarp. Therefore, B5’s preservation above the channel floor at the southeastern end of ch2 suggests that the 6th event at least involved simultaneous movement on both wedge-bounding faults. The lack of convergence of surfaces B4 and B5 in ch2 further suggests that this coincidence was characteristic of this system. Ongoing, perhaps simultaneous, movement on both bounding faults goes some way to explaining the paradoxically smaller tilts observed on the higher surfaces of the wedge when compared to more steeply inclined lower surfaces, although the latter are also probably chasing the dropping baselevel at the downstream end (see also section 5.4 below).

Of the 6 preserved surfaces, the lowest (B6) is not a surface as such but rather the present day topography. This event is inferred based on a displacement of the B surface but is assigned within the time marked by occupation of the A surface. Although no obvious active trace of the MUT crosses terrace A immediately north of the river, GPR imaging shows that the last rupture occurred during occupation of terrace A (Corboz, 2004). This is supported by the profile of ch2 (profile 9), which along with chIa clearly grades down from the last MUT rupture to the downstream A surface. This demonstrates that the A surface was indeed already occupied prior to the last rupture. The channels, which by that stage would have been draining the B surface only, rather than the Hawkins catchment, were warped and blocked by the rupture, leading to their virtual abandonment, although ch2 still occasionally carries water. This evidence supports the remapping of the MUT through the river dogleg (Section 5.2), where the rupture could best be disguised, leaving only minor displacements on the flexure slip faults that cut through the A surface and outcrop D1. Because fan and swamp deposits have accumulated against the scarp on the upstream footwall, this last event (B6/A0) was measured relative to a projection of the well preserved footwall B0 surface onto the scarp of the MUT. The measurement yielded a slightly lesser uplift estimate than could be derived by measuring the elevation difference between chI and chIa as they cross the scarp of the MUT.

Table 5-1 summarizes the individual uplift events associated with the event horizons discussed above and illustrated in Figure 5-5. In addition to the coseismic slip events recorded in the terrace sequence, a 2.5 m displacement of the strath underlying surface A has been documented, over and above the regional incision (Figure 4-12). Apart from noting that the warping postdates the cutting of the strath, the timing and rate at which this
strain accumulated is unclear, but nevertheless significant. Ekstrom et al. (1992), for instance, documented a pattern of aftershocks following the 1985 Kettleman Hills, California, earthquake that corresponded with readjustment of the anticline core following coseismic strain. Figure 4-13 suggests that much of the warping documented on terrace A is probably accommodated by shuffling on minor faults. Unfortunately, no fault-normal topographic profiles, that may have indicated the degree of warping of the terrace surface, were run along the A terrace on the south side of the river. However, it is probable that in this instance the 2.5 m up-warping of the strath is produced by a combination of co- and immediately post-last-seismic activity, and pre-next-seismic activity.

<table>
<thead>
<tr>
<th>Event</th>
<th>Occupied surface</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>A</td>
<td>B</td>
</tr>
<tr>
<td>B1 - B2</td>
<td></td>
<td>2.2 m</td>
</tr>
<tr>
<td>B2 - B3</td>
<td></td>
<td>0.95 m</td>
</tr>
<tr>
<td>B3 - B4</td>
<td></td>
<td>2.65 m</td>
</tr>
<tr>
<td>B4 - B5</td>
<td></td>
<td>2.4 m</td>
</tr>
<tr>
<td>B5 - B6/A0</td>
<td></td>
<td>2.97</td>
</tr>
</tbody>
</table>

Unfortunately, the net uplift events recorded as terrace offsets do not appear to provide a realistic direct measure of co-seismic fault-plane slip. A 2.2 m vertical coseismic uplift on a 15° fault, that was purely slip-generated, would require a massive fault-plane displacement of 8.5 m, which is unlikely on this or any other reverse fault; Wells and Coppersmith (Wells and Coppersmith, 1994), for instance, do not speculate on earthquake parameter relationships for reverse displacements >3 m, and even then only for rupture lengths greater than 100 km (see Figure 5-9 in section 5.3.6 below). This raises the question of what constitutes a reasonable partitioning of pre-seismic strain, coseismic folding and coseismic slip for the SFD.
Answering the question above requires assumptions to be made regarding the model of thrust faulting and hence the distribution and relative contributions of the various components of strain. The fault zone structure as summarized in the cross section on Map 1 has already been defended above, so this section addresses the strain partitioning. Assuming that pre-seismic strain is not released at rupture, it is unlikely to account for more than 1/3 of total uplift (Nicol and Campbell, 2001 and references therein) and may be up to an order of magnitude less than co-seismic strain (Ruegg et al., 1982). The pre-seismic component preceding each rupture should be recorded in the apparent uplift ascribed to the previous rupture. The overall slip history in Table 5-1 suggests a characteristic total uplift of ~2.2 m from one event to the next, with the final event recording the largest ‘uplift’ at 2.97 m. This last event is temporally correlated with, and has a similar magnitude to, the 2.5 m apparent warping of the strath. Both must record all three components of uplift mentioned above, assuming that they are all expressed here. Given their magnitudes compared with previous events, it thus seems likely that much of the pre-seismic strain preceding the next event has already accumulated. Assuming this to be the case, and further assuming that pre-seismic strain accounts for a full 1/3 of the total uplift, a pre-seismic strain magnitude of ~0.8<1 m is estimated based on the last event and the warped terrace. This leaves 1.6<2 m of coseismic strain, partitioned between co- or immediately post-seismic folding and co-planar fault slip.

The amount of coseismic deformation expressed as surface fault slip in a thrust earthquake may be a small fraction of the overall deformation. The 1980 El Asnam, Algeria, earthquake for instance was accompanied by intense surface folding and reduced surface rupture compared with calculated fault slip at depth, such that the surface displacement was a maximum of half the total slip inferred from horizontal and vertical geodetic displacements (e.g. Philip and Meghraoui, 1983; Ruegg et al., 1982). Due to a lack of trench data, this study will therefore assume that fault slip accounts for 0.8-1 m of uplift. On a 15° fault this amounts to a coseismic surface displacement estimate of between 1.2 and 1.5 m. Wells and Coppersmith (1994) note that average subsurface displacement typically falls somewhere between the average and maximum surface displacement, however the stacked assumptions required to reach the 1.2 – 1.5 m estimate still render its validity questionable.
Flow-parallel projection of
downthrown B0 surface

Slip off surface

Fault trace

Topographic profile

Direction of view trends +1° to 359°

Figure 5-5: Oblique elevation of topographic profiles (numbers shown) showing coseismic uplift markers at Dalethope. Vertical exaggeration x10, all dimensions in metres. B0 and C0 represent the correlatives of the B and C surfaces on the downthrown side. Surfaces B5-4 and B6/A0 record periods of quiescence and varying degrees of lateral pinionting separating coseismic uplift events that led to abandonment of the previous surface. B5, shown horizontal, is the most extensive and best preserved of these surfaces. For profile locations see Figure 4-11.