CHARACTERIZING ACTIVE FAULTS USING 3-D GROUND-PENETRATING RADAR

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Zusammenfassung


Die Änderung der Strukturen entlang des Streichens, welche während der ersten Studie der "Alpine fault zone" identifiziert wurden, werden zusätzlich mit 2- und 3-D GPR Daten entlang einem ca. 300 m langen Abschnitt der Störzone erfasst. Anhand einer Serie von senkrecht zum Bruch verlaufenden 2-D GPR Profilen kön-


Die Deformation, welche in den GPR Daten der Maleme Verwerfungszone sichtbar sind, besteht aus Falten und zwischen den Brüchen gekippten Blöcken. Die Bestimmung des Anteils dieser Deformation ausserhalb der eigentlichen Brüche an den Messungen der Versetzungsraten erfolgt an einem durch GPR identifizierten Reflexionshorizont, welcher einer 24.8 ± 0.4 cal. kyr BP alten alluvialen Oberfläche...
Abstract

The identification and characterization of active faults are critical for studies of regional seismic hazard. In most countries, many active fault zones have no historical record of surface-rupturing earthquakes, principally because historical and instrumental records of seismicity are much shorter than the average recurrence intervals of large earthquakes. To overcome these shortcomings, paleoseismic data are often used to supplement historical and instrumental records. Unfortunately, the complex shallow subsurface structures of active fault zones are often difficult to characterize using traditional paleoseismic techniques like surface mapping and trenching. Where they can be correlated with geological exposures and trenches, high-resolution 3-D ground penetrating radar (GPR) data can be used to reveal both the complicated patterns of deformation across the fault zone and along-strike variations in fault morphology.

Densely sampled and correctly migrated GPR volumes can provide vivid images of complex fault structures viewed from arbitrary directions. Nevertheless, interpretations of these features are limited by our inability to delineate and describe in a quantitative fashion the typically diverse reflection patterns. Using data acquired across gravel units overlying the Alpine Fault Zone in New Zealand, I demonstrate the utility of various geometric attributes in reducing the subjectivity of 3-D GPR data analysis. I use a coherence-based technique to compute coherency, azimuth, and dip attributes and a gray-level co-occurrence matrix (GLCM) method to compute texture-based energy, entropy, homogeneity, and contrast attributes. A selection of the GPR attribute volumes are used to highlight key aspects of the fault zone and provide information that improves our understanding of gravel deposition and tectonic structures at the study site. I show that, when used in tandem, the coherence- and texture-based attribute volumes can significantly improve the efficiency and quality of 3-D GPR interpretation, especially for complex data collected across active fault zones.

To investigate how the structures identified in the initial Alpine fault zone survey change along strike, I analyze additional 2- and 3-D GPR data that I acquired along a ~300 m section of the fault zone. From a series of fault-perpendicular 2-D GPR profiles, I identify three left-stepping en-echelon fault strands. Two regions of warped strata are interpreted to result from transpressive folding between overlapping strands; these are locations where displacement is transferred from one fault to
the next. The pattern of shallow fault segmentation may also explain the anomalous surface fault displacements reported by previous investigators.

The same geometric attribute techniques that were developed and applied to the Alpine fault zone data set are used to interpret three 3-D GPR data sets collected across three different types of active fault zone in New Zealand that have different deformation styles: the strike-slip Wellington fault zone, reverse faults of the Ostler fault zone, and normal faults of the Maleme fault zone. Using these techniques, I demonstrate how some attributes are more successful at visualizing certain structural or depositional characteristics than others. The GPR data sets and associated attribute volumes show details of near-surface fault geometry that are not obvious from surface mapping. They also reveal evidence for off-fault deformation and gravitational collapse and topple structures.

Active fault slip rates can be difficult to determine using conventional paleo-seismic techniques, especially at locations where deformation extends many meters from the geologically recognized fault trace. In offshore regions, studies based on densely spaced reflection seismic data tied to stratigraphic logs demonstrate that active faults can have variable displacement rates over relatively short distances and short time intervals. I show how high-resolution 3-D ground-penetrating (GPR) data tied to trench-derived stratigraphic logs provide similar information for active faults in onshore regions. To investigate recent ($\leq 24.4$ kyr) fault activity within the Taupo Rift of New Zealand, I analyze a large 3-D GPR data set that I acquired across 10 fault strands within the Maleme fault zone. After correlating three prominent GPR reflection horizons with three faulted chronostratigraphic units observed within a trench, I extrapolate the geometries of the horizons over a $\sim 150 \times 250$ m area of the fault zone and determine slip-accumulation patterns and rates. Average slip rates and slip-rate patterns derived from the GPR data show variability for time intervals $\leq 12.5$ kyr. The results suggest that at least four surface-rupturing earthquakes are required for these faults to exhibit uniform slip rates characteristic of their long-term behavior.

Deformation imaged within the Maleme fault zone GPR data includes folding and tilting of the blocks between the faults. I assess the contribution of off-fault deformation to slip-rate measurements made from a GPR reflection horizon that correlates with a fault-displaced $24.8 \pm 0.4$ cal. kyr BP alluvial surface. By measuring the geometries of the undeformed parts of this reflection horizon several meters from the fault, I show that drag folding and horizontal-axis rotations of the hanging-wall and footwall accommodate $\sim 50\%$ of total extension. I propose that unrecognized off-fault deformation of this kind may explain why geologically determined fault slip rates for the central and southern Taupo Rift are anomalously low when compared to geodetic estimates.
Chapter 1

Introduction

1.1 Paleoseismology and its role in seismic hazard analysis

Paleoseismology is the study of prehistoric earthquakes. More specifically, its goals are to understand how often earthquakes have occurred in the past, when they occurred and their approximate size. The impetus for most paleoseismic studies is society’s need to assess the probability and severity of future earthquakes. Although large magnitude (M > 6) earthquakes occur infrequently, they have produced some of the most devastating natural disasters over the past decade (Table 1.1).

Before 1970, the assessment of earthquake hazard in industrialized countries such as the United States and the USSR was based almost solely on historical earthquake records [McCalpin and Nelson, 1996]. Unfortunately, the relatively short records of historical and instrumental seismicity were insufficient to estimate the probabilities and recurrence frequencies of large earthquakes on a local scale. Allen [1975] was one of the first to point out that by including information on the timing and size of earthquakes interpreted from the late Quaternary geological record,

<table>
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<th>Year</th>
<th>Location</th>
<th>Deaths</th>
<th>Magnitude</th>
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<tr>
<td>2008</td>
<td>Sichuan, China</td>
<td>&gt;69,000</td>
<td>7.9</td>
</tr>
<tr>
<td>2005</td>
<td>Kashmir, Pakistan</td>
<td>~86,000</td>
<td>7.6</td>
</tr>
<tr>
<td>2004</td>
<td>Sumatra, Indonesia</td>
<td>~228,000</td>
<td>9.1</td>
</tr>
<tr>
<td>2003</td>
<td>Bam, Iran</td>
<td>~31,000</td>
<td>6.6</td>
</tr>
<tr>
<td>2001</td>
<td>Gujarat, India</td>
<td>~20,000</td>
<td>7.6</td>
</tr>
<tr>
<td>1999</td>
<td>Izmit, Turkey</td>
<td>~17,000</td>
<td>7.6</td>
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Table 1.1: Selected earthquakes for the past ten years with fatalities >10,000. Adapted from the National Earthquake Information Center, Golden, CO, U.S.A, http://earthquake.usgs.gov/regional/world/world_deaths.php
long-term earthquake statistics for specific fault zones could be extrapolated more reliably into the future. More recently, seismic hazard analyses undertaken in many countries have used paleoseismic data to supplement historical and instrumental records of seismicity [Wesnousky, 1986; McCalpin and Nelson, 1996; Yeats et al., 1997; Stirling et al., 1998; Thenhaus and Campbell, 2003; Pace et al., 2006].

The most commonly used field techniques in paleoseismology are 1) surface mapping of landforms and 2) subsurface mapping of stratigraphy affected by past earthquakes [McCalpin, 1996; Yeats et al., 1997]. Surface mapping yields information on the locations and dimensions of faults. Where faulted geomorphic surfaces can be dated, they can provide information on slip rates, displacements, and constraints on the ages of past deformation events. Subsurface mapping techniques include drilling, trenching and geophysical investigations. By investigating the shallow subsurface of fault zones, information on near-surface fault geometry and deformation attributable to past earthquakes can be derived. Trenches offer the possibility of determining fault history based on stratigraphic relations between faults and late Quaternary sediments. In addition to characterizing subsurface geology across fault zones, geophysical methods are useful for detecting buried faults that have no surface expression. Because small and moderate-sized earthquakes rarely produce significant deformation at or near the Earth’s surface, paleoseismic investigations will usually only yield evidence from large magnitude (M > 6) earthquakes.

1.2 Principles of ground-penetrating radar (GPR)

Ground-penetrating radar (GPR) is a relatively young geophysical technique that has only found widespread application over the past ~20 years (see Annan [2002] for a chronological history). The GPR technique uses electromagnetic fields to detect structures within the subsurface on the basis of their electromagnetic properties. It is analogous to the seismic technique in that it uses reflection and transmission measurements. The key media parameters in GPR surveying are the dielectric permittivity (ε) and electrical conductivity (σ).

1.2.1 Acquisition of GPR data

The most common acquisition configuration is the common-offset surface-to-surface surveying geometry, in which the transmitting and receiving antennas are kept at a fixed distance from each other as they are moved across the ground (Figure 1.1a). Such a configuration is similar to that used in single-channel seismic reflection profiling. At each station location, an electromagnetic pulse is emitted by the transmitting antenna. This energy is reflected at subsurface boundaries of contrasting material properties (ε and σ) and is recorded by the receiving antenna. The recorded trace is a time varying function of the energy returned from different reflecting interfaces within the ground. An example cross-section of a common-offset GPR profile
1.2. Principles of ground-penetrating radar (GPR)

Figure 1.1: Principles of the (a) common-offset and (b) common-midpoint methods

is shown in Figure 1.2a. Important parameters defining a GPR survey are the nominal antenna frequency, the recording time window, the time-sampling interval, the spatial sampling interval, the antenna spacing, and the antenna orientation.

The attenuation of the returned GPR pulse results from a combination of electrical and scattering losses in the subsurface. As a consequence, for surveys over typical geological materials the depth of investigation is limited to a few tens of meters. The resolution of GPR data is controlled by the wavelength of the propagating wave and increases with increasing frequency content. Antennas with nominal center frequencies in the range 25-200 MHz are typically used in geological surveys. Conventionally, GPR data is recorded along profile lines, providing a 2-D representation of the subsurface geology (e.g. Figure 1.2a). Recent improvements in the acquisition of GPR data have made it practical to record dense 3-D surveys such that investigators can now image volumes of the subsurface (e.g. Figure 1.3; Beres et al. [1995]; Lehmann and Green [1999]; Grasmueck and Viggiano [2007]). These types of surveys are usually performed by moving the antennas along parallel, densely spaced lines. To produce reliable images of the subsurface, it is important to determine accurately the positions of the antennas. Several techniques have been developed recently for automatically recording accurate antenna positions as each
Figure 1.2: Examples of data from (a) a common-offset GPR profile across the Alpine fault zone, New Zealand and (b) a nearby common-midpoint profile (CMP), from which the normal-moveout (NMO) of reflections can be used to estimate subsurface velocities.
1.2. Principles of ground-penetrating radar (GPR)

Figure 1.3: An example portion of a processed 3-D GPR volume from Gemmi Pass, western Swiss Alps.

GPR trace is acquired; they include a laser theodolite [Lehmann and Green, 1999], a rotary laser positioning system [Grasmueck and Viggiano, 2007], and differential GPS [Streich et al., 2006].

Because the positions of the transmitting and receiving antennas are fixed for common-offset surveys, additional common-midpoint profiles (CMPs or expanding spread surveys) should be performed to estimate ground velocities (Figure 1.1b). By progressively increasing the offset between the antennas, the normal-moveout (NMO) of subsurface reflections can be used to estimate subsurface velocities (Figure 1.2b).

1.2.2 Processing of GPR data

Because the principles of typical GPR and seismic reflection data acquisition are very similar, many of the GPR processing steps replicate those used in seismic data analysis. Important steps include: 1) time-zero adjustments, 2) signal enhancement, 3) filtering to reduce noise, 4) corrections for topography, and 5) migration. To produce images of the subsurface that can be directly related to subsurface geology, the GPR data should be converted from time to depth using estimates for the subsurface velocities.
1.3 Applications of GPR in paleoseismic studies

A number of investigations have demonstrated the potential for the detection of active faults using 2-D GPR profiles (e.g. Wasatch fault - Smith and Jol [1995]; San Andreas fault - Cai et al. [1996]; Alpine fault - Yetton and Nobes [1998]; Roer Graben - Demanet et al. [2001]; Portland Hills fault - Liberty et al. [2003]). Unfortunately, isolated 2-D profiles may be contaminated by out-of-plane reflections and may not be sufficient for imaging highly heterogeneous structures, including faults distinguished by complicated geometries. In contrast, correctly processed and migrated 3-D GPR data allow the interpreter to observe high resolution images of structures in any direction with approximately equal fidelity.

From data collected across the San Andreas fault in California, Gross et al. [2002, 2003] first demonstrated the potential of 3-D GPR surveying in a paleoseismic investigation. Using sophisticated processing techniques originally developed for 3-D seismic data processing, they successfully imaged two steeply dipping strands and an offset paleochannel; deformation they attributed to the most recent 1906 San Francisco earthquake. A subsequent 3-D GPR data set acquired across the Wellington fault in New Zealand imaged the first GPR fault-plane reflections from an active strike-slip fault [Gross et al., 2004]. Tronicke et al. [2006] imaged faulted strata in a 3-D GPR data set acquired within the Maleme fault zone, New Zealand. By measuring offsets of reflection horizons, they were able to accurately determine recent vertical displacements across several normal fault strands.

1.4 Objectives and structure of this thesis

The main goal of this thesis is to develop new methods for characterizing active faults using GPR. More specifically, my thesis focuses on 1) improving 3-D GPR data processing and interpretation methods, and 2) demonstrating the potential for 2-D and 3-D GPR in paleoseismic investigations.

Attribute techniques developed to improve the interpretation of 3-D reflection seismic data have potential applications to 3-D GPR data. In this thesis, I demonstrate how geometric attributes calculated from 3-D GPR data can be used to improve the interpretation of the near-surface structure of active fault zones. Chapter 2 details the theoretical aspects of two computer codes I wrote to calculate coherence- and texture-based attributes from 3-D GPR data. These codes are used to compute attribute volumes from 3-D GPR data acquired over the active Alpine fault zone in New Zealand. From the various attribute volumes, I identify complex tectonic and depositional structures that are not obvious from inspection of the original data.

In Chapter 3, I present results from an expanded GPR survey across the Alpine fault zone that includes the 3-D GPR survey described in Chapter 2. A series of 2-D GPR profiles covering a large area (~500 x 500 m) of the fault zone reveal en echelon faults separated by zones of transpressive folding that are not evident from
surface mapping. The interpreted structures help explain the anomalous pattern of fault-displaced river terraces that had been mapped by previous investigators.

Chapter 4 is a comparative study of three 3-D GPR data sets collected across different active fault zones in New Zealand that have different deformational styles: the strike-slip Wellington fault zone, reverse faults of the Ostler fault zone and normal faults of the Maleme fault zone. To improve our interpretations, the processed GPR volumes are transformed to coherence- and texture-based attributes using the methods outlined in Chapter 2. In addition, by using complementary geomorphological and geological observations, this study demonstrates how 3-D GPR surveys can contribute valuable information to paleoseismic investigations.

In Chapter 5, I investigate fault displacement patterns and slip-rate variability along an array of normal faults within the Maleme fault zone, New Zealand using 3-D GPR tied to stratigraphic logs from a paleoseismic trench. By correlating the GPR images with three syn-faulted chronostratigraphic horizons observed in the trench, the geometries of the horizons are extrapolated over a ~150 x 250 m area of the fault zone. Displacements of the reflection horizons are measured along ten strands to determine recent (<24.4 kyr) slip-accumulation patterns and rates.

In Chapter 6, I use a portion of the Maleme fault zone data set to assess the contribution of off-fault deformation to fault displacement. The lowest reflection horizon imaged by the data correlates with a fluvial surface that was originally continuous and horizontal, but has been offset, folded, and tilted by faulting. By projecting the planar parts of this reflection horizon on to the fault plane, the total displacement field across the fault is measured.

Finally, in Chapter 7 I review briefly some general conclusions of my work and outline some topics worth further investigation.
Chapter 2

Visualization of active faults using geometric attributes of 3D GPR data: An example from the Alpine fault zone, New Zealand

Alastair F. McClymont, Alan G. Green, Rita Streich, Heinrich Horstmeyer, Jens Tronicke, David C. Nobes, Jarg Pettinga, Jocelyn Campbell, and Robert Langridge

*Geophysics*, 73, B11-B23.

2.1 Abstract

Three-dimensional ground-penetrating radar (GPR) data are routinely acquired for diverse geological, hydrogeological, archeological, and civil engineering purposes. Interpretations of these data are invariably based on subjective analyses of reflection patterns. Such analyses are heavily dependent on interpreter expertise and experience. Using data acquired across gravel units overlying the Alpine fault zone in New Zealand, we demonstrate the utility of various geometric attributes in reducing the subjectivity of 3-D GPR data analysis. We use a coherence-based technique to compute the coherency, azimuth, and dip attributes and a gray-level co-occurrence matrix (GLCM) method to compute the texture-based energy, entropy, homogeneity, and contrast attributes. A selection of the GPR attribute volumes allows us to highlight key aspects of the fault zone and observe important features not apparent in the standard images. They also provide information that improves our understanding of gravel deposition and tectonic structures at the study site. A new depositional/structural model largely based on the results of our analysis of GPR attributes includes four distinct gravel units deposited in three phases and a well-defined fault trace. This fault trace coincides with a zone of stratal disruption and
shearing bound on one side by upward tilted to synclinally folded stratified gravels and on the other side by moderately dipping stratified alluvial fan gravels that may have been affected by lateral fault drag. When used in tandem, the coherence- and texture-based attribute volumes can significantly improve the efficiency and quality of 3-D GPR interpretation, especially for complex data collected across active fault zones.

2.2 Introduction

Attributes used extensively for the interpretation of 3-D reflection seismic data are relatively uncommon in 3-D ground-penetrating radar (GPR) analyses. Recent advances in the acquisition and processing of 3-D GPR data have created new opportunities for mapping the shallow subsurface [Lehmann and Green, 1999, 2000; Young and Lord, 2002; Streich et al., 2006, 2007; Streich and van der Kruk, 2007; Grasmueck and Viggiano, 2007]. Although correctly migrated GPR data volumes are capable of providing vivid images of complex geological units viewed from arbitrary directions, our knowledge of many subsurface features is limited by our inability to delineate and describe in a quantitative fashion the typically diverse reflection patterns.

Most analyses of 3-D GPR data are based on subjective assessments of reflection pattern character. Such assessments are heavily dependent on interpreter expertise and experience. Over the past few decades, similar issues in reflection seismology have been addressed by the introduction of attributes [Chopra and Marfurt, 2005]. A seismic attribute is a quantitative measure of a seismic characteristic of interest. The extensive use of attributes has improved significantly the quality and efficiency of 3-D seismic interpretation in the petroleum industry.

Most seismic investigations are designed to achieve one or more of the following goals: (1) determine the distribution of key physical properties, (2) map the boundaries and identify the lithologies/facies of geological units, (3) estimate the locations and quantities of fluids, (4) delineate potential fluid-flow pathways, and (5) reconstruct the geological histories of the investigated regions. Careful analysis and interpretation of the processed data are necessary to realize these objectives. To complement the expertise and experience of the interpreter and to increase the efficiency of analyzing large volumes of 3D seismic data, a wide variety of mathematically defined attributes are now available. Attributes can be used to emphasize features distinguished by specified characteristics (e.g., reflection amplitudes, frequencies, dips, azimuths, and continuity, and general texture and form) or to extract information that is not obvious in standard images. Some attributes are determined directly from the processed data, whereas others are based on picked horizons. Combinations of attributes are the basis for automated or semi-automated seismic facies classification schemes.

The use of single-trace seismic attributes increased significantly after the pub-
2.2. Introduction

lication of Taner et al.’s 1979 classic paper on complex-trace analysis, and new multi-trace seismic attributes have been introduced nearly every year since the mid-1980’s (see comprehensive review by Chopra and Marfurt [2005]). Two suites of multi-trace attributes that have proven particularly useful in the analysis and interpretation of 3-D seismic data are the coherence-based coherency, dip, and azimuth attributes [Marfurt et al., 1998] and the texture-based energy, entropy, homogeneity, and contrast attributes (note, that some of these attributes are given different names by different authors; West et al. [2002]; Gao [2003, 2004]; Chopra and Alexeev [2006]).

The goals of GPR investigations are similar to those of seismic studies (points 1 - 5 above), except the key physical properties are electromagnetic rather than elastic/aneleastic and the fluids of interest are shallow water occurrences and contaminants rather than deep oil and gas deposits. Moreover, the form and appearance of typical GPR and reflection seismic data sets are very similar. Yet, attributes based on 3-D GPR data have only been reported in a relatively small number of publications [Grasmueck, 1996; Young et al., 1997; Sénéchal et al., 2000; Corbeanu et al., 2002; Tronicek et al., 2006]. Most of these papers were either concerned with single-trace attributes or multi-trace attributes computed from picked horizons. Young et al. [1997] were the first and only group to present coherence, dip, and azimuth attributes computed directly from 3-D GPR data. Although they successfully imaged the boundaries of a sandstone channel, the shallow ~8 m depth penetration, long ~2 m dominant wavelength, relatively coarse 0.61 x 1.22 m grid spacing, and small number (2626) of traces allowed only limited structural information to be extracted from their data. Other groups of researchers [Rea and Knight, 1998; Tercier et al., 2000; Moysey et al., 2003, 2006; Dafflon et al., 2005] have developed and applied a variety of statistical and texture analysis techniques to 2-D GPR data sets. Quantitative measures of texture require sufficient data samples to represent each facies [Gao, 2003]. Consequently, the resolution of texture images derived from 2-D cross-sections is constrained by the dimensions of the necessarily large analysis windows. This problem is significantly reduced for 3-D data sets as a result of the additional information available from the volume of samples surrounding each sample point. Furthermore, by including information from both the inline and crossline directions, interpretational bias is reduced [Gao, 2003].

We have used complementary information provided by coherence- and texture-based attributes to interpret a relatively large (28304 traces) high-resolution 3-D GPR data set collected across a section of the Alpine fault zone in New Zealand that is rich in structural detail. Our primary geological objective was to map shallow sediment deformation produced by past earthquakes along the fault zone. From the multitude of attribute volumes, we have identified subtle features that would have been missed using traditional approaches to 3-D GPR interpretation.

After reviewing briefly the geology of our study site, we describe the acquisition of the 3-D GPR data set across the Alpine fault zone and the subsequent data
processing. We then outline the essential elements of the two suites of attribute techniques and apply them to the Alpine fault zone data set. Based on their information content, we choose a reduced suite of attribute volumes to help us interpret the data in terms of a simple structural/tectonic model.

### 2.3 Alpine fault zone study site

The Alpine fault is a major continental transform fault zone on the South Island of New Zealand that accommodates relative motion between the Pacific and Australian Plates (Figure 2.1a). Despite average plate-convergence rates of 30-40 mm/yr, the fault has not ruptured during the ~150 years of European settlement in the country [Wells et al., 1999]. Paleoseismic investigations are ongoing in an attempt to understand the potential effects of future earthquakes along the fault [Cooper and Norris, 1995; Yetton, 1998, 2002; Sutherland et al., 2006, 2007].

At our Calf Paddock study site, the Alpine fault zone has a transpressive character, rupturing various late Pleistocene and Holocene fluvial deposits of the Maruia River (Figure 2.1a). A succession of offset terraces and stream channels provide direct evidence for recent vertical and horizontal displacements [Wellman, 1952;
2.4. GPR data

Unfortunately, interpretation of the offsets is complicated by the probable presence of overlapping fault strands [Yetton, 2002]. A wall erected across the principal fault strand in 1962 as a rough gauge of aseismic strain (Figure 2.1b) shows no evidence of disturbance.

An early investigation at Calf Paddock involved the acquisition of gravity and seismic refraction data [Garrick and Hatherton, 1974] and the drilling of a vertical hole 39 m southeast of the monitoring wall center (Figure 2.1b). The hole passed through 30 m of gravels before penetrating basement schist to 83 m depth. Although the hole intersected a steeply dipping crushed zone in schist bedrock, it did not transect the principal fault plane. As a consequence, Garrick and Hatherton [1974] inferred that the fault must dip more than $65^\circ$ to the southeast at this site. Yetton [2002] conducted a paleoseismic study of the site that included the excavation and face-logging of trenches that crossed the projected location of the fault (Figure 2.1b).

2.4 GPR data acquisition and processing

Our 3-D GPR data were recorded with a semi-automated acquisition system that comprised a standard PulseEKKO GPR unit linked to a self-tracking laser theodolite with automatic target recognition capabilities [Lehmann and Green, 1999]. We used a broadside acquisition geometry with 1.0-m-offset 100 MHz antennas to record GPR traces (sampling rate 0.5 ns) at ~0.25 m intervals along sixty-one approximately 116-m-long lines separated by ~0.5 m. The survey was located on a terrace above the paleoseismic trench site and recording lines were oriented perpendicular to and centered about the principal fault trace (Figure 2.1b).

After applying a standard median filter to remove low-frequency system-dependent noise (wow), coordinates were assigned to the individual traces and the airwave arrivals aligned using a crosscorrelation technique. The irregularly spaced traces were then interpolated onto a regular 0.25 x 0.5 m grid using Delaunay triangulation (e.g. Streich et al. [2006]). Division by spatially averaged Hilbert transforms (i.e., the amplitude envelopes; Gross et al. [2004]) of the data and frequency filtering yielded traces in which coherent reflections were observed from near the surface to at least 300 ns (Figure 2.2a; note, that this and most other cross-sections presented in this contribution are plotted at a vertical exaggeration of 3:1; for the time sections, this vertical exaggeration is based on a velocity of 0.08 m/ns).

Semblance analyses of reflections observed on common-midpoint (CMP) data collected at different locations (Figure 2.1b) yielded velocities of 0.08 ± 0.01 m/ns. A strongly dispersive groundwave observed on all CMPs was attributed to a thin low-velocity (~0.07 m/ns) surface soil layer that acted as a waveguide [van der Kruk et al., 2006; Streich et al., 2006]. Diffractions from boulders within this layer exhibited the same dispersive character in the common-offset data (Figure 2.2a). They could not be collapsed using standard migration techniques, but they were well
Figure 2.2: (a) Unmigrated inline GPR cross-section at y = 15 m. Blue arrows indicate a steeply dipping diffraction tail affected by dispersion in an overlying waveguide. (b) As for (a) but after 3-D f-k filtering to remove steeply dipping diffractions. (c) As for (b) after 3-D topographic migration, depth conversion, and f-xy deconvolution. Black arrows delineate the projected location of the Alpine fault zone. Vertical exaggeration is 3:1
2.5. Migrated GPR data

separated from the more shallowly dipping events in the frequency-wavenumber ($f$-$k$) domain. They were effectively removed by 3-D $f$-$k$ filtering (Figure 2.2b).

To produce a reliable structural image of the subsurface it was necessary to migrate the data using an algorithm that accounted for the topographic relief across the survey site [Lehmann and Green, 2000; Heincke et al., 2005]. Of the wide range of tested velocities, the 0.08 m/ns velocity estimate based on the CMP data analyses produced the most coherent migrated images. This value was used for the migration and time-to-depth conversion. Application of a 3 x 3-trace $f$-$xy$ deconvolution filter reduced minor random noise while preserving short wavelength features in the migrated images. Finally, the airwaves and groundwaves were muted (Figure 2.2c).

2.5 Topography and migrated GPR data

The gridded surface elevations determined by the self-tracking laser theodolite illustrate clearly the location of the fault trace and the vertical and lateral offsets of a fluvial terrace (Figure 2.1c). The terrace is vertically offset by $\sim$1.5 m (downthrown to the northwest), and an abandoned stream channel is dextrally displaced by $\sim$10 m (Figure 2.1c). These observations are consistent with the Yetton [2002] geomorphological investigation.

The chair diagram representation of the migrated and time-to-depth converted GPR volume in Figure 2.3 shows a plethora of nearly planar reflections with variable dips and dip directions. The structural contrast generated by the fault is evident from the change in reflection pattern at $x = \sim$55 m, which coincides with the vertical offset of the terrace.

2.6 Selected geometric attributes

Rather than follow the normal procedure of interpreting 3-D GPR reflection facies from multiple vertical and horizontal sections [Beres et al., 1995, 1999; Grasmueck, 1996; Green et al., 2003; Gross et al., 2004], we wish to classify objectively the different reflections on the basis of their 3-D geometric attributes. Numerous geometric attributes have been developed for the analysis of 3-D reflection seismic data [Tuner et al., 1994]. They include coherence-based attributes that are helpful for characterizing structures [Bahorich and Farmer, 1995; Marfurt et al., 1998; Cohen and Coifman, 2002] and texture-based attributes that are valuable for facies discrimination [West et al., 2002; Gao, 2003, 2004; Chopra and Alexeev, 2006]. We have written MATLAB codes for generating coherence- and texture-based attributes from the processed Alpine fault zone GPR data volume.
2.6.1 Calculating coherence-based attributes

Bahorich and Farmer [1995] introduced the coherence cube as a means to locate faults within 3-D reflection seismic data. Coherence is defined by waveform similarity in both the inline and crossline directions. Faults and other discontinuities are distinguished by abrupt truncations of coherence. The original algorithm of Bahorich and Farmer [1995] used normalized cross-correlation to estimate coherence. Marfurt et al. [1998] developed a second generation algorithm using semblance as a coherency measure. Their algorithm provides estimates of the coherency, dip, and azimuth for the most coherent dipping plane within a 3-D analysis window of data surrounding each sample point. Unlike other methods that derive dip and azimuth from picked horizons, these attributes are calculated directly from the data.

Our MATLAB code for estimating coherency, dip, and azimuth from 3-D GPR data is based on the Marfurt et al. [1998] semblance algorithm (Figure 2.4). For the Alpine fault zone data, semblance is calculated for hypothetical planes with ranges of apparent dips $\theta_x$ (inline) and $\theta_y$ (crossline) varying between $-40^\circ$ and $+40^\circ$, the minimum and maximum values corresponding to the steepest dips observed in the topographically migrated data. The coherence of each sample point is defined to be the maximum semblance calculated from all combinations of apparent dips (Figure 2.5). The apparent dip pair with the highest semblance is then used to find the true dip ($\theta$) and azimuth ($\psi$) of the plane as follows:

$$\theta = \tan^{-1} \sqrt{\tan \theta_x^2 + \tan \theta_y^2}$$  (2.1)
2.6. Selected geometric attributes

![Figure 2.4](image)

**Figure 2.4:** Sketches showing the principle of the semblance-based algorithm for calculating coherence-based attributes from 3-D GPR data. (a) Top (x, y) view of 3 x 3 (dark-gray shading) and 5 x 5 (light and dark-gray shading) rectangular-analysis windows centered on an analysis point. (b) A 2-D cross-section (x, t) showing a coherent moderately dipping reflection. The analysis window extends over 5 traces in the x-direction and has a time-width of $\Delta t$. The algorithm searches for the most coherent dip in both the x- and y-directions ($\theta_x$ and $\theta_y$). These apparent dips are used to calculate true dip and azimuth relative to north. (c) An example of a relatively low-coherency shallow dipping reflection.

And

$$\psi = \tan^{-1}\left(\frac{\tan \theta_y}{\tan \theta_x}\right)$$  \hspace{1cm} (2.2)

Examples of applying the semblance algorithm to two reflections observed in the Alpine fault zone GPR data set are shown in Figure 2.5. One reflection is practically horizontal and the other has a noticeable apparent southwesterly dip (in the NW-SE sections shown in many figures, the southwesterly dipping reflections have apparent southeasterly dips).

### 2.6.2 Calculating texture-based attributes

In addition to determining the geometry of individual GPR reflections, we wish to describe and delineate objectively any distinctive reflection patterns associated
Figure 2.5: Coherency as measured by semblance for two typical points within the Alpine fault zone GPR volume. (a) Cross-section extracted from the topographically migrated data. Vertical exaggeration is 3:1. (b) and (c) Volumes of data surrounding the points. (d) and (e) Semblance values determined for 196 combinations of apparent dip (14 x 14 values).
with the various geological units without predefining their boundaries. To achieve this goal, we employ texture analysis techniques originally developed for the automated and semi-automated interpretation of 2-D digital images [Haralick et al., 1973; Weszka et al., 1976]. Such techniques are already applied in the analysis of diverse medical, biological, satellite, airborne, and side-scan sonar images as well as gravity, magnetic, and seismic data. Although the concept of texture in everyday life is intuitively obvious, it is not precisely defined. For many purposes, texture is simply a measure of the visual characteristics of objects or features. As examples, textures may be described as simple or complex, smooth or rough, coarse or fine, linear or curvilinear, parallel or divergent, ordered or non-ordered, flat or undulating, even or uneven etc. For 2-D and 3-D digital data, texture attributes provide quantitative representations of these qualitative characteristics.

Texture attributes computed for a given data point determine the statistical relationships between all data within a predefined 2-D or 3-D analysis window centered on that point. The data are pixels or voxels in 2-D and 3-D digital imaging terminology, respectively, and the texture analysis windows are texels (textural elements). The size of the analysis window is defined according to the scale of textures observed in the data. To convert an entire 2-D or 3-D data set to a suite of texture-attribute planes or volumes, we need to compute texture attributes for all data points.

Following the work of West et al. [2002], Gao [2003, 2004], and Chopra and Alexeev [2006] on the application of texture analysis techniques to 2-D and 3-D seismic data, our algorithm for computing texture-attribute volumes from 3-D GPR data is based on the gray-level co-occurrence matrix (GLCM) method of Haralick et al. [1973], the principles of which are reviewed in section 2.11. The computational process begins by discretizing the GPR amplitudes to a limited number of intensity or gray levels $N_g$. Chopra and Alexeev (2006) have demonstrated that $N_g = 16$ levels are sufficient to obtain meaningful statistics from 3-D seismic data. A GLCM is then created at each data point by determining the spatial distribution and frequency or co-occurrence of intensity values within the respective 3-D analysis window. A range of statistics applied to the GLCMs then defines the attributes that quantitatively describe the textures.

For 3-D reflection seismic or GPR data, the analysis window consists of $N_x \times N_y \times N_z$ data points in the inline, crossline, and vertical directions [Gao, 2003]. After specifying the orientation and distance, each GLCM is obtained by summing the co-occurrences $(i, j)$ throughout the 3-D analysis window. Although GLCMs may be calculated for any 3-D orientation [West et al., 2002], we limit our analyses to the three orthogonal directions $x$, $y$, and $z$ ($t$). Prior to computing the statistical relationships, each GLCM is normalized such that the sum of all elements is equal to one.

Figure 2.6 shows GLCMs computed along the orthogonal directions for two 3-D analysis windows extracted from different parts of the Alpine fault zone GPR volume. To capture the fabric of the reflections, we use $9 \times 7 \times 30$ analysis windows
(i.e., 9 traces in the x-direction, 7 traces in the y-direction, and 30 time samples, equivalent to a volume of 2.0 x 3.0 x 0.6 m). In contrast to analysis windows used to calculate coherence-based attributes, those used here should be large enough to capture repeating amplitude patterns within the data. However, overly large windows will smooth features within the transformed attribute volume. The data are rediscretized to 16 intensity levels from 1 (red) to 16 (blue). One analysis window contains continuous flat reflections (Figures 2.6b, 2.6d, 2.6f, and 2.6h), whereas the other contains discontinuous moderately dipping reflections (Figures 2.6c, 2.6e, 2.6g, and 2.6i).

The different textures in the two analysis windows control the distribution of co-occurrences in the resultant GLCMs. For example, the strong continuous reflections within the first analysis window produce a tight distribution along the diagonals of GLCM \(_x\) and GLCM \(_y\) (Figures 2.6d and 2.6f). In contrast, the dipping and more chaotic reflections within the second analysis window produce a wider distribution about the diagonals (Figures 2.6e and 2.6g).

To derive meaningful texture information from the GLCM representations of our 3-D GPR data, we need to analyze them statistically. Haralick et al. [1973] proposed fourteen different statistical operations for extracting a variety of texture information from the GLCMs. Fortunately, most of the useful texture information in wavefield data is contained in four statistical parameters (West et al. [2002]; Chopra and Alexeev [2006]; see Section 2.11): energy, entropy, homogeneity and contrast (inertia).

Energy supplies information on texture uniformity. It is lowest when all elements of the GLCM are equal and is helpful for emphasizing reflector continuity and geometry. Entropy is a measure of disorder or complexity. It is large for non-uniform textures that generally yield low GLCM values. Although energy and entropy are related, when used together they may draw attention to different features in the data.

Homogeneity yields details on the similarity of pixels and is a useful indicator of overall image smoothness and reflector continuity. It is high for GLCMs that have elements located near the diagonal. Contrast (or inertia) highlights local image variations and differences between adjacent data values. It is high for GLCM elements scattered away from the diagonal. When homogeneity and contrast are used together they provide greater discrimination capabilities than either attribute alone.

From three GLCMs in Figure 2.6 and the four equations 2.3 to 2.6 in Section 2.11, we derive twelve statistical measures for each analysis window. These computations are repeated for every sample point in the original GPR volume, thus yielding a suite of 12 texture attribute volumes.
Figure 2.6: Computations of texture-based attributes for the Alpine fault zone GPR volume. (d) Cross-section extracted from the topographically migrated data. Vertical exaggeration is 3:1. Analysis windows centered on each analysis point are extracted from the 3-D volume. They should be large enough to characterize the texture of each reflection facies. (b) Analysis window extracted from a portion of the GPR volume where relatively flat and continuous reflections have been rediscretized into 16 intensity levels ranging from 1 (red color) to 16 (blue color). (c) Second analysis window extracted from a portion of the volume where the reflections have moderate dips and are only quasi-continuous. (d) and (e) GLCMs calculated along the x-axis of the analysis windows shown in (b) and (c). (f) and (g) As for (d) and (e), but for the y-axis. (h) and (i) As for (d) and (e), but for the time-axis.
2.7 Applications of coherence and texture-based attributes to the Alpine fault zone GPR data

2.7.1 Coherence-based attributes

The size of the coherency analysis window is a controlling factor in the resolution of the attribute volumes [Marfurt et al., 1998]. Depending on the data, overly small windows may yield noisy attribute volumes and overly large windows may result in excessive smoothing and loss of detail. To address this issue, we have tested a range of analysis windows on the migrated Alpine fault zone GPR data. In Figure 2.7 we present the results of three such tests in the form of horizontal slices extracted at 6 m depth from the coherency, azimuth, and dip attribute volumes. The first three attribute slices are based on analysis windows containing 3 x 3 traces (0.5 x 1 m; Figures 2.7b - 2.7d), the second three are based on 5 x 3 traces (1 x 1 m; Figures 2.7e - 2.7g), and the final three are based on 5 x 5 traces (1 x 2 m; Figures 2.7h - 2.7j).

All calculations were performed using a vertical window of 10 ns (equivalent to ~0.4 m in the depth-converted volumes). Samples for which the analysis windows overlap the borders of the GPR volume are excluded from all figures. Although the azimuth and dip attributes can be displayed together in a single figure using hue and intensity to represent azimuth and dip magnitude [Marfurt et al., 1998], we find it easier to interpret the attributes using separate figures. Azimuths and dips for which the coherencies < 0.5 are blank in Figures 2.7c, 2.7d, 2.7f, 2.7g, 2.7i, and 2.7j.

Except for the region between x = 20 m and x = 50 m, the GPR reflections have generally low coherencies (Figures 2.7b, 2.7e, and 2.7h). This is a consequence of the types of sediments being imaged; coarse fluvial gravels are rarely well layered. Azimuths and, to a lesser extent, dips change abruptly at x = 55 m (Figures 2.7c, 2.7d, 2.7f, 2.7g, 2.7i, and 2.7j), coincident with the estimated surface trace of the fault. Reflections southeast of the fault have 210 - 250° azimuths and moderate 20 - 35° dips, whereas those immediately to the northwest have ~340° azimuths and somewhat shallower 10 - 25° dips. Further to the northwest at x = 0 - 30 m, the reflections are mostly subhorizontal. All of this information can be discerned in all three suites of attributes. For the further analysis of our data, we have used the attribute volumes based on the larger asymmetric (1.0 x 2.0 m) window, because their resolution is sufficient and they contain less noise than those based on the smaller asymmetric (0.5 x 1.0 m) and symmetric (1.0 x 1.0 m) windows.

Cross-sections through the attribute volumes also provide useful structural information. Relatively continuous reflections in the original data of Figure 2.8a are emphasized in the coherency section of Figure 2.8b. We interpret two bounding surfaces that separate domains of different reflection coherency and azimuth on either side of the fault. Southeast of the fault, a prominent shallow reflection (s1-bs in Table 2.1 and Figures 2.8b and 2.8c) separates shallow easterly dipping reflections from underlying weakly coherent southwesterly dipping ones. Northwest of the
2.7. Applications of geometric attributes to the Alpine fault zone GPR data

Figure 2.7: (a) Horizontal slice extracted at 6 m depth from the fully processed Alpine fault zone GPR volume. (b) - (d) Coherency, azimuth, and dip at 6 m depth calculated using a 3 x 3 (0.5 x 1 m) trace analysis window. (e) - (g) As for (b) - (d) but for a 5 x 3 (1 x 1 m) trace analysis window. (h) - (j) As for (b) - (d) but for a 5 x 5 (1 x 2 m) trace analysis window. The algorithm performed a dip-search up to a maximum value of ~40°. Azimuths and dips with coherencies < 0.5 are shown blank. Various cross-sections along A-A’ in (a) are shown in Figures 2.8 and 2.12. Arrows identify the projected location of the principal strand of the Alpine fault zone.
fault, there is an onlap relationship between shallow nearly horizontal features and a dipping reflection (s2-bs in Table 2.1 and Figures 2.8b and 2.8c). The azimuth cross-section shows excellent discrimination between the different reflection facies on either side of the fault, demonstrating that the division of facies can be traced to at least 15 m depth below the surface (i.e., f2-tf and f1-af in Figure 2.8c). Complementary structural information supplied by the original 3-D data, the three coherence-based attribute volumes, and the texture-based attributes (see next section) can be interpreted in terms of four reflection facies and two bounding surfaces (Table 2.1).

A selection of horizontal azimuth attribute slices at 2.5, 3.0, 5.0, and 7.0 m depth is sufficient to characterize the geometries of the principal structures recorded in the 3-D GPR data. In Figure 2.9, azimuths with coherencies < 0.5 are shown blank and azimuths with dips < 5° are plotted gray.

At 2.5 m depth, we only see reflections from the topographically higher (up-thrown) block southeast of the fault (Figures 2.9a and 2.9e). Predominantly horizontal reflections (gray in Figure 2.9e) within radar facies f3-fl (Figure 2.8c, Table 2.1) are transected by the northeast- and southwest-facing banks of an ancient stream channel (blues and yellows in Figure 2.8c). Sediments draped over the fault scarp produce the band of pink at x ~55 m. Similar features are observed in the azimuth slice at 3 m depth, except the orientation of the stream channel changes abruptly near x = 70 m (Figure 2.9f). This change is not obvious in the original GPR image in Figure 2.9b.

On the 5.0 m and 7.0 m depth slices, the reflections can be divided into three domains:

- mostly subhorizontal reflections at x = 0 - 25 m,
- northwesterly dipping reflections at x = 25 - 55 m, and
- southwesterly dipping reflections that extend from the fault near x = 55 m to at least the southeastern boundary of the survey.

Close inspection of these deeper slices demonstrates that reflections within a ~20 m wide zone are deflected ~30° from a general southwest to west-southwest orientation as the fault is approached from the southeast. This deflection is illustrated best in the automatically generated rose diagrams of Figure 2.10, which show azimuths within 10-m-wide strips extracted at 7 m depth from the azimuth attribute volume. The ancient stream channel at 3 m depth also appears to be deflected by roughly the same amount (Figure 2.9f). We interpret the deflection of the ancient stream channel and deeper reflections as the result of displacement-induced drag close to the fault. Drag in this instance refers to deflected markers that are convex in the direction of slip (c.f. Reches and Eidelman [1995]; Grasemann et al. [2005]). The channel on the shallowest slice at z = 2.5 m has not been deflected, indicating that it has been significantly less affected by fault displacement(s).
2.7. Applications of geometric attributes to the Alpine fault zone GPR data

Figure 2.8: (a) Cross-section from original GPR volume (for location see Figure 2.7a). (b) - (c) Coherence-based attribute cross-sections calculated using the semblance-based coherency algorithm and 5 x 5 trace analysis windows. (b) Coherency. Two bounding surfaces are identified (s1-bs and s2-bs described in Table 2.1). (c) Azimuth. Unreliable data points with coherencies < 0.5 are shown blank. There are three radar facies (f1-af, f2-tf and f3-fl; see Table 2.1 for details) characterized by distinctive azimuths. A fourth radar facies (f4-fl in Table 2.1) is inferred above the s2-bs boundary. Arrows identify the projected location of the principal strand of the Alpine fault zone. Dotted lines delineate a zone of reduced reflection continuity and steeper dips (best seen on the texture-based attribute in Figure 2.12a). Vertical exaggeration is 3:1.
### Table 2.1: Description and interpretation of radar stratigraphy. Labeling system is after Neal [2004]

<table>
<thead>
<tr>
<th>Radar facies</th>
<th>Description</th>
<th>Geological Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>f1-af</td>
<td>Sub-parallel southwesterly dipping reflections with limited lateral continuity. Reflections have moderate (20° - 35°) dips.</td>
<td>Foreset beds of a prograding alluvial fan. Their orientation indicates a northeasterly source. This deposit was probably formed during a period of late Pleistocene aggradation sourced from the Maruia River catchment [Suggate, 1965].</td>
</tr>
<tr>
<td>f2-tf</td>
<td>Sub-parallel northwesterly dipping reflections that are moderately continuous. Reflections have shallow to moderate (10° - 25°) dips.</td>
<td>Post-seismic alluvium beds deposited in the accommodation space provided by the repeatedly downthrown block northwest of the principal fault trace. The current alignment of the dipping strata with the fault plane suggests they were tilted upward by the fault.</td>
</tr>
<tr>
<td>s1-bs</td>
<td>Laterally continuous undulating reflection.</td>
<td>Erosional surface formed by the Maruia River downcutting into radar facies f1-af on the southeast (hanging wall) side of the fault zone.</td>
</tr>
<tr>
<td>f3-fl</td>
<td>Subhorizontal reflections. that are moderately continuous.</td>
<td>Fluvial deposits of the Maruia River.</td>
</tr>
<tr>
<td>s2-bs</td>
<td>Laterally continuous to discontinuous undulating reflection. Sub-horizontal reflections of radar facies f4-fl onlap this surface</td>
<td>Contact at base of late Holocene river gravels resting unconformably over tilted alluvium beds on the northwest (footwall) side of the fault zone.</td>
</tr>
<tr>
<td>f4-fl</td>
<td>Subhorizontal reflections that are moderately continuous. They onlap radar surface s2-bs.</td>
<td>Late Holocene fluvial deposits of the Maruia River.</td>
</tr>
</tbody>
</table>
2.7. Applications of geometric attributes to the Alpine fault zone GPR data

2.7.2 Texture-based attributes

As for the test examples shown in Figure 2.6, the GLCMs we used are based on 2.0 x 3.0 x 0.6 m analysis windows. Horizontal slices extracted at 6 m depth from the twelve texture attribute volumes are displayed in Figure 2.11; these attribute volumes are generated from GLCMs calculated for adjacent points in the x, y, and z directions. Although the numerical values of the color scales vary for each plot, we only highlight relative variations (i.e., high to low values). Some attributes emphasize certain features better than others. For example, because the dipping reflections northwest of the fault mostly strike in the y direction and those to the southeast strike in other directions, all four attribute volumes based on GLCM_y show excellent discrimination between the different reflection patterns on either side of the fault (Figure 2.11). Reflections northwest of the fault have higher energy, lower entropy, higher homogeneity, and lower contrast than those to the southeast. Similarly, entropy_x and contrast_x based on GLCM_x and homogeneity_z and contrast_z based on GLCM_z provide very good differentiation between the subhorizontal reflections at x = 0 - 30 m and the moderately dipping reflections at x = 30 - 55 m.

Three of the twelve texture attributes are sufficient for objectively delineating the different reflection patterns recorded in the Alpine Fault GPR data: contrast_x, homogeneity_y, and contrast_z. Figure 2.12 shows cross-sections extracted from these attribute volumes along profile A-A’ of Figure 2.7a. Each attribute cross-section is shown as a semi-transparent layer overlying a variable area plot of the original GPR data (positive amplitudes are dark shades). Some regions of low signal-to-noise ratio at depths > 13 m produce anomalously high attribute values.

High contrast_x values occur where reflections show little continuity along the profile. The high values at x = 45 - 60 m outline a distinct zone of reduced reflection continuity across the fault and increased reflection dip on its northwestern side (Figure 2.12a). Assuming that the disrupted reflections are a consequence of fault movement, we interpret this zone as a band of shearing and stratal disruption within a broader zone of shear-induced folding identified in the coherence-based attribute volumes. In addition, moderate and low contrast_x values improve discrimination of radar facies f3-fl from f1-af on the southeastern side of the fault (compare Figures 2.8c and 2.12a).

The homogeneity_y attribute highlights continuous dipping reflections perpendicular to the plane of the cross-section (Figure 2.12b), whereas the contrast_z attribute highlights high-frequency reflections within the cross-section. The location of the principal strand of the Alpine fault zone is well delineated on the homogeneity_y cross-section (Figure 2.12b). Thin subhorizontal reflections of radar facies f3-fl and f4-fl yield high contrast_z values, whereas the underlying dipping reflections produce low values (Figure 2.12c). This attribute differentiates between radar facies f4-fl and f2-tf on the northwestern side of the fault and does a better job of defining radar facies f3-fl on the southeastern side of the fault than the azimuth attribute (Figure 2.8c).
Figure 2.9: Horizontal slices extracted from (a)-(d) the original Alpine Fault volume and (e)-(h) the azimuth volume at depths of 2.5, 3.0, 5.0, and 7.0 m. Azimuths with coherencies < 0.5 are shown blank and azimuths with dips < 5° are plotted grey. Arrows identify the interpreted location of the principal strand of the fault zone.
2.7. Applications of geometric attributes to the Alpine fault zone GPR data

Figure 2.10: Rose diagrams of azimuth values sampled within 10-m-wide fault-parallel corridors from northwest to southeast across the fault zone. Dashed lines define the strike of the fault zone. On the southeastern side of the fault zone, the dominant azimuth changes from WSW to SW at \( x > 75 \) m. The number of values \( n \) used for the azimuth estimates are shown at the bottom of each diagram.

Figure 2.11: Horizontal slices extracted at 6 m depth from the suite of texture-attribute volumes calculated from the Alpine Fault GPR data. Each column shows, in descending order, the attributes calculated for the \( x \)-, \( y \)- and time- (or \( z \)) directions. Dashed lines identify the projected location of the fault zone.
Figure 2.12: Cross-sections of the most useful texture attributes extracted from the respective Alpine fault zone volumes along profile A-A’ (Figure 2.7a). Note the color scale at the base of the figure. The texture attributes are plotted as semi-transparent colors overlying a variable area representation of the original GPR data. Annotations refer to the reflection facies described in the text and Table 2.1. Arrows identify the projected location of the principal strand of the Alpine fault zone. Dotted lines define a zone where reflections have reduced continuity and steeper dips (highlighted by the Contrast attribute in (a)). Vertical exaggeration is 3:1.
2.8 Geological interpretation

Cross-sections extracted from the original 3-D GPR data and two of the most informative attribute volumes are plotted at true scale in Figure 2.13 and a sketch of our geological interpretation is presented in Figure 2.14. The four GPR facies identified within the gravels correspond to three stages of deposition:

- an episode of late Pleistocene aggradation [Suggate, 1965] that involved the progradation of an alluvial fan from northeast to southwest (GPR facies f1-af in Table 2.1);
- a period of post-seismic aggradation that led to a distinct sequence of alluvial sediments accumulating in the accommodation space provided by the repeatedly downthrown block northwest of the principal fault trace (GPR facies f2-tf in Table 2.1);
- Holocene base-level lowering, during which the Maruia River incised into the aggradational valley fill and left a thin veneer of fine- to coarse-grained flood deposits blanketing the degraded river terraces (GPR facies f3-fl and f4-fl in Table 2.1).

Of primary importance is the steep southeasterly dip ($\approx 80^\circ$) of both the principal fault strand and the zone of shearing and stratal disruption across the main fault zone. These structures can be followed from the surface to at least 15 m depth at this location (Figure 2.13). This is generally consistent with the steeply dipping crushed zone encountered in the drill hole and the moderately to steeply southeasterly dipping shears of the main fault zone observed in the nearby 1.0-1.5-m-deep trenches (Figure 2.1b; Yetton [2002]). Similarly steeply dipping expressions of the Alpine fault zone are observed in trenches $\sim$12 km to the southwest and $\sim$45 km to the northeast of our study site [Yetton, 2002]. These observations do not preclude the Alpine fault zone from becoming shallower dipping at greater depth [Norris et al., 1990; Norris and Cooper, 1995; Davey et al., 1995, 1998; van Avendonk et al., 2004].

In the near-surface ($<\sim 15$ m), the Alpine Fault Zone is characterized by a $\sim$50-m-wide zone of deformation that can be partitioned into three overlapping domains. A $\sim$30-m-wide domain that forms the northwest footwall to the main fault zone is characterized by asymmetric synclinal folding, within which we observe upward tilting of older river gravel beds approaching the principal fault strand (Figures 2.8a, 2.8c, 2.13a, 2.13b, and 2.14). A second $\sim$20-m-wide domain that forms the southeast hanging wall is characterized by lateral fault drag of alluvial fan deposits and an ancient river channel that approaches the principal fault strand (Figures 2.9, 2.10, and 2.14). The two domains of folding are divided by a $\sim$15-m-wide steeply southeasterly-southeasterly dipping zone of shearing and stratal disruption.
Figure 2.13: The same cross-sections as shown in Figures 2.8a, 2.8c and 2.12a, but plotted with no vertical exaggeration. The azimuth (b) and contrast (c) sections show that both the principal fault strand (arrows) and the zone of shearing and stratal disruption (defined by dotted lines) dip steeply (80°) to the southeast.
2.9 Conclusions

Our 3-D GPR data have provided significant new information on fault-generated deformation within late Pleistocene and Holocene fluvial deposits overlying the Alpine fault zone in New Zealand. A variety of 3-D coherence- and texture-based attribute representations of the data allowed us to emphasize important characteristics of the fault zone, extract information not obvious in the standard GPR images, classify the depositional facies, and describe key structures.

Reflection coherency, azimuth, and dip attributes were calculated using 1.0 x 2.0 x 0.4 m analysis windows. The coherency attribute yielded only limited new knowledge, but the azimuth and dip attributes enabled us to differentiate between prograding alluvial fan deposits southeast of the fault and tilted alluvium beds to the northwest. In addition, the azimuth attribute helped us map the principle fault strand to ~15 m depth and identify a zone of lateral fault drag on the southeast side of the fault.

A multitude of texture-based attributes, which were generated from gray-level co-occurrence matrices (GLCM) using 2.0 x 3.0 x 0.6 m analysis windows, provided
information overlapping and complementary to that supplied by the coherence-based attributes. For example, a thin layer of shallow-dipping fluvial sediments that was only partially defined by the azimuth attribute was well defined by a combination of $\text{contrast}_x$ and $\text{contrast}_z$ attributes. The texture attributes also highlighted subtle features not recognized in the coherence-based attribute volumes. In particular, high $\text{contrast}_x$ values revealed a 15-m-wide zone of discontinuous moderately-dipping reflections that we interpreted to be a domain of intense shearing centered about the principal fault trace.

Our interpretation of the 3-D GPR data and associated attribute volumes was constrained by information provided by earlier geological, paleoseismological, and geophysical investigations. The resultant depositional / structural model of the Alpine Fault Zone at the Calf Paddock study site included distinct gravel units formed during three phases of deposition and a well-defined principal fault trace. Differential motion generated a band of intense shearing on both sides of the fault, the upward tilting of formerly flat-lying gravels on the northwest side and lateral fault drag on the southeast side.

Many critical details of our analysis strategy and depositional / structural model were fundamentally dependent on the 3-D nature of the Alpine fault zone GPR data set. Based on this experience and experience gained by others in analyzing seismic data, we propose that various statistical, texture, and automated classification techniques recently developed for 2-D GPR data would be even more effective when applied to 3-D GPR data. Nevertheless, we would argue that pragmatic approaches to 3-D GPR data analysis should continue to include a large degree of interpreter quality control in addition to applications of semi-automated or automated classification techniques.

2.10 Acknowledgements

We greatly appreciate the contributions of the ETH-University of Canterbury field crew. This project was supported by grants from the Swiss National Science Foundation and ETH Zurich.

2.11 Appendix A: GLCM method and statistical parameters

2.11.1 GLCM method

The gray-level co-occurrence matrix (GLCM) method is comprehensively described in a number of publications [Haralick et al., 1973; Reed and Hussong, 1989; Gao et al., 1998; West et al., 2002]. Here, we review briefly the method for 2-D digital images using an $N_x \times N_y$ analysis window. The extension to 3-D digital data
sets is relatively straightforward [Gao, 2003, 2004; Chopra and Alexeev, 2006]. A GLCM is defined by the number of times that data with intensity level \( i \) occurs with a specified spatial relationship (i.e., orientation and distance) relative to data with intensity level \( j \), where \( i = 1, 2, 3 \) \( N_g \) and \( j = 1, 2, 3 \) \( N_g \). The \( N_g \times N_g \) matrix \( P \) may be evaluated for arbitrary orientations and distances within the analysis window. According to Reed and Hussong [1989], effective texture analyses can be performed using two GLCMs, one with adjacent data points in the x direction (orientations 0° and 180°) and one with adjacent points in the y direction (orientations 90° and 270°). To illustrate the method, consider the following data that have been discretized to \( N_g = 4 \) intensity levels in an \( N_x = 5 \) by \( N_y = 5 \) analysis window:

\[
\begin{array}{cccc}
3 & 3 & 2 & 2 \\
2 & 1 & 1 & 2 \\
1 & 3 & 4 & 4 \\
1 & 2 & 3 & 2 \\
2 & 2 & 2 & 1 \\
\end{array}
\]

The GLCM for this analysis window evaluated for adjacent data points in the forward and reverse x directions is:

\[
\begin{array}{cccc}
(i,j) & 1 & 2 & 3 & 4 \\
1 & 4 & 6 & 1 & 0 \\
2 & 6 & 6 & 4 & 0 \\
3 & 1 & 4 & 2 & 2 \\
4 & 0 & 0 & 2 & 2 \\
\end{array}
\]

The \((i, j)\) element of this GLCM is the number of times that data points with intensity \( i \) occur adjacent to data points with intensity \( j \). For example, the number 4 at \( i = 2 \) and \( j = 3 \) represents four occurrences of a data value with intensity level 2 adjacent to a data value with intensity level 3 along the x direction. Since the calculations are performed in the forward and reverse directions, adjacent data values with equal intensities (i.e., the diagonal values) count twice and the matrix is symmetric.

### 2.11.2 Statistical parameters

The GLCM-based statistical parameters that we use are:

\[
Energy = \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} P_{i,j}^2, \tag{2.3}
\]

\[
Entropy = \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} P_{i,j} \log P_{i,j}, \tag{2.4}
\]
Homogeneity = \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} \frac{1}{1 + (i - j)^2} p_{i,j}, \quad (2.5)

and

Contrast \ (\text{inertia}) = \sum_{i=1}^{N_g} \sum_{j=1}^{N_g} (i - j)^2 p_{i,j}, \quad (2.6)

where \( p_{i,j} \) is the element in the \( i^{th} \) row of the \( j^{th} \) column of the GLCM.
Chapter 3

High-resolution GPR surveying of the northern Alpine fault, New Zealand

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published in GPR '08, Proceedings of the Twelfth International Conference on Ground-Penetrating Radar, June 16-19, Birmingham, United Kingdom

3.1 Abstract

The Alpine fault is a major continental transform on the South Island of New Zealand that is capable of generating large (>7.5) magnitude earthquakes. Paleoseismic investigations are ongoing in an attempt to explain the potential effects of future earthquakes along the fault. We have acquired high-resolution 2- and 3-D GPR data over a large area (~500 x 500 m) that straddles a northern part of the fault zone where surface mapping of a series of faulted river terraces and channels reveals a complicated earthquake history. Our data reveal three left-stepping enechelon fault strands. In addition, two regions of warped strata are interpreted to result from transpressive folding between overlapping strands, such that displacement is transferred from one fault to the next. This pattern of shallow fault segmentation helps explain the anomalous surface fault displacements reported by previous investigators.
3.2 Introduction

The Alpine fault zone is a major dextral transform on the South Island of New Zealand (Figure 5.1a). It extends for at least 800 km and accommodates around two-thirds of the 30-40 mm/year of relative convergence between the Pacific and Australian Plates [Sutherland et al., 2007]. Because the fault has not ruptured during the past ~150 years of European settlement in New Zealand, most of our present understanding of recurrence intervals and potential damaging effects of future major (>magnitude 7.5) ruptures is based on paleoseismic investigations [Cooper and Norris, 1995; Sutherland et al., 2006]. At a few locations, fluvial terrace sequences are offset by the fault, thus providing a potentially useful late Pleistocene - Holocene chronology of fault movement. At such locations it is important to investigate the fault zone in sufficient detail to confidently identify the youngest strands [Yetton, 2002].

At our Calf Paddock study site, the Alpine fault zone has a transpressive character, rupturing various late Pleistocene and Holocene fluvial deposits of the Maruia River (Figure 3.1b). A succession of terraces and stream channels that are offset across the principal fault strand (s1) provide direct evidence for recent accumulated vertical and horizontal displacements (Figure 3.1b). Elevation profiles measured along the locally upthrown and downthrown sides of the principal fault strand (s1) show that vertical and horizontal displacements increase in a southwesterly direction away from the active bed of the Maruia River (Figures 3.1b and 3.2). Unfortunately, the pattern of offsets measured from the oldest terraces is not well understood [Yetton, 2002]. Vertical and horizontal displacements measured from the oldest terrace tread and riser (TR3) are smaller than those measured on the younger terrace tread and riser (TR2) to the northeast (Figure 3.2). Although this pattern may be explained by a mechanism of displacement transfer to overlapping fault strands, there are no obvious fault scarps that would confirm this interpretation [Yetton, 2002].

A 3-D GPR survey acquired in 2003 across the principal fault strand (3D1 in Figures 3.1b and 3.3), imaged a steeply dipping fault plane (~80°) and revealed evidence for off-fault folding and tilting within a broad ~50-m-wide zone of deformation [McClaymont et al., 2008]. In 2006, we collected additional 2- and 3-D GPR data over a much larger area (~500 x 500 m) to investigate along-strike variations in fault zone structure and attempt to image the postulated overlapping fault strands (Figure 3.3).

3.3 Acquisition and Processing

The new data were acquired semi-automatically using a PulseEKKO GPR unit linked to a GPS receiver. Information from a GPS base station provided the means to obtain highly accurate differential GPS coordinates (± 0.1 m). We collected a
Figure 3.1: (a) New Zealand Plate boundary setting and location of the Calf Paddock survey site (black dot) across the Alpine fault. (b) Topography of the survey area. Geographic projection is New Zealand Map Grid. The x and y axes define the local coordinate directions shown in Figures 3.2, 3.3, 3.4, and 3.5; ‘+’ defines the origin of this coordinate system. Dotted white lines - elevation profiles in Figure 3.2; dashed white lines - principal fault strand s1 and secondary strands s2 and s3 interpreted from the GPR data; dotted black lines - edges of abandoned terrace risers TR1, TR2, and TR3 carved by the Maruia River; solid black lines - fault-perpendicular profiles A-A’, B-B’, C-C’, D-D’, E-E’, F-F’, G-G’, and H-H’ displayed in Figure 4; white rectangle 3D1 - site of a previous 3-D GPR survey [McClymont et al., 2008].
Figure 3.2: (a) Elevation profiles along the locally upthrown (blue) and downthrown (red) sides of the fault (profile locations shown in Figure 3.1b). Vertical exaggeration is 10. Vertical displacement vectors are measured at 4 locations along the fault. Horizontal displacement vectors are based on two offset terrace risers (TR2 and TR3) and the trough of an offset channel (indicated by curved arrows in (b)).

Figure 3.3: Map showing the locations of the 100 MHz profiles (red), 200 MHz profiles (blue) and densely sampled 3-D data sets (green) collected at Calf Paddock. GPR cross-sections of the labeled profiles are shown in Figure 3.5a. Fault strands interpreted from the data are delineated by dashed lines. Geographic projection is New Zealand Map Grid. Origin of this coordinate system is denoted with a ‘+’; the y-axis is approximately parallel to fault strike.
3.4 Interpretation

Cross-sections from a series of fault-perpendicular GPR profiles along a ~300 m length of the fault zone are displayed in Figure 3.4 (locations are shown in Figure 3.1b). Abrupt changes in reflection geometry allow the principal fault strand of the fault zone (s1) to be identified on all cross-sections. On the downthrown (northwest) side of the fault, reflections dip to the northwest, whereas reflections on the

<table>
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<th>Nominal antenna frequency</th>
<th>100 MHz Profiles</th>
<th>200 MHz Profiles</th>
<th>3D1 3-D Survey</th>
<th>3D2 3-D Survey</th>
</tr>
</thead>
<tbody>
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<td>Sampling rate</td>
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<td>0.5 ns</td>
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<tr>
<td>Number of lines</td>
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<td>82</td>
</tr>
<tr>
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<td>0.5 m</td>
<td>0.25 m</td>
</tr>
<tr>
<td>Average step size (inline-dir.)</td>
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<td>0.1 m</td>
<td>0.25 m</td>
<td>0.15 m</td>
</tr>
<tr>
<td>Number of traces</td>
<td>27603</td>
<td>63605</td>
<td>28304</td>
<td>17384</td>
</tr>
</tbody>
</table>

Table 3.1: Acquisition parameters for the GPR data.

series of 2-D profiles using 100 MHz and 200 MHz antennas oriented roughly parallel and perpendicular to fault strike. A small 3-D data set covering an area of ~20 x 30 m was acquired over the site of a previous ~2-m-deep paleoseismic trench (Yetton [2002]; Figure 3.3). Acquisition parameters for the data sets, including the 2003 3D1 survey [McClymont et al., 2008], are summarized in Table 4.1.

After applying a standard median filter to remove low-frequency system-dependent noise (wow), the processed differential GPS coordinates were assigned to the individual traces. Corrections, and any required transformations, together with interpolation and binning were achieved during early stages of the processing [Lehmann and Green, 1999]. The stacked traces were scaled, bandpass filtered, and migrated using a topographic migration code that accounted for the relatively large topographic variations across the fault scarp and terrace risers (Lehmann and Green [2000]; Figure 3.1b). The new 3-D survey and 2-D profiles data were migrated individually using a constant average velocity of 0.075 m/ns that was determined from multiple common-midpoint profiles (CMPs) recorded within the survey area. The different migrated data sets were corrected to a common depth (z) datum by aligning matching reflections at profile tie points. This process ensured that picked reflection horizons could be correlated between the 100 MHz and 200 MHz profiles. Maximum depth uncertainties related to this correction were about ± 0.5 m. Finally, all data were loaded into a commercial seismic interpretation package.
Coherency- and texture-based attributes calculated from our two 3-D surveys allow us to more accurately determine the geometries of the dipping reflections and identify different depositional facies across the fault zone (McClymont et al. [2008]; Figure 3.4). The GPR reflections within the downthrown block are oriented parallel to the strike of the principal fault strand (240°) with 10-25° dips and ~340° azimuths; these reflections define facies f2-tf. In contrast, the majority of reflections within the upthrown block are oriented oblique to fault strike with steeper 20-35° dips and 210-250° azimuths; these reflections define facies f1-af. Based on their reflection geometries, the sedimentary layers within the downthrown block are probably fluvial deposits that were originally horizontal, but have been tilted subsequently by fault movement. In contrast, sediments within the upthrown block were probably deposited within a late Pleistocene alluvial fan that prograded from northeast to southwest. During a more recent stage of Holocene base-level lowering, the Maruia River incised into the aggradational valley fill and left a thin (1-3 m) veneer of fine- to coarse-grained flood deposits that blanketed the degraded river terraces.

In addition to the principal fault strand (s1), we have identified two overlapping subsidiary strands (s2 and s3; Figures 3.1b, 3.3, and 3.4). Although they do not have topographic scarps, they are evident from truncations to otherwise continuous reflections (e.g. the offset horizon in Figure 3.4a) and from abrupt changes in reflection pattern. In profiles A-A’, B-B’, and C-C’, strand s2 marks the division between relatively horizontal reflections to the northwest and dipping reflections from facies f2-tf to the southeast (Figures 3.4a-3.4c). Similarly, in profiles F-F’, G-G’, and H-H’, strand s3 separates facies f2-tf from facies f1-af (Figures 3.4f-3.4h). Fault lineaments (n=174) picked from all cross-sections dip steeply to the southeast, with an average dip of 78 ± 8°.

When plotted in map view the three strands form a distinctive left-stepping en-echelon pattern (Figure 3.1b and 3.3). A bend or stepover to the left on a dextral fault creates a region of local compression that can produce such transpressional structures as pressure ridges, thrust faults, and folds [Weldon et al., 1996]. Based on our analysis of all profiles, we suggest that the domains of tilted layering within facies f2-tf represent transpressive folding between the overlapping strands (Figure 3.4).

To investigate the 3-D geometry of the warped strata, we semi-automatically picked some of the more prominent reflections comprising facies f2-tf (Figures 3.4b, 3.4h, and 3.5a). A 3-D perspective of three of the interpolated picked horizons (H1, H2, and H3) and fault strand lineaments is shown in Figure 3.5b. Horizons H1 and H2 and horizon H3 represent two regions of transpressive folding between the overlapping strands that have similar geometries:

1. Where fault strands s1 and s2 overlap, horizons H1 and H2 have shallow (10-25°) dips to the northwest and strike subparallel to the average strike of the
3.4. Interpretation

Figure 3.4: (This page and overleaf) Cross-sections of selected 100 MHz profiles (locations shown in Figure 3.1b). Interpreted fault strands s1, s2, and s3 are delineated by dotted lines. Radar facies f1-af and f2-tf are described in the text. The x coordinates are relative to the fixed point shown in Figures 1 and 3. The insets in (b) and (h) show zooms of the tilted strata (identified with arrows) that comprise facies f2-tf. TR2 and TR3 indicate terrace risers shown in Figure 3.1b. Vertical exaggeration is 3.
3 GPR Surveying of the Northern Alpine Fault
fault zone (050°). Southwest of the southwestern tip of strand s2, the warped horizon tapers and becomes N-S striking, oblique to the average strike of the fault zone.

2. Where strands s1 and s3 overlap, horizon H3 mostly dips shallowly to the northwest and strikes subparallel to the average strike of the fault zone. Further to the southwest, the warped region tapers and the horizon strike becomes progressively more N-S oriented (Figure 3.5b).

We suggest that the warped regions and the subsidiary fault strands s2 and s3 that are not obvious at the surface, have accommodated significant components of horizontal and vertical displacement within the fault zone. Consequently, displacements measured across the northeastern and southwestern tips of the principal fault strand (s1) will be lower than displacements measured along its central section. Indeed, geomorphic measurements across strand s1 show a trend of diminishing displacement toward the southwestern end of the fault that is broadly consistent with this interpretation (Figure 3.2a). The tapering of the warped regions probably coincides with a change in the concentration of deformation from one fault strand to the next (Figure 3.5b). The northeastern end of strand s1 intersects the active bed of the Maruia River (Figure 3.1b). Consequently, we cannot discriminate between the effects of displacement transfer and recent erosion by the river.

That the off-fault deformation is not apparent from the surface topography has important implications for studies that use reconstructions of offset geomorphic features to derive slip rates. Because strain can be widely distributed in the vicinity of fault tips, geomorphic markers in these regions may experience less displacement than those from a less diffuse region of the fault zone. Consequently, geomorphic reconstructions of past offset may be problematic for localities where the shallow part of the fault zone is highly segmented. As demonstrated at our Calf Paddock site, GPR surveys have the potential to characterize along-strike changes in fault zone morphology and identify sites amenable to geomorphic reconstructions.

3.5 Conclusions

Migrated 2- and 3-D GPR data acquired across a section of the northern Alpine fault zone have imaged subsurface structures that were not evident at the surface. Whereas surface mapping identified a single fault trace, our data revealed a segmented fault zone consisting of three left-stepping en-echelon fault strands. Between the overlapping strands, we interpreted regions of warped strata as transpressive folding. The location of one of these folds coincides with a trend of diminishing displacements measured from offset geomorphic markers across the principal fault trace. By improving our understanding of the shallow fault structure at this location, we have demonstrated the utility of 2- and 3-D GPR surveying in geomorphic investigations of active faults.
Figure 3.5: (a) Perspective view of selected 2-D GPR profiles. Black lines delineate fault lineaments (a-f) picked on each cross-section. Yellow lines are the picked horizons H1, H2, and H3 shown in 3.5b. (b) Perspective view of the picked horizons and picked fault strands s1 (black lines), s2 (red lines), and s3 (blue lines). Letters a-f are the cross-sectional fault lineaments corresponding to those shown in 3.5a. The x and y coordinates are relative to the fixed point shown in Figures 3.1 and 3.3.
3.6  Acknowledgements

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Chapter 4

Characterization of active faults using 3-D GPR data

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submitted to Journal of Geophysical Research

4.1 Abstract

Where they can be correlated with geological exposures and trenches, 3-D Ground-penetrating radar (GPR) data can contribute critical subsurface information to paleoseismic investigations. Because active faults are typically characterized by complicated near-surface structures that vary with the styles of faulting and the types of rock that are ruptured, GPR data can be difficult to interpret. We have acquired 3-D GPR data sets across three active fault zones within New Zealand that have different deformation styles: the strike-slip Wellington fault zone, reverse faults of the Ostler fault zone, and normal faults of the Maleme fault zone. To improve our interpretation of the processed GPR volumes, we employed two suites of geometric attributes. The first suite was computed using a coherence-based algorithm. It provided estimates of the coherency, azimuth, and dip of reflections. The second suite quantified the volumetric textures of reflections, which allowed different reflection facies to be defined objectively. We have demonstrated how some attributes were more successful at visualizing certain structural or depositional characteristics than others. For example, the coherency attribute was an excellent tool for highlighting normal faults within volcanic deposits of the Maleme fault zone, whereas the texture-based attributes were most useful for discriminating between the gravel and metasediment units juxtaposed by the Wellington fault zone. Our GPR data sets and associated attribute volumes showed details of near-surface fault geometry that
were not obvious from surface mapping and they revealed evidence for off-fault deformation, and gravitational collapse and topple structures.

4.2 Introduction

The identification and characterization of active faults are critical for studies of regional seismic hazard (e.g., Thenhaus and Campbell [2003]). Historical and instrumental records of seismicity do not completely characterize the earthquake cycle of many active fault zones, because the records are generally much shorter than the repeat times of the largest earthquakes [McCalpin and Nelson, 1996]. This is a particular problem in New Zealand, where relatively complete records of historic and instrumental seismicity exist for only the past ~150 and ~50 years, respectively, whereas large active faults on the islands are estimated to have recurrence intervals of many hundreds of years [Stirling et al., 1998]. Consequently, paleoseismic studies are important for characterizing the numerous active faults that extend along the length of the country (e.g., Berryman [1980]; Van Dissen et al. [1994]; Van Dissen and Berryman [1996]; Benson et al. [2001]; Villamor and Berryman [2001]; Sutherland et al. [2007]).

Paleoseismic investigations of active faults are primarily based on surface mapping and trench excavations [McCalpin and Nelson, 1996]. In many regions, surface mapping allows the major faults to be identified. However, the surface expressions of faulting are subject to modifications by erosion and burial, such that subtle features of past earthquakes (e.g., distributed faulting at fault bends and stepovers, tension fractures, and folding) may not be evident on standard geological and topographic maps. Trenching can be used to determine the amount of fault-related movement, but a single excavation is unlikely to capture all recent surface faulting events [Michetti et al., 2005]. Three-dimensional (3-D) high resolution ground-penetrating radar (GPR) surveying of active faults provides a non-invasive cost-effective means to complement observations from surface mapping and trenching (e.g., Gross et al. [2002, 2003, 2004]; Green et al. [2003]; Grasmueck and Viggiano [2007]; Tronicke et al. [2006]; Amos et al. [2007]).

Although 3-D GPR does not provide direct evidence of past earthquake parameters, it can be used to extrapolate or interpolate information based on exposures and trenches. It can also be used to optimize the location of trenches that provide critical ground-truth information, including samples that can be dated. Over the past 15 years, 2-D GPR profiling has been used extensively in paleoseismic studies to investigate active faults (e.g., Wasatch fault - Smith and Jol [1995]; Portland Hills Fault - Liberty et al. [2003]). Unfortunately, isolated 2-D profiles may be contaminated by out-of-plane reflections and may not be sufficient for imaging highly heterogeneous structures, including faults distinguished by complicated geometries. In contrast, correctly processed and migrated 3-D GPR data allow the interpreter to observe high resolution images of structures in any direction with approximately
equal fidelity.

We demonstrate here the utility of 3-D GPR data for investigating different types of active fault that have ruptured within different types of near-surface material. To minimize the subjectivity of our interpretations and to extract maximum structural information, we take advantage of attribute volumes computed from the processed GPR volumes. A wide range of mathematically defined attributes are commonly used to aid the interpretation of 3-D seismic data. The form and appearance of typical GPR and reflection seismic data are similar. Yet, attributes based on 3-D GPR data have only been reported in a relatively small number of publications [Grasmueck, 1996; Young et al., 1997; Sénéchal et al., 2000; Corbeanu et al., 2002; Tronicke et al., 2006; McClymont et al., 2008].

Attributes are used to enhance features distinguished by specified characteristics (e.g., reflection amplitudes, frequencies, dips, azimuths, and continuity, and general texture and form) or to reveal information that is not obvious in standard images. Some attributes are calculated explicitly from the processed data, whereas others are based on picked horizons. We employ two independent methods, originally developed for the interpretation of 3-D seismic data, to generate different types of attribute volume. Complete details on these methods and their application to 3-D GPR data recorded across the Alpine fault zone in New Zealand are contained in McClymont et al. [2008]. The first is a coherence-based technique that quantifies the geometry of reflections (i.e., coherency, azimuth and dip; Marfurt et al. [1998]). The second is a texture-based technique that allows different patterns of reflections associated with changes in subsurface geology to be delineated [West et al., 2002; Gao, 2003]. From the multitude of coherence- and texture-based attribute volumes we are able to identify subtle characteristics of active faulting that might be missed using traditional approaches to 3-D GPR interpretation.

We acquired 3-D GPR data sets across three active fault zones in New Zealand (Figure 4.1). The faults sampled in our surveys represent a range of deformation styles: the strike-slip Wellington fault zone, normal faults of the Maleme fault zone, and thrust faults of the Ostler fault zone (Figures 4.1 and 4.2). The processed GPR volumes are rich in detail, illustrating the complex 3-D characteristics of fault zones in the near surface. Specific geometrical characteristics of reflections in each data set (e.g., dip, azimuth, and texture) are enhanced by calculating different geometric attributes. The parameters used to transform the GPR volumes to attribute volumes are determined according to the resolution and types of structure present in the data. We demonstrate the suitability and different types of information provided by each attribute for characterizing subsurface structure at each location.

After reviewing briefly the tectonic setting of New Zealand, we introduce the local geology and hazards/risks associated with the faults at our Wellington, Ostler, and Maleme fault zone survey sites. Relevant geomorphic and trenching data are presented for each site. We then summarize the data acquisition and processing procedures for the three 3-D GPR data sets. Subsequently, we outline the essential
Figure 4.1: New Zealand plate boundary setting and locations of the GPR surveys across the 1) Wellington, 2) Maleme, and 3) Ostler fault zones. Major faults (bold lines) and tectonic domains including the Marlborough fault system (MFS), North Island Dextral Fault Belt (NIDFB), and Taupo Rift Zone (outlined by dotted lines) are annotated.
Figure 4.2: (a) Location of the GPR survey across the Wellington-Hutt Valley segment of the Wellington fault zone (bold line), which traverses the capital city of Wellington. Shaded relief topography and fault trace data were obtained from the GNS Science Active Faults Database (http://data.gns.cri.nz/af/). Rectangles mark the locations of paleoseismic trenches and/or geological exposures. (b) As for (a), but for the Maleme fault zone. Some of the longer fault traces (> 3 km) adjacent to and north of the fault zone are plotted as dotted lines. Segment domains are from Rowland and Sibson [2001]. (c) As for (a), but for the Ostler fault zone. The irregularly shaped white filled body outlines the canals and reservoirs forming the Upper Waitaki Power Development Scheme.
elements of the two attribute techniques and apply them to each data set. Based on their information content, we select a reduced suite of attribute volumes specific to each data set to help us interpret the different fault structures. Finally, on the basis of the GPR, surface geology, trench and exposure data, very preliminary interpretations of the shallow fault structures are presented for each location.

4.3 Tectonic setting

New Zealand straddles the boundary between the obliquely converging Australian and Pacific plates (Figure 4.1). Pacific-Australia convergence decreases southward from ~45 mm/yr at 35°S to ~35 mm/yr at 45°S and acquires a progressively larger margin-parallel component [Wallace et al., 2004b]. Active tectonics on the North Island are dominated by the consequences of westward subduction of the Pacific Plate beneath the Australian Plate at the Hikurangi Subduction Zone (Figure 4.1). Some oblique convergence is accommodated onshore by a combination of right lateral strike-slip faults of the North Island Dextral Fault Belt (NIDFB; Figure 4.1) and rotation of major tectonic blocks [Wallace et al., 2004b]. Back-arc extension within the overriding Australian Plate northwest of the NIDFB is represented by normal faults of the Taupo Rift Zone (Figure 4.1).

West-directed subduction terminates near ~43°S where thick buoyant continental crust of the Pacific Plate intersects the plate margin, giving rise to a zone of oblique continent-continent convergence along the Southern Alps of the South Island. Approximately two thirds of this convergence is accommodated by the right lateral transpressive Alpine fault zone (Norris and Cooper [2001]; Figure 4.1). The remaining deformation is distributed on structures east of the Alpine fault zone, such as the Ostler fault zone [Read, 1984; Davis et al., 2005]. Continent-continent convergence across the Alpine fault zone is linked to the Hikurangi Subduction Zone to the north by the Marlborough fault system (MFS), a series of right lateral strike-slip faults that take on a component of thrust motion as they trend offshore (Figure 4.1).

The Alpine fault zone continues offshore to the south near Fiordland where motion becomes almost pure strike-slip [Hull and Berryman, 1986; Sutherland and Norris, 1995]. Further to the south, Pacific/Australia convergence is increasingly absorbed within the Fiordland Subduction Zone, where the Australian Plate plunges beneath the Pacific Plate (Figure 4.1).

4.3.1 Wellington fault zone setting and survey

The Wellington fault zone is one of a number of major faults forming the NIDFB (Figure 4.1). The Wellington-Hutt Valley (W-HV) segment of the fault zone is ~75 km long and has a variable strike between NNE-SSW and ENE-WSW (Figure 4.2a; Van Dissen et al. [1992]; Langridge et al. [2005]). A future major rupture
of the W-HV segment could have catastrophic implications for the densely populated Wellington region. Although the sense of motion is predominantly strike-slip, variable amounts of vertical displacement have produced prominent scarps along the fault zone, with the sense of vertical motion varying along strike and with time [Begg and Johnston, 2000; Langridge et al., 2005]. The W-HV segment last ruptured ~300-450 cal. years BP. It has an expected recurrence interval of 500-770 years [Berryman, 1990; Van Dissen et al., 1992; Van Dissen and Berryman, 1996]. The average Late Quaternary slip-rate is 6.0-7.6 mm/yr and individual surface rupture events with displacements of between 3.8 and 4.6 m have been observed [Berryman, 1990; Van Dissen and Berryman, 1996].

At our Totara Park study site, the general position of the Wellington fault zone is delineated by a 1.2-2.0-m-high fault scarp (Figures 4.3a and 4.3b). Here, geomorphic observations provide little indication of the subsurface fault geometry. A nearby riverbank exposure (Figure 4.2a) demonstrates that the fault dips steeply (~65°) to the southeast with a horizontally stratified 2-10-m-thick sequence of Holocene alluvial gravels overlying sheared greywacke basement rock and older alluvial gravels (Figure 4.3c). Clasts within the older gravel unit on the northwest side of the fault are imbricated parallel to the fault plane, indicating that, like the greywacke, this unit has been subject to long term shearing over multiple earthquake cycles. The riverbank exposure also reveals a 2-3-m-wide fault-parallel zone of disrupted sediments within the younger gravels on the southeast side of the fault (Figure 4.3c). Unfortunately, slip rates have not been determined at this site, because gravel horizons cannot be dated or correlated across the fault [Berryman, 1990]. A second 8-10-m-deep trench excavated across the fault ~10 km to the northeast of our study site (Figure 4.2a) reveals a similar geological sequence (Figure 4.3d).

The objective of our 3-D GPR survey was to image the subsurface and identify deformation structures that may differ from the nearby riverbank exposure. By constraining along strike variations in fault geometry, we could better estimate the impact of future surface ruptures to the nearby housing community (Figure 4.3a).

### 4.3.2 Maleme fault zone setting and survey

The Taupo Volcanic Zone is a region of back-arc extension on the central North Island that results from westward subduction of the Pacific Plate beneath the Australian Plate. About 75% of the ~10 mm/yr of rift-wide extension is accommodated on steeply dipping (60°-90°) normal faults of the Taupo Rift Zone (Figure 4.1; Villamor and Berryman [2001]; Nicol et al. [2006]). In 1987, the M_L 6.3 Edgecumbe earthquake ruptured multiple fault segments within the onshore part of the rift, causing widespread ground damage that included soil liquefaction, soil and rock slope failures, and localized ground subsidence and cracking [Franks, 1988; Beanland et al., 1989].

The entire rift is partitioned along strike into rift segments. Accommodation
Figure 4.3: (a) Aerial photo showing fault scarp morphology that delineates the Wellington fault zone at our survey site. White rectangle outlines the area of the 3-D GPR survey and arrows indicate sense of strike-slip motion. (b) Ground level view of GPR survey across the Wellington fault zone scarp. Arrow shows location of photograph in (c). (c) Riverbank exposure of the Wellington fault zone. Dashed lines delineate major geologic units and arrows show the principal fault strand. (d) Trench exposure of the Wellington fault zone ~10 km to the northeast of the survey site (Figure 4.2a).
zones transfer displacement between these offset extensional domains by soft-linkage of normal faults [McClay and White, 1995; Rowland and Sibson, 2001]. A major bend in the Taupo Rift Zone takes place in the Okataina area, where fault traces appear to rotate from an ENE-WSW orientation within the Okataina domain in the north to a NE-SW orientation within the Ngakuru Subdomain in the south (Figure 4.2b; Rowland and Sibson [2001]).

The Maleme fault zone forms a dense array of at least 16 discrete fault strands that span an ~2.5 km wide graben at the northern end of the Ngakuru Subdomain (Figures 4.2b and 4.4a; Villamor and Berryman [2001]). Surface and subsurface (GPR) mapping shows that the faults are steeply dipping (~75°), forming a complex pattern of splaying and merging strands [Villamor and Berryman, 2001; Tronicke et al., 2006]. These faults displace a former horizontal surface that developed as a result of sedimentation within a lacustrine environment. Drainage of the lake at ~20 ka resulted in local stream dissection of the surface and deposition of an overlying alluvial unit. The lacustrine and alluvial sediments were subsequently mantled by a series of volcanic tephras originating from nearby and distant volcanic vents; the youngest from the most recent Taupo eruption at 1718 cal. years BP [Villamor and Berryman, 2001]. From scarp height measurements across the Maleme fault zone, Villamor and Berryman [2001] determined a vertical displacement rate of 3.55 ± 0.3 mm/yr over the past ~20,000 yr.

We acquired a 3-D GPR data set across a NW-facing fault scarp at the northern end of the Maleme fault zone that encompasses the site of a paleoseismic trench (Figures 4.2b and 4.4). Our objectives were to correlate the complex fault structures observed on the trench walls with structures observed in the GPR volume and to extrapolate their 3-D geometries beyond the trench. Our correlation of the GPR data with the trench logs is described in section 4.7.3.

### 4.3.3 Ostler fault zone setting and survey

The Ostler fault zone is a N-S trending thrust system on the east flank of the Southern Alps. It can be traced over 50 km, forming a ~3-km-wide zone of low angle reverse faults within the intermontane Mackenzie Basin (Figure 4.2c). Cenozoic sediments within the basin reach a maximum thickness of ~1 km and lie unconformably above greywacke basement [Read, 1984; Blick et al., 1989]. Surface geological mapping and seismic reflection data indicate that the fault was originally a normal fault during the Late Cretaceous-Paleocene and was reactivated as a high angle (50°-60°) thrust fault in the last 2.4 Ma [Ghisetti et al., 2007].

The fault zone has been the subject of a deformation monitoring project since 1966, prior to the construction of the Upper Waitaki Power Development Scheme, which includes a series of canals, dams, and reservoirs located adjacent to and within the fault zone (Figure 4.2c). Continued warping of the ground adjacent to the fault zone suggests a significant component of aseismic deformation [Blick et al.,]
Figure 4.4: (a) Aerial photo showing numerous NE-SW trending fault scarps (arrows identify one of these scarps) and the location of our GPR survey within the Maleme fault zone. Coordinates are New Zealand Map Grid. (b) Ground level view of GPR survey across one of the topographic scarps. Dashed line defines the top of the scarp.
4.4 Data acquisition and processing procedures

1989; Van Dissen et al., 1994]. Nevertheless, vertical displacement of surfaces and strata across the fault zone provide strong evidence for coseismic displacement on a kinematically linked fault system [Van Dissen et al., 1994; Davis et al., 2005]. Two trenches excavated across segments of the fault zone to the north (Figure 4.2c) found evidence for large earthquakes that occurred approximately 3600, 6000, and 10,000 cal. years BP [Van Dissen et al., 1994].

Seismic reflection data indicate that the main fault segment dips 50°-60° westward to depths of ~1.5 km [Ghisetti et al., 2007]. By aligning scarps crosscutting multiple fluvial terraces formed by the Ohau River (Figure 4.2c), Davis et al. [2005] measured an average near-surface fault dip of 50 ± 9°. Using this characteristic dip together with vertical offsets observed on terrace surfaces and surface ages estimated by Read [1984], they calculated shortening rates across the Ostler fault zone of 0.7-1.0 mm/yr. From observations of warped and tilted fluvial terraces at the same location, Amos et al. [2007] postulated that the Ostler fault zone is a listric thrust that shallows with depth. Their 3-D GPR data revealed multiple fault planes with an average dip of 56 ± 9°.

Near our Clearburn study site, two main thrust segments are separated by a ~500-m-wide slip transfer zone (Figure 4.2c). At this location, the fault zone cuts Late Pleistocene glacial outwash gravel. Numerous late-glacial to interglacial braided river channels are carved into the outwash surface (Ghisetti et al. [2007]; Figure 4.5a). From geomorphic mapping, Davis et al. [2005] identify 12 small (30-300 m long) subparallel reverse faults in this area, each of which has maximum vertical displacements of 1-3 m. In noting that the cumulative vertical displacement on these faults is anomalously low compared to that measured on arrays of faults to the north and south, they suggest that most of the displacement deficit is accommodated by fault-related folding on a large hanging wall anticline that warps the outwash surface west of the slip transfer zone (Figure 4.5a).

To investigate the details of small scale faults that do not form obvious scarps, we acquired a 3-D GPR data set across the toe of the Ostler fault zone deformation front, on the eastern side of the slip transfer zone (Figures 4.2c and 4.5).

4.4 Data acquisition and processing procedures

All of our data sets were collected using a Sensors and Software PulseEKKO GPR unit integrated with a self-tracking laser theodolite [Lehmann and Green, 1999]. Acquisition and processing parameters are summarized in Tables 4.1 and 4.2, respectively. The GPR antennas were mounted on a sled. A target prism fixed to a mast on the sled was tracked by the theodolite and coordinates measured as the GPR data were acquired. The traces were recorded at a relatively constant rate (<1 s per trace) along approximately parallel straight lines (i.e., along the x-direction of each survey; Figure 4.6). Coordinate assignments, corrections, and any required transformations, together with interpolation and binning were achieved during early
Figure 4.5: (a) Orthorectified photo of the Ostler fault zone at Clearburn and the location of our GPR survey. Triangles are on hanging wall side of fault zone. West of the main deformation front, an anticlinal bulge warps the surface of the glacial outwash gravels. A network of braided paleochannels (facing arrows identify one of these channels) is etched into this surface on both sides of the fault zone. Coordinates are New Zealand Map Grid. (b) Ground level view across the deformation front (dashed line) of the Ostler fault zone.
4.5. Transformation to geometric attributes

Table 4.1: Acquisition parameters for each 3-D GPR survey

<table>
<thead>
<tr>
<th></th>
<th>Wellington FZ</th>
<th>Maleme FZ</th>
<th>Ostler FZ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nominal antenna frequency</td>
<td>100 MHz</td>
<td>100 MHz</td>
<td>100 MHz</td>
</tr>
<tr>
<td>Antenna separation</td>
<td>1.0 m</td>
<td>1.0 m</td>
<td>1.0 m</td>
</tr>
<tr>
<td>Antenna configuration</td>
<td>Perpendicular-broadside</td>
<td>Perpendicular-broadside</td>
<td>Perpendicular-broadside</td>
</tr>
<tr>
<td>Sampling rate</td>
<td>1.0 ns</td>
<td>0.5 ns</td>
<td>0.4 ns</td>
</tr>
<tr>
<td>Trace length</td>
<td>750 ns</td>
<td>460 ns</td>
<td>500 ns</td>
</tr>
<tr>
<td>Number of traces</td>
<td>10251</td>
<td>32040</td>
<td>51675</td>
</tr>
<tr>
<td>Number of lines</td>
<td>51</td>
<td>120</td>
<td>159</td>
</tr>
<tr>
<td>Average step size (x-dir.)</td>
<td>~0.25 m</td>
<td>~0.15 m</td>
<td>~0.25 m</td>
</tr>
<tr>
<td>Survey dimensions</td>
<td>50.0 x 25.0 m</td>
<td>40.0 x 30.0 m</td>
<td>81.0 x 39.5 m</td>
</tr>
<tr>
<td>Average velocity</td>
<td>0.10 m/ns</td>
<td>0.07 m/ns</td>
<td>0.11 m/ns</td>
</tr>
</tbody>
</table>

stages of the processing (Table 4.2; Lehmann and Green [1999]).

Portions of the data from each processed GPR volume are displayed in Figure 4.6. Favorable ground conditions at each site allowed us to record high resolution data to depths >10 m. All data sets were migrated using a constant average velocity determined from multiple common-midpoint profiles (CMPs) recorded within the survey areas (Table 4.1). A 3-D phase-shift code was used to migrate the Wellington fault zone data volume [Yilmaz, 2001]. Relatively large topographic variations within the Ostler and Maleme fault zones required the application of a topographic migration code [Lehmann and Green, 2000; Heincke et al., 2005]. The constant velocity used to migrate each data volume was used to convert the respective migrated GPR volume from time to depth (Table 4.1). Finally, the coherency of reflections was enhanced and uncorrelated noise reduced by applying a gentle $f$-$xy$ deconvolution filter (3 x 3 traces) to each data set.

4.5 Transformation to geometric attributes

To improve and guide our interpretation of each data set, the processed GPR volumes were converted to multiple geometric attribute volumes. We used two independent techniques, originally developed for 3-D reflection seismic data, to generate coherence- and texture-based attribute volumes [Marfurt et al., 1998; Gao, 2003; McClymont et al., 2008]. Coherence-based attributes are useful for characterizing structures like faults, stratigraphic discontinuities, or channels [Bahorich and Farmer, 1995; Marfurt et al., 1998, 1999; Marfurt, 2006], whereas texture-based attributes are helpful for facies discrimination [West et al., 2002; Gao, 2003, 2004;
Figure 4.6: Portions of processed and migrated 3-D GPR volumes from the (a) Wellington fault zone, (b) Maleme fault zone, and (c) Ostler fault zone surveys. Top mutes mask the airwaves and groundwaves. Blue lines are the ground surfaces. (a) Radar facies B basement of metasediments and older deformed gravel; G older gravel sequence; S Holocene gravel sheets. No vertical exaggeration. (b) MR1, MR2, and MR3 are reflections associated with boundaries between different layers of tephra and paleosol that can be traced across the normal faults. Dashed rectangle defines the walls of a former paleoseismic trench. Vertical exaggeration of 2. (c) DG - deformed outwash gravels; UG - undeformed outwash gravels. White arrows identify two faults. Vertical exaggeration of 2.
4.5. Transformation to geometric attributes

<table>
<thead>
<tr>
<th>Processing Step</th>
<th>Wellington FZ</th>
<th>Maleme FZ</th>
<th>Ostler FZ</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Assignment of coordinates to the traces</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.</td>
<td>Median (dewow) filter</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3.</td>
<td>Trace editing</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.</td>
<td>Alignment of first arrivals</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5.</td>
<td>Interpolation and trace binning</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6. Grid size</td>
<td>Inline (x-dir.) 0.25 m</td>
<td>0.15 m</td>
<td>0.25 m</td>
</tr>
<tr>
<td></td>
<td>Crossline (y-dir.) 0.5 m</td>
<td>0.25 m</td>
<td>0.25 m</td>
</tr>
<tr>
<td>7.</td>
<td>Amplitude scaling</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8.</td>
<td>Bandpass filtering</td>
<td></td>
<td></td>
</tr>
<tr>
<td>9. 3-D Migration</td>
<td>phase shift</td>
<td>topographic</td>
<td>topographic</td>
</tr>
<tr>
<td>10.</td>
<td>Time to depth conversion</td>
<td></td>
<td></td>
</tr>
<tr>
<td>11.</td>
<td>3 x 3 trace f-xy deconvolution filtering</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12.</td>
<td>Muting of air- and ground-waves</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4.2: Processing parameters for each 3-D GPR survey

Chopra and Alexeev, 2006]. We have written two MATLAB codes to generate the different suites of attributes.

4.5.1 Calculating coherence-based attributes

Coherency is a measure of reflection continuity that can be used to determine the dominant dip and azimuth of reflections. Unlike most methods that provide dip and azimuth estimates using picked horizons, coherence-based attributes are calculated directly from the data. Our algorithm performs scans over a range of dips in the x- and y-directions to find the most coherent planar reflections within 3-D analysis windows of data surrounding each sample point [Marfurt et al., 1998; McClymont et al., 2008]. The sizes of the analysis windows are dependent on the reflection characteristics. If the analysis windows are too small, the attribute volume may appear noisy. Overly large windows may result in excessive smoothing and loss of detail [McClymont et al., 2008]. Semblance values are calculated for a range of apparent dips in the x- and y-directions over a vertical time window. The size of the time window is usually chosen to be large enough to encompass a full cycle of the dominant signal wavelength, but it can be shortened or lengthened depending on the interpretation objective. For example, shorter time windows will highlight short duration low coherency features like channel edges, whereas longer windows will enhance near-vertical low coherency structures like faults [Marfurt et al., 1998]. The maximum dip tested should be greater than the steepest dipping reflections observed in the data.

We define the coherency at each sample point as the maximum semblance com-
putered from the dip scan. The apparent dip pair \((\theta_x, \theta_y)\) that yields the maximum semblance is used to calculate the true dip \(\theta\) and azimuth \(\psi\) for each sample point [Marfurt et al., 1998; McClymont et al., 2008]. This process is repeated for all sample points until the original GPR volume is transformed into coherency, dip and azimuth attribute volumes.

4.5.2 Calculating texture-based attributes

Whereas coherence-based attributes are excellent tools for structural characterization of a GPR volume, texture-based attributes provide information useful for radar facies discrimination [McClymont et al., 2008]. In general terms, texture is simply a measure of the visual characteristics of objects or features (e.g., smooth or rough, linear or curvilinear, parallel or divergent). For 2-D and 3-D digital data, texture attributes provide quantitative representations of these qualitative characteristics. Like the 3-D coherence-based attributes, volume texture attributes are computed for predefined 3-D analysis windows centered about each data point. For 3-D reflection seismic or GPR data, each analysis window consists of \(N_x \times N_y \times N_z\) data points in the inline, crossline, and vertical directions [Gao, 2003]. It should be large enough to capture the repeating patterns that represent a particular texture. The data points are voxels in 3-D digital imaging terminology and the texture analysis windows are texels (textural elements). To convert an entire 3-D data set to a suite of texture-attribute volumes, we compute texture attributes for all data points.

Our method of calculating texture attributes is based on the gray-level co-occurrence matrix (GLCM) method of Haralick et al. [1973], as modified for 3-D data by West et al. [2002], Gao [2003], Gao [2004], and Chopra and Alexeev [2006]. For volume texture analysis, a GLCM is a symmetric matrix that quantitatively describes the spatial relationships and relative occurrences of reflection amplitudes within a 3-D analysis window. It has a dimension of \(N_g \times N_g\), where \(N_g\) is the number of gray levels or intensities used to quantify the GPR amplitudes. Our GPR data were originally sampled with 16-bit precision \((N_g = 65536)\). By resampling the data to 4-bit precision \((N_g = 16)\), we reduce the dimensions of the GLCM and the number of computations required for each data point without significantly altering the texture quality [Chopra and Alexeev, 2006; McClymont et al., 2008].

A GLCM is created for each analysis window by counting the number of co-occurrences of a particular spatial arrangement (orientation and distance) of reflection amplitudes. For example, we could generate a GLCM in which each element \((i, j)\) of the matrix represents the number of times that a data point with amplitude \(i\) is adjacent to a data point with amplitude \(j\) along the \(x\)-direction of the analysis window. Although GLCMs may be calculated for any 3-D orientation [West et al., 2002], we limit our analyses to the three orthogonal directions \(x, y,\) and \(z\) \((t)\). Consequently, at each data point we generate three GLCMs: \(\text{GLCM}_x, \text{GLCM}_y,\) and \(\text{GLCM}_z\).

The distribution of values within a GLCM varies according to the texture of
reflections within the analysis window. For strong continuous reflections, a GLCM calculated for a direction within the plane of the reflections (i.e., x- and y-directions for horizontal reflections) will have high values along the diagonal. Discontinuous or incoherent reflections will have more occurrences of high values farther from the diagonal [McClymont et al., 2008]. Rather than evaluate the individual GLCMs in a qualitative fashion, we quantify statistically the distribution of co-occurrences to obtain measures of texture. Haralick et al. [1973] derived 14 different statistical measures of texture from GLCMs. We use four of these to quantify the distribution within each GLCM [West et al., 2002; Chopra and Alexeev, 2006; McClymont et al., 2008]: energy, entropy, homogeneity, and contrast (inertia).

Energy is a measure of textural uniformity within an image. It is lowest when all elements of the GLCM are equal and is useful for emphasizing reflection continuity and geometry. Entropy is a measure of disorder or complexity and is high for non-uniform textures that typically yield low GLCM values. Homogeneity measures the similarity of voxels within an analysis window. It is a useful indicator of overall image smoothness and reflection continuity. It is high for GLCMs that have elements concentrated near the diagonal. Contrast (or inertia) highlights local image variations and differences between adjacent voxels. It is high when the elements of the GLCM are scattered away from the diagonal.

4.6 Interpretation of the Wellington fault zone data set

A portion of the processed Wellington fault zone GPR volume is shown as the chair diagram of Figure 4.6a. This data set exhibits a complex pattern of dipping reflections with variable continuity. After analyzing individual cross-sections and horizontal slices, Gross et al. [2004] were able to identify three distinct radar facies (Figure 4.6a). Two of these facies can be correlated with major geological units seen at the nearby riverbank exposure; S corresponds to the deposit of Holocene gravel sheets and B matches greywacke and older deformed gravels exposed on the NW side of the principal fault trace (Figure 4.3c).

4.6.1 Application of coherence-based attributes to the Wellington fault zone data set

To aid our interpretation, the original GPR volume was first converted to coherency, azimuth and dip attribute volumes. We tested a range of parameters to determine the optimum size of the trace analysis window. Figure 4.7 shows horizontal slices extracted from coherence-based attribute volumes calculated using 3 x 3 (i.e., 3 traces in the x-direction and 3 traces in the y-direction; 0.5 x 1 m) and 5 x 5 (1 x 2 m) trace analysis windows. A vertical window of 10 ns (equivalent to ~0.5 m
in the depth-converted volumes) was used for all calculations. Since cross-sections of the original data revealed some reflections with relatively steep dips, we used a large maximum search dip of $80^\circ$ for the computation of both sets of attributes. For the Wellington fault zone example, the coherency, dip, and azimuth attribute volumes based on the larger analysis window (Figures 4.7c, 4.7e, and 4.7g) provided sufficient resolution while being smoother and easier to interpret than those based on the smaller analysis window (Figures 4.7b, 4.7d, and 4.7f). For this reason, our interpretation of the Wellington fault zone was based on the attribute volumes calculated with the larger analysis window.

Coherency attribute slices highlight the contrast in continuity of reflections across the fault zone (Figure 4.7c). Reflections from the sheared greywacke/gravel basement (B) northwest of the fault have little continuity with low coherency values, whereas reflections from the Holocene gravel sheets (S) southeast of the fault are relatively continuous with moderately high coherency values.

Azimuth and dip attribute slices reveal significant changes in the orientation of reflections across the fault. Figures 4.7e and 4.7g demonstrate that many reflections from the greywacke/gravel basement dip moderately to steeply ($>40^\circ$) either to the northwest (purple) or southeast (green). At the riverbank exposure, shear planes within the greywacke unit and imbricated clasts within the older gravel unit are oriented subparallel to the principal fault strand on its northwest side (Figure 4.3c; Berryman [1990]). We interpret this fabric as the result of extensive fault-parallel shearing within a broad zone next to the steeply dipping fault plane. In contrast, the majority of reflections from the Holocene gravel sheets show little evidence of deformation. They are mostly horizontal, reflecting their recent fluvial origin (Figures 4.7e and 4.7g).

Cross-sections through the original and attribute volumes calculated with the 5 x 5 trace analysis windows allow us to visualize vertical changes in reflection geometry (Figure 4.8). The coherency attribute cross-section shows clearly the difference in thickness of the Holocene gravel sheets on the northwest (up thrown) and southeast (down thrown) sides of the fault (Figure 4.8b). The surface forming the base of the gravel sheets has a vertical separation of 3 m across the fault plane; without a well constrained age for this surface we cannot estimate a vertical slip rate. Unfortunately, laterally offset piercing points are not evident in our data. From geomorphic observations of river terrace risers ~500 m to the northeast of our GPR survey, Berryman [1990] reports horizontal and vertical displacements of 20.0 and 1.4 m, respectively.

The coherency attribute also highlights the steep-dipping reflection that Gross et al. [2004] interpreted as a fault-plane reflection (delineated by arrows). Multiple high coherency fault-parallel reflections on the NW side of the principal fault strand indicate a concentrated 5-m-wide zone of intense shearing within the basement unit (Figure 4.8b). Low coherencies within a 2-3 m wide zone separate the fault plane from the undeformed Holocene gravel sheets on the down-thrown block (DS in
4.6. Interpretation of the Wellington fault zone data set

Figure 4.7: Horizontal slices extracted at 5 m depth from the Wellington fault zone original GPR and coherence-based attribute volumes. (a) Original GPR data. Cross-sections along A-A’ are shown in Figures 4.8 and 4.10. (b) and (c) Coherency calculated using 3 x 3 and 5 x 5 trace analysis windows, respectively. (d) and (e) Azimuth calculated using 3 x 3 and 5 x 5 trace analysis windows, respectively. (f) and (g) Dip calculated using 3 x 3 and 5 x 5 trace analysis windows, respectively. The algorithm performed a dip search up to a maximum value of 80°. Azimuths and dips with coherencies <0.5 are plotted gray and azimuths and dips for near horizontal features are blank. B - basement of metasediments and older deformed gravel; S - Holocene gravel sheets. Arrows identify the interpreted location of the principal fault strand.
Figure 4.8: A-A’ cross-sections extracted from the Wellington fault zone original GPR and coherence-based attribute volumes calculated using 5 x 5 trace analysis windows (see location in Figure 4.7a). (a) Original GPR data. (b) Coherency. (c) Azimuth. (d) Dip. Azimuths and dips with coherencies <0.5 are plotted gray and azimuths and dips for near horizontal features are blank. B - basement of metasediments and older deformed gravel; DS disrupted sediments; G older gravel sequence; S - Holocene gravel sheets. Dashed lines delineate the base of S and DS. Facing arrows show interpreted location of the principal fault strand. Double-headed arrows indicate the approximate width of the fault zone. No vertical exaggeration.
4.6. Interpretation of the Wellington fault zone data set

Figure 4.8b) that formed as a consequence of fault rupture during the Holocene. The combined width of the zone of concentrated shearing and disrupted sediments is 7-8 m, broadly consistent with a fault-zone width of 8-10 m proposed by Gross et al. [2004].

The horizontal slices of Figures 4.7e and 4.7g, combined with the cross-sections of Figure 4.8, demonstrate the horizontal nature of the Holocene gravel sheets. Although the azimuth and dip attributes provide little additional information about the low coherency basement and older gravel sheet reflections, they show quantitatively the geometry of the fault-plane reflections; the relatively continuous reflections dip 70° at an azimuth of ~130° (green in Figure 4.8c; burnt orange in Figure 4.8d).

4.6.2 Application of texture-based attributes to the Wellington fault zone data set

In addition to the three coherence-based attributes, we have calculated 12 texture-based attribute volumes (i.e., energy, entropy, homogeneity, and contrast in the x, y, and z directions) for the Wellington fault zone data set. We used analysis windows with dimensions 7 x 5 x 20 samples (1.5 x 2.0 x 1.0 m) to capture changes in reflection pattern associated with the four radar facies shown in the cross-sections of Figure 4.8.

Horizontal slices extracted at 5 m depth from each of the 12 texture-based attribute volumes are displayed in Figure 4.9. Although the numerical values of the color scales vary for each plot, we only highlight relative variations (i.e., low to high values). Each row corresponds to the three orthogonal directions used to generate GLCMs for each data point and the columns show the four statistical measures used to evaluate each GLCM. Some attributes emphasize certain reflection patterns better than others. For example, the textural differences between the relatively continuous horizontal reflections from the Holocene gravels and the semi-continuous steeply dipping reflections from the greywacke/gravel basement across the fault are well defined in the entropy, homogeneity, and contrast attribute volumes calculated from GLCM$_x$ and GLCM$_z$ (Figure 4.9).

We are able to delineate the different reflection patterns recorded in the Wellington fault zone GPR data using just three of the twelve texture attributes: entropy$_z$, homogeneity$_x$, and contrast$_x$. Figure 4.10 shows cross-sections extracted from these attribute volumes along profile A-A’ of Figure 4.7a. Steeply dipping reflections from the sheared basement produce moderate to high contrast$_x$ values, whereas the semi-continuous shallowly dipping reflections from the older gravel deposits produce only moderate contrast$_x$ values (Figure 4.10c). On an individual inline cross-section of the original GPR volume (Figure 4.8a), these differences between the two reflection patterns are difficult to recognize. Because the texture attribute cross-sections are calculated from multiple inlines, they provide a more robust indication of changes to the 3-D reflection patterns.
Figure 4.9: Horizontal slices extracted at 5 m depth from the suite of texture-attribute volumes calculated from the Wellington fault zone GPR data. Each column shows in descending order the attributes calculated for the x-, y-, and z-directions. B-basement of metasediments and older deformed Holocene gravels; S-Holocene gravel sheets. Dashed lines identify the interpreted location of the principal fault strand.
4.6. Interpretation of the Wellington fault zone data set

Figure 4.10: A–A’ cross-sections extracted from the most useful Wellington fault zone texture-based attribute volumes (see location in Figure 4.7a). Note the color scale at the base of the figure. The texture attributes are plotted as semi-transparent colored layers overlying a variable area representation of the original GPR data. B, G, S, dashed lines and arrows are as in Figure 4.8. The lower dotted line separates reflections with high signal to noise (above) from those with low signal to noise (below). No vertical exaggeration.
Although the layered nature of the Holocene gravel sheets is recognizable from the original data (Figure 4.8a), the texture attributes allow us to delineate accurately the lower boundary of this facies; the continuous horizontal reflections are well defined by high homogeneity and low contrast values (S in Figures 4.10b and 4.10c). High entropy values, which correlate with high frequency reflections, generally coincide with radar facies S (Figure 4.10a).

The dotted lines in Figure 4.10 delineate the depths to which we can make reliable interpretations of the radar facies. Homogeneity is uniformly low for B and G at shallow depths (<10 m), but increases with depth as the signal-to-noise (S/N) of the reflections decreases (Figure 4.10b).

Certain texture attributes are also useful for identifying reflections from above-ground objects, which are a common source of noise in GPR data. High entropy values at the northwestern end of the cross-section between z = 10 and z = 17 m (Figure 4.10a) coincide with high frequency reflections from nearby trees (Figures 4.3a and 4.3b) that were initially identified after analyzing unmigrated cross-sections of the original data.

### 4.7 Interpretation of the Maleme fault zone data set

A portion of the processed Maleme fault zone GPR data is displayed in Figure 4.6b. Relatively continuous reflections are disrupted by a previous trench excavation (rectangular region), and by the complex zone of faulting beneath the topographic scarp.

#### 4.7.1 Application of coherence-based attributes to the Maleme fault zone data set

Unlike the Wellington fault zone, the material contrasts caused by faulting are not large enough to produce fault-plane reflections in the Maleme fault zone GPR volume. Rather, the faults are evident from offsets of the otherwise continuous reflections from the shallow-dipping stratigraphy (Figure 4.6b). To highlight these sharp changes, a relatively small trace analysis window was used to generate the coherence-based attribute volumes. After testing a range of dimensions, we determined an optimum trace analysis window size of 3 x 3 samples (0.3 x 0.5 m) and a time width of 5 ns (corresponding to 0.18 m in the depth-converted volume).

The relatively gentle dips observed on cross-sections of the original GPR volume (Figure 4.6b) allowed us to limit the dip search to a maximum of 20°. Horizontal slices extracted at z = 4.5 m from the original, coherency, azimuth, and dip volumes are displayed in Figures 4.11a-4.11d. The coherency attribute is effective in highlighting the locations of reflection discontinuities, including fault strands, and the former walls of the trench, which extends from z = 1 to 7 m within
4.7. Interpretation of the Maleme fault zone data set

the GPR volume. We interpret two fault strands south of the trench that merge into a narrow zone of closely spaced faults north of the trench (mf1, mf1a, and mf1b; Figure 4.11b). Two subparallel lineations observed on the footwall block to the south of the main fault strands (shown by the red arrow in Figure 4.11b) are only observed at shallow depths, suggesting that they are either man made or represent disconnected extension fractures formed between faults.

The azimuth slice shows that, except for the regions near the fault strands, azimuths are predominantly southeast (green in Figure 4.11c). Near the fault strands, a combination of flexure, block rotation and the mantling of sediments across the northwest-facing scarps produce northwest azimuth values (red-pink in Figure 4.11c). These effects are also observed on the corresponding dip attribute slice, where reflections are generally gently dipping (<10°), but become steeper over 5-to-10-m-wide zones that straddle the fault strands (outlined by thick blue lines in Figure 4.11d).

A cross-section extracted from the GPR volume south of the trench shows fault strands mf1a and mf1b (Figure 4.12a; profile B-B’ of Figure 4.11a). These faults coincide with low coherence lineaments on the corresponding coherency attribute cross-section (Figure 4.12b). Although mf1a has a steep dip (~80°) that is characteristic of normal faults within the Taupo Rift Zone, mf1b dips moderately (~40°) to the southeast, exhibiting an apparent reverse sense of slip (Figures 4.12a and 4.12b). Villamor et al. [2006] observe false reverse faults from numerous trenches excavated within the Taupo Rift Zone. They are most likely the result of fault toppling, where the over steepened block on the upthrown side of a fault topples towards the downthrown side, rotating the fault plane and producing the geometry of a reverse fault.

4.7.2 Application of texture-based attributes to the Maleme fault zone data set

Texture-based attributes were calculated from the original GPR volume using analysis windows with dimensions of 7 x 5 x 30 samples (0.90 x 1.00 x 0.53 m). Generally, the continuous reflections from the stratigraphic horizons have monotonously uniform textures. Substantial changes in texture coincide with reflection discontinuities at the fault strands and at the walls of the trench. These are evident on the horizontal slices extracted at 4.5 m depth from the contrast x and contrast y attribute volumes (Figures 4.11e and 4.11f). High contrast x values highlight NE-SW-oriented discontinuities, including the fault strands identified on the coherency attribute images (Figures 4.11b and 4.11e). Conversely, high contrast y values emphasize NW-SE-oriented discontinuities such as the trench walls (Figure 4.11f).

High values of contrast x in the cross-section of Figure 4.12c show the two main fault strands more clearly than the coherency values of Figure 4.12b. This difference is attributed to the different analysis window sizes used to generate the two types of attribute; the relatively small windows used to generate the coherence-based at-
Figure 4.11: Horizontal slices extracted at 4.5 m depth from the Maleme fault zone original GPR volume and various attribute volumes. White arrows identify interpreted fault strands. Rectangles outline paleoseismic trench location. Red arrow in (b) points to an interpreted extension fracture. Thick blue lines in (d) define a zone of stratal folding with dips generally >10° that surround the fault strands. Cross-sections along profile B-B’ in (a) are shown in Figure 4.12. For the coherency related attributes, a dip search up to a maximum value of 20° was used.
4.7. Interpretation of the Maleme fault zone data set

Figure 4.12: Cross-sections extracted from the Maleme fault zone original, coherency, and contrast attributes along profile B-B’ shown in Figure 4.11a. Arrows identify interpreted faults shown on the time slices of Figure 4.11. Vertical exaggeration of 2.
tribute volumes emphasize subtle deformation features, including multiple overlapping strands and short discontinuities created by faulting, whereas the larger analysis windows used to calculate texture-based attributes smooth over sharp discontinuities while highlighting the broader zones of deformation.

4.7.3 Trench stratigraphy and correlation with GPR data

Geologic logs from the trench walls revealed alternating layers of fine- and coarse-grain alluvium overlain by a succession of tephra and associated paleosol (Figure 4.13a). This well dated stratigraphic sequence is recognized at other localities within the Taupo Rift Zone [Villamor and Berryman, 2001]. Abrupt changes in grain size produce strong reflections in our GPR data. Although the trench logs reveal more deformation than can be observed in the corresponding GPR cross-sections, at least three reflection horizons correlate with prominent tephra-paleosol boundaries logged in the walls of the trench (MR1, MR2, and MR3 in Figures 4.13b and 4.13c). The ages of surfaces MR1, MR2, and MR3 are constrained by radiocarbon dating to 1718 ± 30, 15425 ± 325 and 24352 ± 780 cal. years BP, respectively [Froggatt and Lowe, 1990; Villamor and Berryman, 2001; Alloway et al., 2007].

4.8 Interpretation of the Ostler fault zone data set

A portion of the processed Ostler fault zone GPR volume is displayed in Figure 4.6c. Reflections from bedding within the deformed outwash gravels dip gently to the east, parallel to the topography west of the deformation front. East of the break in the slope, stratigraphic bedding is effectively horizontal (Figure 4.6c).

4.8.1 Application of coherence-based attributes to the Ostler fault zone data set

Like the Wellington fault GPR volume, the processed Ostler fault zone GPR volume contains numerous short events superimposed on more continuous reflections (Figure 4.6c). A number of shallow- to moderate-dipping reflections from reverse faults are imaged in the original GPR volume. We estimate that these reflections have maximum dips of ~30°. To suppress the short wavelength events and emphasize the more continuous events, coherency, azimuth, and dip attribute volumes are generated from the GPR data using a relatively large 5 x 5 trace analysis window (1 x 1 m) and a dip search of up to 35°. A short vertical analysis window of 4 ns (equivalent to ~0.25 m in the depth converted volume) is used to resolve the shallowly dipping faults.

Figure 4.14 shows time slices extracted at 8 m depth from the coherence-based attribute volumes. The coherency attribute delineates clearly the boundaries of shallow channels (Ch in Figure 4.14b) within the horizontally layered gravels east of the
4.8. Interpretation of the Ostler fault zone data set

Figure 4.13: Correlation of GPR data with paleoseismic trench observations. (a) Geologic log from north wall of paleoseismic trench. Arrows point to lithologic contacts correlated with GPR reflections. (b) Outline of trench superimposed on a cross-section extracted from the 3-D GPR data. Yellow shading follows tephra-paleosol contacts correlated with GPR reflections MR1, MR2, and MR3. Airwave and groundwave obscure reflections from parts of the MR1 horizon within the footwall. Dotted lines show the corresponding positions of horizon MR2 in the (a) trench and (b) GPR sections. (c) As for (b), but without interpretation.
deformation front, but provides few details on structures within the deformed gravels. Fortunately, the azimuth attribute slice (Figure 4.14c) supplies important structural details within this region. It shows how west-dipping fault-plane reflections (of1, of2, of2a, and of2b; purple) can be distinguished from east-dipping stratigraphic reflections (green). In addition, this slice vividly illustrates the orientation of the north- (red) and south-facing (blue) channel banks. The dip attribute slice demonstrates that the fault-plane reflections are generally steeper dipping than the stratigraphic reflections (Figure 4.14d).

We can combine information contained in the coherence-based attribute volumes to highlight features that exhibit certain geometrical properties such as fault-plane reflections. For example, Figure 4.15 displays a series of horizontal slices extracted at 6.5, 7.5, 8.5, and 12.0 m depth from the original and azimuth attribute volumes. To emphasize the relatively coherent reflections from the moderately dipping faults, we only plot events that have high coherencies >0.5 and dips >20° in the azimuth attribute slices. From these, we identify at least three curvilinear fault strands that traverse the width of the survey area (of1, of2 and of3 in Figure 4.15). Two of these strands (of2 and of3) appear to bifurcate at different depths into separate strands in the southern half of the survey (of2a, of2b, of3a, and of3b in Figures 4.15e - 4.15g). These complex features would have gone unnoticed had we based our interpretation solely on the original data (Figures 4.15a - 4.15c). As depths approach 12.0 m, reflections from the fault strands are no longer visible (Figure 4.15h).

4.8.2 Application of texture-based attributes to the Ostler fault zone data set

Texture-based attribute volumes were calculated from the original GPR volume using analysis windows with dimensions 7 x 7 x 30 samples (1.5 x 1.5 x 0.7 m). Reflections from the deformed outwash gravels (west side of horizontal slices in Figure 4.16) have distinctly different textures than those of the undeformed outwash gravels (east side of the slices) in attribute volumes calculated from GLCMx. The horizontal slices in Figure 4.16 emphasize subtle textural changes across the width of the fault zone. Curvilinear features highlighted by low homogeneityx and high contrastx correspond to the faults identified on the coherence-based attribute images. By inspecting horizontal slices from both the coherence- and texture-based attributes, we are able to pick the faults with greater confidence.

Figure 4.17a shows a cross-section through the original GPR volume (profile C-C′ of Figures 4.14a and 4.15a), on which we indicate three of the fault strands identified in the attribute slices (of1, of2 and of3). Unlike fault of1, which forms the deformation front, faults of2 and of3 have no obvious surface expression (Figure 4.5b). They would not have been identified from surface mapping. In addition, reflections from the layered gravels above and between faults of1 and of2 have a
4.8. Interpretation of the Ostler fault zone data set

Figure 4.14: Horizontal slices extracted at 8 m depth from the Ostler fault zone original GPR and coherence-based attributes. (a) Original GPR data. Cross-sections along C-C’ are shown in Figure 4.17. (b) Coherency. (c) Azimuth. (d) Dip. The algorithm performed a dip search up to a maximum value of 35°. Azimuths and dips with coherencies <0.5 are plotted gray. Ch interpreted troughs of channels. Arrows identify individual fault strands.
Characterization of active faults using 3-D GPR data

Figure 4.15: Horizontal slices extracted from the Ostler fault zone (a) - (d) original GPR volume and (e) - (h) azimuth attribute volume at depths of 6.5, 7.5, 8.5, and 12.0 m. Shallowly dipping (<20°) and low coherency (<0.5) elements are not plotted in (e) - (h) in order to highlight coherent reflections from the moderately dipping (>20°) faults. Arrows identify individual fault strands.
4.8. Interpretation of the Ostler fault zone data set

Figure 4.16: Horizontal slices extracted from the Ostler fault zone (a) - (d) homogeneity, and (e) - (h) contrast attribute volumes at the same depths as shown in Figure 4.15. Arrows identify individual fault strands.
warped, concave-down geometry (inset of Figure 4.17a). Long wavelength (>50 m) fault-related folding is an important mechanism for accommodating shortening across the Ostler fault zone [Davis et al., 2005; Amos et al., 2007]. Our GPR data suggest that this process may also be happening on a smaller scale (wavelengths <10 m).

On the original GPR sections, the fault-plane reflections are only evident at depths shallower than 10 m (Figure 4.17a). At greater depths, at least one reflection horizon from the layered gravels (OR1), which can be semi-automatically picked throughout the GPR volume (Figure 4.17b), is vertically offset by ~2 m and ~3 m across the projections of faults of1 and of2, respectively. Furthermore, relatively low homogeneity values and high contrast values delineate subparallel linear structures that follow the projection of the dipping fault-plane reflections to depths >10 m (Figures 4.17c and 4.17d). We attribute the low homogeneity and high contrast values to a decrease in reflection amplitude where the layered strata are disrupted by faulting.

4.9 Discussion

Figure 4.18 summarizes our geological interpretations of the three GPR data sets. The three models are plotted at the same scale. Major structural features interpreted from the original GPR and attribute volumes are shown for each.

4.9.1 Wellington fault zone

Three of the major geological units in our Wellington fault zone model (Figure 4.18a) match those identified by Gross et al. [2004]. We interpret an extra unit of disrupted sediment, which corresponds to the low coherency zone between the steeply dipping fault-plane reflections and the high-coherency reflections of the Holocene gravel sheets on the downthrown block (Figures 4.8b and 4.18a). Based on other interpretations made at the nearby river exposure (Figure 4.3c; Berryman [1990]; J. Begg, pers. comm.), the disrupted zone either represents scarp-derived colluvium deposited in a tension crack created by ground rupture or results from coseismic disruption of the layered gravel sheets on the downthrown block. Disrupted clastic units that parallel the fault plane have also been observed within exposures of the fault zone at other locations (e.g., Figure 4.3d; Berryman [1990]).

The fault plane geometry measured in the GPR data (Figures 4.8c and 4.8d) is very similar to that measured at the riverbank exposure, where the fault plane dips at 65° to the southeast (Figure 4.3c). These observations suggest that along several hundred meters of fault strike the near-surface morphology of the Wellington fault zone is relatively uniform. From the attitude of the fault plane and sense of vertical offset, we infer that this branch of the Wellington fault zone has experienced a component of extension, possibly associated with overlapping en-echelon
4.9. Discussion

Figure 4.17: (a) Cross-section extracted from the Ostler fault zone GPR original volume along profile C-C’ (see location in Figure 4.15a). Three faults are identified on the basis of fault-plane reflections and truncations of shallowly dipping strata (of1, of2, and of3). The inset in (a) shows a zoom into the fault plane of one of the faults (of1). (b) Fence diagram showing semi-automatically picked horizon OR1, which is vertically offset by ~2 m and ~3 m across of1 and of2, respectively. The picked horizon does not include the reversed dips of the faults because the picking algorithm has difficulties tracking reflections with reduced continuity near the faults. (c) and (d) Profiles C-C’ cross-sections extracted from the homogeneity x, and contrast x attribute volumes.
traces. We estimate that the fault zone at this location has a minimum width of 8 m. Because we have not imaged any piercing points from which vertical and strike-slip separation can be measured, we would not recommend opening a trench at this location.

4.9.2 Maleme fault zone

Our interpretation of the Maleme fault zone GPR volume also shows a complicated pattern of merging and overlapping fault segments. Closely spaced overlapping segments define a principal fault strand that bifurcates into two strands near the center of our survey area (Figures 4.11 and 4.18b). The NE-SW-oriented principal strand has a steep dip that is characteristic of normal faults within the Taupo Rift Zone, whereas the shallower dipping subsidiary strand has an apparent reverse geometry that is probably the result of fault toppling (Figure 4.12). Fault-toppled deposits have been observed at numerous trench excavations within the Taupo Rift Zone. They occur where faults exhibit listric geometries; material on the over-steepened upthrown side of the fault tends to topple towards the downthrown side, rotating the fault plane and producing an apparent reverse geometry [Villamor et al., 2006]. It is difficult to reconstruct past fault movements from trench logs that feature these types of structures. We suggest that 3-D GPR surveying can be used to identify such complicating features and help find more favorable locations for trenching.
4.9. Discussion

Cross-sections from the Maleme fault zone GPR data provide reliable images of offset stratigraphic horizons that can be used to measure fault offsets (Figure 4.12). Nevertheless, precise offset measurements are difficult to make because the layers between the reflection horizons taper toward the fault plane and are deflected in the direction of fault slip. In reflection seismology, the curvature of stratigraphic markers near faults is usually interpreted as the result of distributed strain (e.g., Chapman and Meneilly [1991]; Lamarche et al. [2006]); it can be accounted for in fault displacement measurements by projecting the planar parts of the horizon onto the fault plane. Unfortunately, because certain deposits (e.g., tephra) mantle pre-existing fault scarps in the Taupo Rift Zone, not all horizon curvature can be attributed to tectonic deformation [Villamor and Berryman, 2001]. This effect can be significant on the scale of high resolution GPR surveys, such that fault offsets based on this projection technique would represent maximum estimates of displacement.

A minimum estimate of accumulated displacements from past earthquakes could be obtained by measuring the vertical separation of chronostratigraphic horizons near the fault plane. On the coherency attribute slice of Figure 4.11b, the zone of brittle fault deformation is narrowest at y = 28 m. At this location, reflection horizon MR2 is vertically offset 1.2 m across fault strand mf1. This horizon corresponds to the base of the Rotorua tephra unit (Figure 4.13), which has a radiocarbon age of 15425 ± 325 cal. years BP [Froggatt and Lowe, 1990; Alloway et al., 2007]. Using these constraints, we calculate a minimum vertical displacement rate of 0.08 mm/yr for this fault.

4.9.3 Ostler fault zone

Our Ostler fault zone GPR volume provides images of a complicated pattern of linked faulting (Figure 4.18c). At least three shallow- to moderate-dipping (20-30°) faults are identified within the weakly layered glacial outwash gravels. Two of these faults bifurcate into separate strands in the southern half of the survey area (of2 and of3). These types of segments are interpreted by Davis et al. [2005] as strands within the Quaternary and Tertiary sediments that merge to a single strand within the greywacke bedrock; a single earthquake probably ruptures multiple strands at the surface. Davis et al. [2005] report average fault dip measurements for the Ostler fault zone of 50 ± 9°, whereas Amos et al. [2007] determined a value of 56 ± 9°. Both values are significantly steeper than our observations. Faults observed in our GPR data exhibit progressively shallower dips approaching the deformation front (of3 to of1; Figures 4.17 and 4.18c). Their morphology resembles that of hanging wall collapse scarps, which form by gravitational collapse of the overhanging scarp [Philip et al., 1992]. We expect fault dip to progressively increase at greater distances from the deformation front.

Published slip rates for the Ostler fault zone are based exclusively on geomorphic interpretations (e.g., Davis et al. [2005]; Van Dissen et al. [1994]). Although
Van Dissen et al. [1994] estimated the timing of past earthquakes from dated colluvial deposits within a shallow trench, they did not uncover fault-offset stratigraphic markers from which a subsurface slip rate could be determined. Interpretations of our GPR data indicate that fault strands of 2 and of 3 vertically offset a shallow (<7 m depth) reflection horizon within the outwash gravels by ~5 m (OR1; Figure 4.17b). Based on this interpretation, we would recommend the excavation of a deeper paleoseismic trench at our survey site to obtain datable material from horizon OR1.

Reverse faults are difficult to characterize on the basis of topographic expression for three reasons: (1) erosion tends to remove topographic scarps, (2) deformation may only be expressed as folding at the surface, and (3) the fault planes may shallow as they approach the surface, often following sedimentary boundaries. Our Ostler fault zone GPR data not only reveal faults not evident at the surface, they allow us to determine their subsurface geometry (Figures 4.17 and 4.18c). In addition, the multiple strands and the fault-related folding we interpret (Figure 4.17a) suggest that deformation is taking place on a smaller scale than can be observed at the surface.

4.9.4 Role of surface lithologies on fault character

We suggest that the complexity of faulting in the near surface is largely dependent on the competency of the faulted material. Although we observe a wide region of fault-parallel shearing within the gravel/greywacke basement of the Wellington fault zone, the strand of the fault that has moved in the past few tens of thousands of years seems to be relatively narrow. It appears to be controlled by the basement rock in the shallow subsurface (Figure 4.18a). Conversely, faults within the Maleme fault zone and Ostler fault zone surveys are widely distributed, offsetting poorly consolidated glacial outwash gravels and unwelded volcanic ashes and fluvial gravels, respectively. At these locations, the faults merge and change orientation within just a few meters along strike (Figures 4.18b and 4.18c).

4.10 Conclusions

We have demonstrated the applicability of 3-D GPR surveying for subsurface mapping of active faults in a range of tectonic settings. The processed GPR volumes showed details of faulting on a scale of a few tens of centimeters to depths >10 m. Many of the identified deformation structures were not evident at the surface. Using surface-based observations from geological exposures and trenches as ground truth, our 3-D volumes allow complicated fault zone geometries to be extrapolated along and across fault strike.

In addition to faulting, we interpreted collapse/topple structures and fault-related folding within the Maleme and Ostler fault zone data sets. By identifying the loca-
4.10. Conclusions

 tion of collapse/topple features and off-fault folding, future paleoseismic trenches can be excavated at locations where fault deformation is less complicated. Our data sets also showed how the geometry of faulting can change over relatively short distances along strike.

We have demonstrated the benefits of using geometric attributes to represent 3-D information on 2-D displays. Coherence- and texture-based attributes were successfully used to emphasize structures and to identify changes in reflection pattern associated with different lithological units. Some attributes were more informative than others. For example, the coherency attribute was an excellent tool for highlighting normal faults within the volcanic deposits of the Maleme fault zone, whereas the texture-based attributes were most useful for discriminating the different facies juxtaposed by the Wellington fault. In addition, the parameters used to calculate geometric attributes from the processed GPR volumes were dependent on the resolution and structural characteristics of each data set:

1) For the Wellington fault zone data, relatively large analysis windows (1 x 2 m) and a wide range of search dips (0-80°) were required to generate reliable coherence-based attributes that captured the geometry of the reflections from the steeply dipping fault plane and the sheared greywacke/gravel basement. Coherency and certain texture attributes, which were calculated using 1.5 x 2.0 x 1.0 m analysis windows, provided excellent discrimination between reflections from the Holocene gravel sheets and reflections from sheared greywacke/gravel basement and an older gravel deposit.

2) Steeply dipping faults within our Maleme fault zone data set did not produce reflections in the GPR data, but were evident from truncations of the shallowly dipping reflections generated at stratigraphic boundaries. As a consequence, individual faults could be imaged on the basis of their low coherency. To avoid smoothing out these structures, relatively small trace analysis windows (0.3 x 0.5 m) with a time width of 5 ns (corresponding to 0.18 m in the depth converted volume) were used to generate coherence-based attribute volumes. Conversely, the broader deformation that characterized the fault zone was highlighted in contrast attribute volumes that were calculated using analysis windows with dimensions 0.90 x 1.00 x 0.53 m. Azimuth and dip attributes were used to visualize stratal folding surrounding the fault strands.

3) Because reflections from reverse fault planes within our Ostler fault zone data set were shallow dipping, we used a relatively short time window of 4 ns (equivalent to ~0.25 m in the depth converted volume) and a narrow range of search dips (0-35°) to calculate coherence-based attributes that highlighted these structures. Similarly, homogeneity and contrast attributes calculated with a short vertical dimension (0.30 x 0.50 x 0.18 m) provided comparable definition of the faults and allowed us to image their continuation to depths >10 m.
4.11 Acknowledgements

We greatly appreciate the contributions of Ralf Gross, Jens Tronicke and other members of the ETH, University of Canterbury, and Victoria University of Wellington field crews. This project was supported by grants from the Swiss National Science Foundation and ETH Zurich.
Chapter 5

Fault displacement accumulation and slip-rate variability within the Taupo Rift (New Zealand) measured using high-resolution 3-D ground-penetrating radar

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submitted to Tectonics

5.1 Abstract

In offshore regions, studies based on densely spaced reflection seismic data tied to stratigraphic logs demonstrate that active faults can have variable displacement rates over relatively short distances and short time intervals. Here, we demonstrate how high-resolution 3-D ground-penetrating (GPR) data tied to trench-derived stratigraphic logs provide similar information for active faults in onshore regions. To investigate recent (<24.4 kyr) fault activity within the Taupo Rift of New Zealand, we analyze 3-D GPR data acquired over 10 fault strands within the Maleme fault zone. After correlating three prominent GPR reflection horizons with three faulted chronostratigraphic units observed within a trench, we extrapolate the geometries of the horizons over a ~150 x 250 m area of the fault zone and determine slip-accumulation patterns and rates. Profiles of cumulative fault displacement measured for horizons older than 12.5 kyr exhibit characteristic displacement distributions. By calculating average cumulative displacements with time for five practically complete fault strands, we obtain robust slip-rate estimates. Slip rates are
variable for time intervals \( \leq 12.5 \) kyr long, suggesting that at least four earthquakes are required for these faults to exhibit uniform slip rates characteristic of their long-term behavior.

### 5.2 Introduction

Probabilistic seismic analyses that include paleoseismic observations generally require knowledge of two earthquake parameters: their characteristic magnitudes and their average recurrence intervals [Thenhaus and Campbell, 2003]. Surface displacement measurements and appropriate empirical relationships provide estimates of earthquake magnitude, and the ages of deformation attributable to earthquakes supply information on recurrence intervals [Wells and Coppersmith, 1994; Hemphill-Haley and Weldon, 1999; Cowie and Roberts, 2001]. Although average slip distributions derived from populations of faults can exhibit common generic properties [Manighetti et al., 2005], slip rates and ground-rupture patterns for individual faults can vary with space and time (e.g., Schwartz and Coppersmith [1984]; Fumal et al. [1993]). Accordingly, it is necessary to understand how displacements accumulate along the lengths of faults over several earthquake cycles.

Historical records are much shorter than earthquake recurrence intervals in most parts of the world, such that fault slip rates must be determined using a variety of paleoseismic techniques. These include trenching and geomorphic and geodetic observations in onshore regions, and borehole studies and the analysis of reflection seismic data in offshore areas.

In onshore regions of tectonic extension, trenching can be used to determine slip rates on active faults (e.g. Pantosti et al. [1996]; Langridge et al. [2000]; Nicol et al. [2006]); trenching allows offset marker horizons to be measured and dated. In addition, the occurrence of earthquakes can be inferred from observations of colluvial wedge deposits, progressive displacement of older units, and fault terminations. A key limitation of trenching is that slip rates derived at single points along fault strands may not be representative of the maximum or average displacements along the entire lengths of the faults [Cowie and Roberts, 2001; Kim and Sanderson, 2005]. Although multiple trench excavations along fault strands would reduce such uncertainties (e.g., McCalpin [2005]), it is more common to use geomorphic observations of fault scarps to place constraints on how displacements may have changed along the strikes of faults (e.g., Pantosti et al. [1996]; Berryman et al. [1998]; Personius and Mahon [2005]). Unfortunately, modification of fault scarps by erosional and depositional processes can make such interpretations difficult. Interpretations can be particularly problematic in immature fault zones, where faults can be highly segmented and smaller scarps associated with the simultaneous rupture of multiple strands may not be apparent.

In offshore regions that have undergone extension, dense high-resolution 2- and 3-D reflection seismic data tied to dated stratigraphic logs have allowed both the
spatial and temporal variations in slip rate to be measured over multiple earthquake cycles [Petersen et al., 1992; Mansfield and Cartwright, 1996; Nicol et al., 1997, 2005]. Taylor et al. [2004], Bull et al. [2006], and Lamarche et al. [2006] used densely spaced reflection seismic data to determine syn-sedimentary displacements along and across active faults within the offshore part of the Taupo Rift on the North Island of New Zealand. These data revealed how faults grew and accommodated slip over relatively short time intervals (≥2.4 kyr) during the past 1.3 Myr. By elucidating their short-term slip-rates, Bull et al. [2006] demonstrated that displacement patterns on active faults were variable with time and earthquakes with markedly different magnitudes may have occurred along the same fault strand. Clearly, sparse trenching, geomorphic observations and geodetic surveying would only partially characterize the threat of future ground ruptures along the onshore extensions of these faults.

It would be advantageous to have densely spaced reflection seismic data across tectonically active onshore regions, where seismically active faults represent a greater hazard to populated regions. Unfortunately, dense land-based reflection seismic surveying is a very expensive and logistically challenging endeavor. However, low-cost 3-D ground-penetrating radar (GPR) surveying offers a viable alternative to onshore reflection seismic surveying; 3-D GPR surveying is capable of providing vivid subsurface images that can be correlated with trench excavations (e.g., McClymont et al. [submitted]).

We have used a combination of trenching and high-resolution 3-D GPR surveying to investigate spatial and temporal variability of displacement accumulation along active faults within the onshore Taupo Rift (Figure 5.1). For our study, we have targeted faults of the Maleme fault zone, which were active during the formation of a well-bedded and well-dated sequence of alluvium, tephra, and associated paleosol. Fault offsets across some of the marker horizons observed in a trench excavation were well resolved on the GPR data. By correlating measurements of fault displacements in the trench to a large 3-D GPR data set, we were able to extrapolate displacement information along and across fault strike. Our integrated analysis has yielded a high-resolution ~25 kyr history of fault displacement for 10 strands within a ~150 x 250 m area of the fault zone.

We begin by describing the extensional tectonics of the Taupo Rift with emphasis on the Maleme fault zone. Important details of GPR data acquisition and processing are then explained. Because of the relatively large surface area covered by our GPR survey, we outline some novel processing steps that are required to image accurately the subsurface structures. By correlating our processed GPR data with observations made in the trench excavation, we show how the chronostratigraphic horizons can be extrapolated away from the trench walls. Our method for estimating along-strike fault displacements from offsets and interruptions of these horizons is then described. Finally, from the derived cumulative fault-displacement profiles, we analyze spatial and temporal changes in slip rate within a section of the
Figure 5.1: (a) New Zealand plate boundary setting. Major faults are plotted as bold lines. Bold arrows show relative convergence of the Pacific and Australian plates. Box outlines the area shown in (b). (b) Digital terrain model and active faults (black lines) of the Taupo Rift. Shaded relief topography and fault trace data were obtained from the GNS Science Active Faults Database (http://data.gns.cri.nz/af/). White dashed ellipse outlines the Maleme fault zone, within which the GPR survey is located (circle). Star marks the main shock epicenter of the 1987 Edgecumbe earthquake. Geographic projection is New Zealand Map Grid.
5.3. **Tectonic Setting**

5.3.1 The Taupo Volcanic Zone and Taupo Rift

The Taupo Volcanic Zone (TVZ) is a region of active back-arc extension and volcanism resulting from oblique westward subduction of the Pacific Plate beneath the Australian Plate (Figure 5.1a). It has been active for \(~2\) Myr, forming one of the most productive silicic volcanic systems worldwide [Wilson et al., 1995]. The central TVZ is a north-northeast - south-southwest trending 20-30-km-wide zone characterized by high heat flow (700 mW/m²), active normal faulting, and intense shallow (<10 km depth) seismicity [Anderson and Webb, 1994; Bibby et al., 1995]. At its southern terminus, faulting and volcanism have only been active for <0.4 and <0.3 Myr, indicating that the processes defining back-arc extension are migrating southward [Gamble et al., 2003; Villamor and Berryman, 2006b]. The tectonic structure within the TVZ is known as the Taupo Rift [Villamor and Berryman, 2006a].

The northern offshore continuation of the Taupo Rift is the Havre Trough, an actively extending back-arc basin that is well-separated from the Kermadec Ridge volcanic arc to the east (Figure 5.1a; Gamble and Wright [1995]. Unlike the Havre Trough and Kermadec Ridge, active normal faulting and volcanism occur within overlapping domains of the TVZ.

Most of the extension across the Taupo Rift is accommodated on steeply dipping (60°-90°) normal faults [Villamor and Berryman, 2001]. Evidence for active tectonism includes the M\(_{w}\) 6.5 Edgecumbe earthquake of 1987 (Beanland et al. [1989]; Figure 5.1b), which ruptured multiple fault segments within the northern section of the rift, including one \(\sim7.0\)-km-long segment. Extension rates determined from GPS observations decrease from \(\sim15\) mm/yr in the north to <5 mm/yr in the south [Wallace et al., 2004b]. Estimates of extension derived from paleoseismic data are generally lower than the geodetic observations, decreasing from \(\sim7-12\) mm/yr across the northern and central sections of the rift to \(\sim2-6\) mm/yr at its southern end [Villamor and Berryman, 2001; Wallace et al., 2004b; Villamor and Berryman, 2006a; Nicol and Wallace, 2007]. In places, the geologically determined extension rates are less than half the geodetic rates. This suggests that the paleoseismological data are incomplete and/or that some of the extension is accommodated by other processes such as dike intrusion [Villamor and Berryman, 2001; Wallace et al., 2004b; Nicol and Wallace, 2007].
5.3.2 The Maleme fault zone

The Maleme fault zone defines part of the central axis of the Taupo Rift (Figure 5.1b). Surface and subsurface (GPR) mapping demonstrates that it comprises a dense array of at least 16 splaying and merging fault strands that span a ~2.5-km-wide graben (Figure 5.2a; Villamor and Berryman [2001]; Tronicke et al. [2006]; McClymont et al. [submitted]).

We have acquired a large 3-D GPR dataset across several fault strands at the northern end of the Maleme fault zone (Figures 5.1b and 5.2a). A total of 10 distinct fault strands were imaged in our GPR data (s1-s10 in Figure 5.2b). Although most of the fault strands are distinguished by 1-2m-high topographic scarps, many are not evident from surface mapping (compare Figures 5.2a and 5.2b). Fault strands s1, s5, s6, and s7 define a local graben (Figure 5.2b). Although individual strands are non-linear, they all have a general northeast-southwest strike.

These faults transect the sequence of weakly consolidated Late Pleistocene and Holocene alluvium, tephra, and associated paleosols. Because these units are distinguished by low strengths, past ground ruptures have produced complicated shallow deformation patterns that are not visible at the surface. A small portion of our data across fault strand s1 (location shown in Figure 5.2b) has been interpreted by McClymont et al. [submitted] in terms of collapse structures, fault-drag folding, and disconnected extension fractures.

5.4 Data acquisition and processing

To sample efficiently a relatively large ~150 x 250 m region of the fault zone, we used a Sensors and Software PulseEKKO GPR unit linked to a self-tracking laser theodolite [Lehmann and Green, 1999]. Acquisition and processing parameters for the survey are summarized in Tables 5.1 and 5.2. The total survey area comprised 6 regions (F1-F6 in Figure 5.2b). Small gaps between the regions were a consequence of obstacles on the ground, including trees and fence lines.

<table>
<thead>
<tr>
<th></th>
<th>F1-F2</th>
<th>F3-F6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nominal antenna frequency</td>
<td>100 MHz</td>
<td>100 MHz</td>
</tr>
<tr>
<td>Antenna separation</td>
<td>1.0 m</td>
<td>1.0 m</td>
</tr>
<tr>
<td>Antenna configuration</td>
<td>Perpendicular-broadside</td>
<td>Perpendicular-broadside</td>
</tr>
<tr>
<td>Minimum stacks per trace</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>Sampling rate</td>
<td>0.5 ns</td>
<td>0.5 ns</td>
</tr>
<tr>
<td>Trace length</td>
<td>460 ns</td>
<td>460 ns</td>
</tr>
<tr>
<td>Grid spacing</td>
<td>0.15 x 0.25 m</td>
<td>0.25 x 0.5 m</td>
</tr>
</tbody>
</table>

Table 5.1: Acquisition parameters for the six 3-D GPR surveyed regions (F1-F6).
5.4. Data acquisition and processing

Figure 5.2: (a) Active traces mapped at the surface of the Maleme fault zone (teeth marks are on downthrown side) and the location of the GPR survey. Box outlines area shown in (b). (b) Orthorectified photo highlighting the topography measured for GPR traces within the six surveyed regions (F1-F6) and the location of the Huffadine paleoseismic trench. Numbered lines with teeth delineate the locations of normal fault strands s1-s10 identified within the processed GPR volumes. White dashed lines (GA) define the axes of two laterally offset grabens. White and black dots show CMP locations; white dot identifies the control point CMP shown in Figure 5.15. Dotted lines outline the areas of the Tronicke et al. [2006] and McClymont et al. [submitted] 3-D GPR studies. The x and y axes define the local coordinate directions of our GPR surveys. The “+” defines the origin of the coordinate system. Geographic projection in (a) and (b) is New Zealand Map Grid.
Table 5.2: 3-D GPR data processing sequence

<table>
<thead>
<tr>
<th>Processing Step</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Assignment of coordinates to the traces</td>
</tr>
<tr>
<td>2. Median (dewow) filter</td>
</tr>
<tr>
<td>3. Trace editing</td>
</tr>
<tr>
<td>4. Alignment of first arrivals</td>
</tr>
<tr>
<td>5. Correction for daily velocity variations</td>
</tr>
<tr>
<td>6. Interpolation and trace binning</td>
</tr>
<tr>
<td>7. Amplitude scaling</td>
</tr>
<tr>
<td>8. Bandpass filtering</td>
</tr>
<tr>
<td>9. 3-D topographic Migration</td>
</tr>
<tr>
<td>10. Time to depth conversion</td>
</tr>
<tr>
<td>11. 3 x 3 trace f-xy deconvolution filtering</td>
</tr>
<tr>
<td>12. Muting of air- and ground-waves</td>
</tr>
</tbody>
</table>

By tracking a target prism fixed to a mast attached to the recording sled, the theodolite provided coordinates as the GPR data were acquired. Measuring tapes and rope were used to guide the sled along approximately parallel straight lines oriented perpendicular to fault strike (i.e., along the x-direction of our local coordinate system in Figure 5.2b). We used a line spacing of 0.25 m for regions F1 and F2. After reviewing the recorded data, the line spacing was increased to 0.5 m for regions F3-F6. Coordinate assignments, corrections, and any required transformations were made during early stages of the processing (Table 5.2; Lehmann and Green [1999]). Trace interpolation and binning were achieved using a Delaunay triangulation method.

To obtain reliable estimates of subsurface GPR velocities, we recorded multiple common-midpoint profiles (CMP’s) within the survey area (black and white dots in Figure 5.2b). Where possible, two CMP’s were recorded at each location, parallel and perpendicular to fault strike. By measuring the normal moveout (NMO) of the more continuous reflections on each CMP, we obtained velocities of 0.065-0.077 m/ns within the surveyed regions, with an average value of 0.070 m/ns. The small variance in velocities reflects the relatively uniform stratigraphic sequence within the survey area. Temporal variations in GPR ground velocity, primarily due to periods of precipitation, were estimated by repeating measurements (approximately every 2 days) at a control point CMP (white dot in Figure 5.2b). The precipitation only affected GPR velocities very near to the surface. Consequently, to correct our 3-D GPR data for day-to-day velocity changes, it was only necessary to apply appropriate static shifts to the ungridded traces (see Appendix A in section 5.12).

After correcting for day-to-day velocity variations, all data were scaled by divid-
5.5 Migrated 3-D GPR Data

Horizontal slices extracted at different depths from the processed data volumes illustrate the influence of faulting on subsurface reflections (e.g., see Figure 5.3). Vertical offsets on the faults have juxtaposed different stratigraphic reflection patterns. For example, reflections in Figure 5.3b tend to be more continuous on the hanging-wall side of fault strands s1 and s2 than on the footwall. In addition, tilting of the blocks between the faults has resulted in dipping stratigraphic reflections that intersect the horizontal cuts through the data volumes. Reflections between fault strands s2 and s4 dip to the south, oblique to fault strike, whereas reflections within the hanging-wall and footwall blocks of strand s1 parallel fault strike (Figure 5.3b).

Block rotations between faults are best observed on fault-perpendicular cross-sections extracted at different locations along strike (e.g., Figures 5.4 and 5.5). Cross-sections A-A’ and B-B’ also show how the structures can change markedly over just a few tens of meters. On profile B-B’, fault strands s1-s4 cut strata at semi-regular intervals (25-40 m) and dip steeply to the northwest (Figure 5.5). Structures are more irregular on profile A-A’, where fault strand s6 dips steeply to the southeast forming the northwest side of a small asymmetric graben (Figure 5.4). Fault strands s4 and s8 may be connected, but the large gap between data volumes prevents us from determining their relationship (Figures 5.3a, 5.3b, 5.4b, and 5.5b). The individual faults in the near surface have slightly listric geometries, suggesting they may merge at depths of <100 m.

On all cross-sections extracted from the data volumes, we note three prominent reflections H1-H3 that can be followed across and along fault strike (e.g., Figures 5.4 and 5.5). The reflections are shallow dipping, with a progressive increase in dip from H1 to H3 that indicates that the fault-bounded blocks underwent progressive rotation as sediments were being deposited.
Figure 5.3: Horizontal slices extracted from the processed and migrated GPR volumes at depths of (a) $z = 6.3\, \text{m}$ and (b) $z = 9.1\, \text{m}$. The blank and low intensity regions in (a) result from the horizontal slice cutting the Earth's surface, and the muted air- and groundwaves. The low intensity regions in (b) are a consequence of low signal-to-noise data that are automatically attenuated during the processing. Normal fault segments $s_1$-$s_{10}$ are shown in (a) and (b). The origin of the local $(x, y)$ coordinate system is shown by the ‘+’; the $x$-axis is approximately parallel to fault strike. Location of the Hutzader trench is also outlined.
Figure 5.4: Fault-perpendicular cross-sections extracted from the processed and migrated GPR volumes along profile A-A’ in Figure 5.3b. (a) Section without interpretation. (b) Interpreted section. Vertical exaggeration is 3. Prominent reflections H1, H2, and H3 (dashed and dotted lines) can be traced across each fault segment (solid lines). Less continuous fault segments are denoted by '?' symbols. (c) The same cross-section as in (b), but plotted without vertical exaggeration. The x and y coordinates are relative to the fixed point shown in Figures 5.2b and 5.3.
Figure 5.5: As for Figure 5.4, but for profile B-B’ in Figure 5.3b.
5.6 Correlation with trench data

A paleoseismic trench excavated prior to our GPR survey has allowed us to correlate the GPR reflections with the subsurface geology (Figures 5.2b and 5.3). The Huffadine paleoseismic trench was approximately 10 m long and extended to depths up to 5 m. We compared trench observations with two parallel GPR cross-sections extracted from one of the GPR volumes at ~1 m offsets to the northeast and southwest of the northeast and southwest trench walls, respectively (cross-sections closer to the trench were influenced by reflections/diffractions from the trench walls). A comparison of the geology exposed on the northeast wall and the nearby GPR cross-section is shown in Figure 5.6 (see also Appendix B in section 5.13).

5.6.1 Trench stratigraphy

The stratigraphic sequence mapped in the Huffadine trench, which is observed at other locations within the Taupo Rift, represents a depositional history from more than 25 ka to the present [Froggatt and Lowe, 1990; Villamor and Berryman, 2001; Nairn, 2002; Alloway et al., 2007]. Multiple closely spaced (<0.5 m intervals) and upward splaying fault strands are observed in the trench (Figure 5.6b). They form an ~2-m-wide zone of complex deformation. The multiple fault strands within this zone are represented on the GPR cross-sections by the single s1 fault strand (Figures 5.2b, 5.3b, 5.4b and 5.5b).

Sediments exposed in the trench comprise alternating layers of fine and coarse grain fluvial sands and gravels overlain by a succession of volcanic tephra deposits and their associated paleosols (Figure 5.6b). The fluvial units were deposited in streams and rivers at the end of a period of lacustrine sedimentation that began before ~64 ka [Villamor and Berryman, 2001]. Volcanic airfall deposits subsequently mantled this surface. Within the trench, the Te Rere tephra is the lowest of a sequence of tephas deposited during episodic volcanic eruptions within the TVZ since ~25 ka. The tephas include all phases associated with a volcanic eruption (e.g., ignimbrite, fall deposits). They are separated by paleosols that represent considerable periods of quiescence [Shane, 2000]. Individual tephra-paleosol pairs are usually easy to identify in cross-section (Figure 5.6a). As a consequence of weathering, paleosols usually contain slightly higher percentages of clay and silt than their parent material. Calibrated radiocarbon ages (cal. kyr BP) for the tephas identified in the trench are provided by Alloway et al. [2007].

Correlation of reflections with well-dated geological units in the Huffadine trench allows ages to be confidently assigned to GPR horizons H1, H2, and H3. By progressively removing the deformation observed on the two trench walls, we see evidence for seven single-event displacements that have variously offset the different stratigraphic horizons (Table 5.3; full details of the northeastern trench wall restoration are contained in section 5.13). Large offsets between closely spaced strati-
Figure 5.6: Correlation of GPR data with observations on the northeastern wall of the Huffadine paleoseismic trench. (a) Photo showing the lithologic contrast between the base of the Taupo tephra (coarse grained sediment) and the top of the pre-Taupo-eruption paleosol (fine grained sediment). (b) Geologic log from northeastern wall of paleoseismic trench (location shown in Figures 5.2 and 5.3). Arrows identify lithologic contacts correlated with GPR reflections. (c) Outline of trench log superimposed on nearby GPR cross-section. Yellow lines follow lithologic contacts correlated with GPR reflections (see (b)). Airwave and groundwave phases obscure reflections from parts of the H1 horizon within the footwall. (d) As for (c), but without interpretation.
5.6. Correlation with trench data

<table>
<thead>
<tr>
<th>Event</th>
<th>Cumulative Displacement</th>
<th>Age</th>
<th>Reflection</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.15 ± 0.10</td>
<td>0.9 ± 0.8</td>
<td>H1</td>
</tr>
<tr>
<td>2</td>
<td>0.58 ± 0.10</td>
<td>3.6 ± 1.9</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>0.71 ± 0.10</td>
<td>7.5 ± 2.0</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>1.22 ± 0.10</td>
<td>12.5 ± 3.0</td>
<td>H2</td>
</tr>
<tr>
<td>5</td>
<td>1.28 ± 0.10</td>
<td>18.6 ± 3.2</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>1.99 ± 0.20</td>
<td>18.6 ± 3.2</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>2.46 ± 0.20</td>
<td>24.4 ± 0.8</td>
<td>H3</td>
</tr>
</tbody>
</table>

Table 5.3: Cumulative displacements and approximate ages for seven surface-rupturing earthquakes determined from trench observations and GPR reflections H1, H2, and H3. Reflection H1 was only affected by event 1, whereas H2 was affected by events 1-4 and H3 was affected by all events.

graphic horizons are interpreted to represent an earthquake, with the timing of rupture constrained by the ages of the bounding stratigraphic horizons (e.g., Nicol et al. [2006]; Villamor et al. [2007]). Using the single-event displacements identified during restoration of the trench, we derive estimates for the displacement ages of each major GPR reflection horizon (Table 5.3).

The contact between the base of the Taupo tephra (1.7 ± 0.1 cal. kyr BP; Figures 5.6b and 5.17a; Sparks et al. [1995]; Alloway et al. [2007]) and the top of a paleosol that formed above the Whakatane tephra correlates with the strong H1 reflection observed in our GPR data (Figures 5.6b-5.6d). We cannot determine from the geology a minimum age for faulting that offset this contact (see event 1 of Appendix B in section 5.13). Nevertheless, historical records indicate that there have been no surface rupturing earthquakes in this region since the beginning of widespread European settlement in New Zealand at ~1850 AD (i.e., 0.1 cal. kyr BP). Based on this information, we assign a tentative age of 0.9 ± 0.8 cal. kyr BP (i.e., the mean of the two bounding ages ± the difference between the mean and the bounding ages) to the displacement of reflection horizon H1 (Table 5.3).

The basal contact of the Rotorua tephra (15.4 ± 0.3 cal. kyr BP; Figures 5.6b and 5.17a; Hajdas et al. [2006]; Alloway et al. [2007]) coincides with reflection horizon H2 (Figures 5.6b-5.6d). Within the trench, Rotorua tephra consists of fine to coarse ash and hosts lapilli clasts of up to 0.2 m in diameter. It is overlain by the Rotoma tephra and paleosol (9.5 ± 0.1 cal. kyr BP; Hajdas et al. [2006]; Alloway et al. [2007]). Fault displacement that occurred between deposition of these layers and displaced reflection horizon H2 is assigned an approximate age of 12.5 ± 3.0 cal. kyr BP (Table 5.3).

Where exposed in the Huffadine trench, the base of the Te Rere tephra airfall deposit (24.4 ± 0.8 cal. kyr BP; Figures 5.6b and 5.17a; Reimer et al. [2004]; Alloway et al. [2007]; Lowe et al. [2008]) coincides with reflection horizon H3 (Figures 5.6b-5.6d). This unit comprises a mixture of tephra and loess silt. It lies unconformably
above the alluvial beds. From the trench log, different amounts of displacement between the tops and bottoms of the offset Te Rere tephra layer are observed, indicating that faulting occurred during deposition of this layer (Figure 5.6b; see Event 7 of Appendix B in section 5.13). Accordingly, we assign a 24.4 ± 0.8 cal. kyr BP age to the initial displacement of reflection horizon H3 (Table 5.3).

5.7 Measurements of fault displacement

A seismic interpretation package was used to pick reflection horizons H1 - H3 throughout the GPR volumes. Their high continuity and distinctive waveforms allowed them to be picked semi-automatically on almost every trace, across and along fault strike. Based on the picked horizons, we have measured individual cumulative displacements along all fault strands transecting the GPR volumes.

Our GPR horizon maps show details of past fault deformation that are not evident at the surface. For example, when rendered in 3-D, picked reflection horizon H2 reveals evidence for the partial collapse of an over-steepened footwall of fault strand s1 and a local horst structure that formed between fault strands s2 and s5 (Figure 5.7). Partially collapsed footwalls have been observed in trenches excavated elsewhere within the Taupo Rift. They usually occur where the fault steepens towards the surface [Villamor et al., 2006]. The horst structure illustrated in Figure 5.7 coincides with a change in structural style from northwesterly dipping faults in the southwestern part of the survey area to southeasterly dipping faults in the northeastern part. The southeasterly dipping faults s5 and s6 form the northwestern side of an emergent graben (Figure 5.7).

Displacement estimates were made on fault-perpendicular cross-sections every 0.5 m along the length of each fault. Like the trench logs (Figure 5.6b), we observed multiple offsets of the GPR reflection horizons beneath the topographically defined fault scarps. Where they could be identified, we summed the vertical separations of reflection offsets to obtain the total throw for each horizon (e.g., Figure 5.8). Because the faults were steeply dipping in the near surface (70-90°) and mostly had pure dip-slip components, the throws approximated the fault displacement.

Although the alluvial and volcanic deposits observed in the trench provide excellent chronostratigraphic horizons from which to measure past displacement, they have been subject to a complex interplay of tectonic, depositional, and erosional processes that pose challenges for our paleoseismic interpretation. Interpretations based on the airfall tephras are particularly problematic, because they are observed elsewhere to mantle topography rather than form horizontal layers within the accommodation space created by fault offset [Villamor et al., 2006, 2007]. Consequently, it is difficult to discriminate between sedimentary layers that mantle a pre-existing fault scarp and those that are folded in association with fault movement. For this study, we measure only the brittle deformation that could be observed across each fault strand and attribute any horizon curvature to depositional
Figure 5.7: Picked reflection horizon H2 showing evidence for partial collapse of the footwall of fault strand s1 and a local horst structure formed between strands s2 and s5. Black lines outline the projection of the trench excavation on to the horizon. Vertical exaggeration is 2.
processes. Given that we neglect the influence of fault-drag folding and block rotations, our measurements are likely to be minimum estimates of past fault displacements. We assume that any erosion of the footwalls is insignificant, because thin tephra layers are preserved on both sides of the trenched fault. Since we are investigating a relatively thin (<10 m) stratigraphic sequence near the surface, differential compaction of the footwall and hanging-wall sediments is also considered to be negligible.

A mean smoothing filter (with length \( n = 20 \) samples) is applied to all displacement profiles to reduce scatter due to measurement error. We estimate that the vertical resolution of our GPR data and measurement errors contribute to displacement uncertainties of ± 0.3 m at each sample point. Although the downthrown parts of horizon H1 are clearly imaged within the hanging-walls of the faults, its superposition with the airwaves and groundwaves makes it difficult to pick within the footwalls of the faults (Figures 5.6c and 5.6d). Therefore, displacement measurements for this horizon are probably systematically underestimated.
5.8 Results

5.8.1 Cumulative fault-displacement profiles

Displacement measurements for fault strand s1 are shown in Figure 5.9a (this is the only fault strand for which we have trenching information within our survey area). On this and all other fault-displacement diagrams, measurements are plotted on a N049°E axis. This direction coincides with the y-axis of our GPR surveys and approximates the average strike of the faults (Figure 5.3). The displacement profiles for H1, H2, and H3 represent the systematic accumulation of earthquake slip on this fault strand over the past $0.9 \pm 0.8$, $12.5 \pm 3.0$, and $24.4 \pm 0.8$ cal. kyr BP, respectively (Table 5.3).

The profiles for H2 and H3 are approximately symmetric; displacements are
largest near the profile centers (y = 75 m) and decrease towards the ends of the profile. We place the location of the northeast fault tip at y = 175 m, where vertical offsets of the GPR reflection horizons taper to zero. Although we have not sampled the southwest fault tip, simple extrapolation suggests that the entire length of the fault is roughly 200 m (Figure 5.9a). The short-wavelength variations superimposed on the broader displacement patterns may be due to off-fault folding. Alternatively, local minima and maxima may have been inherited from smaller strands that linked together as the fault grew (e.g., McLeod et al. [2000]). The H1 displacement profile for this fault strand has a displacement maximum well away from the center of the fault and is relatively flat elsewhere.

The H1, H2, and H3 horizon displacements derived from retro-deformation of the trench logs are also plotted in Figure 5.9a (see Appendix B in section 5.13). Trench observations suggest that an earthquake displaced the H1 horizon by 0.15 m at approximately 0.9 ± 0.8 cal. kyr BP (Table 5.3). Yet the GPR data demonstrate that the rupture on this same horizon attained a maximum displacement of ~1 m at y = 125 m (Figure 5.9a). From the trench, we determine that the H2 horizon is offset by at least four single-event displacements, causing 1.22 m of cumulative displacement since 12.5 ± 3.0 cal. kyr BP. In contrast, the GPR data reveal a displacement maximum of 2.9 m at y = 75 m. Although at least seven single-event displacements have offset horizon H3 since 24.4 ± 0.8 cal. kyr BP (see Appendix B in section 5.13), the limited trench dimensions do not allow us to measure the total offset (Figure 5.6b). For this reason, it is no surprise that the trench-derived cumulative displacement of 2.46 m is less than the ~2.7 m indicated by the GPR data at the same location. The GPR data reveal a maximum displacement for H3 of 4.3 m at y = 75 m (Figure 5.9a).

A second example of a displacement profile is presented in Figure 5.9b; fault strand s2 displaces reflection horizons H1, H2, and H3 by maximum amounts of 1.1, 3.8, and 4.1 m, respectively. Despite having a shorter length, the s2 displacement maxima are comparable to the estimates for s1. Displacements of horizons H2 and H3 decrease rapidly to the northeast on strand s2, from ~4 m at y ~90 m to 0 m at y ~150 m (Figure 5.9b). This pronounced change coincides along strike with increases in displacement on overlapping strands s5 and s6 (Figures 5.3, 5.7, and 5.10a). Thus, our data indicate that displacement has been transferred between the overlapping strands, even though they dip in opposite directions and do not clearly link in the shallow subsurface.

To understand better the collective behavior of the faults, Figure 5.10b shows along-strike profiles of the combined cumulative displacement across all faults observed within the GPR volumes. Displacement profiles for the H2 and H3 horizons have similar geometries, with displacement maxima at around y = 100 m (Figure 5.10b). In contrast, the displacement profile for H1 does not have a single well-defined maxima.

The profiles are somewhat asymmetric, primarily because only partial sections
5.8. Results

Figure 5.10: (a) Map of all fault strands offsetting the H3 reflection horizon. The thickness of each fault strand represents the displacement along strike. Dotted lines outline the areas of the six GPR regions; faults have been interpolated across small (<10 m) gaps between these regions. (b) Cumulative displacement profiles for each reflection horizon measured along all fault strands. (c) as for (b), but only for strands s1, s2, s5, s6, and s7.
of some faults are imaged (Figure 5.10a). To remove the influence of incomplete fault strands, we have excluded strands s3, s4, s8, s9, and s10 from the displacement profiles shown in Figure 5.10c. As a result, the H2 and H3 displacement profiles in Figure 5.10c are more symmetric than those in Figure 5.10b, but the H1 horizon continues to have an irregular shape (Figure 5.10c).

The lack of correlation between the individual and combined-fault-displacement profiles for the youngest (H1) and oldest (H2 and H3) horizons suggests that the displacement distribution that accumulated over relatively recent times (≤0.9 ± 0.8 cal. kyr BP) may not be characteristic of the long-term displacement patterns. It appears that multiple earthquakes are needed for faults to acquire characteristic displacement profiles (i.e., with displacement maxima located near their centers and tapering towards their tips; Figures 5.9 and 5.10c).

5.8.2 Influence of location on displacement measurements

The individual and cumulative displacement profiles show how fault displacement can change over a few meters along fault strike (Figures 5.9, 5.10b, and 5.10c). In view of these observations, displacements measured at a single location may not be representative of displacements along the entire fault. Figure 5.11 illustrates displacement histories for fault strand s1 determined using three different methods:

1. cumulative displacements measured within the trench (blue line),
2. cumulative displacements determined from the average values of the GPR-derived displacement profiles (red line),
3. cumulative displacements determined from the maximum values of the GPR-derived displacement profiles (green line).

Given that the average values account for along-strike variations, we consider them to be the most representative of fault-displacement history. Since the trench was located near the southwest tip of fault strand s1, displacements measured at this position are up to 0.5 m smaller than the average of displacements measured along the length of the fault (Figure 5.11). Measurements made at the displacement maxima are >1 m greater than the average values, yielding significant over-estimates of the displacement-accumulation rates.

5.8.3 Interval slip rates

The fault-displacement profiles reveal considerable variability over time scales of ≤12.5 kyr. To gain further insight into the spatial and temporal evolution of the faults, we convert displacements measured on the dated H1, H2, and H3 reflection horizons to interval slip rates. Figure 5.12 shows slip rates for fault strand s1 (solid
5.8. Results

Figure 5.11: Plot of cumulative displacement versus horizon age for fault strand s1. Blue line (with error bars) shows cumulative displacements determined by retro-deforming stratigraphy observed in the Huffadine paleoseismic trench (Appendix B in section 5.13). Red and green lines (with error bars) show cumulative displacements derived from the GPR data calculated from the means and maxima of the H1, H2, and H3 displacement profiles shown in Figure 5.9a.
Table 5.4: Ratios of average slip rate for the three independent time intervals constrained by horizons H1-H3 relative to the long-term (0 - 24.4 cal. kyr BP) average slip rate for each fault strand (average slip rates are shown in Figure 5.13).

<table>
<thead>
<tr>
<th>Time interval (cal. kyr BP)</th>
<th>s1</th>
<th>s2</th>
<th>s5</th>
<th>s6</th>
<th>s7</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 - 0.9</td>
<td>2.21</td>
<td>2.46</td>
<td>1.65</td>
<td>7.03</td>
<td>0.56</td>
</tr>
<tr>
<td>0.9 - 12.5</td>
<td>1.11</td>
<td>1.58</td>
<td>1.27</td>
<td>1.14</td>
<td>0.96</td>
</tr>
<tr>
<td>12.5 - 24.4</td>
<td>0.80</td>
<td>0.33</td>
<td>0.69</td>
<td>0.41</td>
<td>1.07</td>
</tr>
</tbody>
</table>

line) and a combination of the more completely sampled strands s1, s2, s5, s6, and s7 (dotted line) for six different time intervals. In general, the plots illustrate how the slip-rates become less variable over progressively longer time intervals. The slip-rate profiles for fault strand s1 and the combined fault strands over the past ~0.9 kyr are highly irregular (Figure 5.12a). At some locations, the slip rates are zero, indicating incomplete fault rupture of the strands, whereas at other locations they are up to 5 times larger than those averaged over the past 24.4 kyr (Figures 5.12a and 5.12f). For time intervals ≥11.6 kyr, the largest slip rates near the centers of the profiles are quite close to their long term averages (Figures 5.12b-5.12f).

Although the profiles shown in Figures 5.12b and 5.12c are calculated for similar time intervals of 11.6 and 11.9 kyr, they exhibit different patterns and rates of slip. Maximum slip rates calculated for the younger time interval (0.9-12.5 cal. kyr BP) are approximately twice as high as those from the older time interval (12.5-24.4 cal. kyr BP), suggesting that slip rates have increased with time. To analyze temporal changes in slip rate on strands s1, s2, s5, s6, and s7, Table 5.4 shows the ratios of their average slip rates over the three non-overlapping time intervals (Figures 5.12a, 5.12b, and 5.12c) relative to their long-term average slip rates (Figure 5.12f); average slip rates are based on the average values of displacement measured along the fault strands. With the exception of strand s7, all faults show a progressive increase in slip rate with time. Although these results suggest that slip rates are increasing with time, they may also reflect short-term variability in the accumulation of fault displacements.

We have also calculated average slip rates for strands s1, s2, s5, s6, and s7 for all six time intervals shown in Figure 5.12. The values in Figure 5.13 reveal generally decreasing variances in average slip rate with increasing time interval. Values for the shortest time interval are widely distributed, ranging from 0.02 to 0.53 mm/yr. Although the range is much smaller for the 11.6, 11.9, and 12.5 kyr time intervals, there is considerable scatter in the values for individual strands. For example, average slip rates for strand s6 vary from 0.03 to 0.12 mm/yr. In contrast, average slip rates for the 23.5 and 24.4 kyr time intervals are very similar to each other (Figure 5.13).

As for the spatial variations of fault displacement, our results suggest that the Maleme faults require significant periods of displacement accumulation before rel-
5.8. Results

Figure 5.12: Slip rate variations along fault strands s1, s2, s5, s6, and s7 for six different time periods. Gray lines are slip rates for strand s1. They are calculated using the H1, H2, and H3 displacement profiles shown in Figure 5.9a; dotted lines are combined slip rates for strands s1, s2, s5, s6, and s7 based on the combined H1, H2, and H3 displacement profiles shown in Figure 5.10c. Time intervals are constrained by the horizon ages shown in Table 5.3 (see also Figure 5.17a).
5.9 Discussion

5.9.1 Comparison with regional-scale studies within the Taupo Rift

From the combination of trench and GPR data, we have estimated displacements along five practically complete fault strands (s1, s2, s5, s6, and s7 with lengths ranging from ~25 to ~200 m) within a limited area of the Taupo Rift. Neither the location of the displacement maximum for the youngest reflection horizon H1 nor the slip rates of all three fault-displaced horizons for time intervals \( \leq 12.5 \) kyr have consistent patterns. These observations are in general agreement with the results of two regional-scale studies within the Taupo Rift [Nicol et al., 2006; Bull et al., 2006]. Nicol et al. [2006] demonstrate that displacement rates along several widely distributed faults are highly variable for time scales \(< 18\) kyr, but that rates become more uniform when averaged across fault zones. Similarly, Bull et al. [2006] show that displacement patterns and slip rates along the Rangitaiki fault in the offshore Taupo Rift are only uniform for time scales \( > 9 \) kyr. The general agreement between
these studies and ours suggests that both short (<200 m) and long (>1 km) faults require several thousand years to establish uniform displacement patterns and slip rates.

5.9.2 Implications for seismic hazard estimates

Since the GPR-based average fault displacement values account for highly variable slip distributions, they are a basis for more reliable estimates of past earthquake magnitudes using standard empirical formulas (e.g., Wells and Coppersmith [1994]; Hemphill-Haley and Weldon [1999]). Furthermore, GPR-based extrapolations of trench observations along and across fault strike provide key information on the distribution of ground rupture along kinematically-linked fault strands. For example, from trench observations of strand s1 we would estimate a single-event displacement of only 0.15 m across the youngest reflection horizon H1, which would indicate that the last earthquake rupture had a relatively small magnitude. In contrast, the GPR data reveal that this earthquake probably ruptured fault strands s1, s2, s5, s6, and s7 simultaneously and produced displacements as large as 1.7 m (Figure 5.10c).

5.9.3 Extension rates

As well as vertical slip rates, we have derived extension rates from measurements of fault heave. The maximum cumulative 4.9 m of heave on the oldest reflection horizon H3 was observed along profile B-B’ as it crossed fault strands s1-s4 (Figure 5.5). It corresponded to an average extension rate of 0.20 ± 0.01 mm/yr over the past ~24.4 kyr. For historical earthquakes, empirical relationships for normal faults suggest that displacements at seismogenic depth are on average 1.6 times larger than at the surface [Wells and Coppersmith, 1994; Villamor and Berryman, 2001]. Applying this correction factor to our estimates yields an extension rate of 0.32 ± 0.02 mm/yr. Moreover, small earthquakes (< M 5.8) that do not produce ground rupture within the Taupo Rift are probably responsible for (non-observed) extension that is roughly 30 % of the observed extension [Villamor and Berryman, 2001; Lamarche et al., 2006]. Combining this information, we estimate that the maximum horizontal displacement within our GPR survey area is 0.43 ± 0.03 mm/yr, which represents ~5 % of the total extension budget across the central Taupo Rift estimated from geodetic observations and other geological information [Wallace et al., 2004b; Nicol and Wallace, 2007].

In order to obtain extension for the complete fault zone, one would have to measure slip rates from all faults along strike-perpendicular transect lines across the entire width of the fault zone. Considering the along-strike variability of fault displacements observed in our 3-D GPR data, we caution against projecting slip rates onto a transect line that are derived from trenches at different locations along
fault strike. Although we have not been able to measure extension across the entire width of the Maleme fault zone, we suggest that it could be achieved using the combined GPR/trenching approach outlined here.

5.10 Conclusions

Our GPR data enabled us to define the geometries of three faulted reflection horizons H1, H2, and H3 that were first identified in a paleoseismological trench. These high resolution GPR data provided vivid images of complex fault deformation not evident at the surface and allowed fault-displacement patterns along ten fault strands s1-s10 to be mapped in the shallow subsurface over a ~150 x 250 m area of the Maleme fault zone. Fault strand s1 intersected the trench, in which a stratified sequence of tephras and their paleosols were seen to overly fluvial sands and gravels. From the fault-displacement patterns, we computed cumulative, average cumulative, and maximum cumulative displacements for all ten fault strands. By combining the displacement information with the ages of the three principal reflection horizons determined from trench data (0.9 ± 0.8, 12.5 ± 3.0, 24.4 ± 0.8 cal. kyr BP), we were then able to estimate space and time-varying slip rates.

GPR-based cumulative displacement estimates for fault strand s1 at the trench were almost identical to cumulative displacements determined by retro-deforming the geological trench logs. However, GPR-based values of the average and maximum cumulative displacements along the ~200 m length of the fault strand were up to 0.5 m and 1.5 m larger than the cumulative displacements measured at the trench, respectively. As a consequence, using fault displacements defined at the trench would have yielded anomalously low estimates of slip rate and earthquake magnitude. Our analyses have illustrated the wide range of displacements that could be observed at an arbitrary point along a fault strand within the Maleme fault zone; multiple trench excavations would be required to determine the average and maximum cumulative displacements of the faults.

Cumulative displacement profiles for all faults were highly irregular for the past 0.9 ± 0.8 cal. kyr BP. For horizons with ages ≥12.5 ± 3.0 cal. kyr BP, progressive accumulation of slip from multiple (≥4) earthquakes produced characteristic cumulative displacement distributions with maxima located near the centers of the faults. In addition, interval slip rates measured from five practically complete fault strands became uniform for time intervals >12.5 kyr. These results suggest that individual faults within the Maleme fault zone require >12.5 kyr to establish uniform displacement patterns and slip rates.
5.11 Acknowledgements

We greatly appreciate the contributions of Nima Riahi and our New Zealand-based field assistants: Nick Horspool, Anne Douglas, and Gemma Salm. This project was supported by grants from the Swiss National Science Foundation and ETH Zurich. Funding for trench data and Villamor’s contribution from New Zealand Foundation for Research, Science and Technology (contract C05X0402) and Royal Society of New Zealand Marsden Fund (contract GNS303).

5.12 Appendix A: Application of day-to-day static corrections

Very near-surface velocity variations resulting from rainwater infiltration created static shifts in our GPR data. This effect is illustrated in Figure 5.14a. Traces between \( y = 5 \) m and \( y = 55 \) were recorded over a three day period without rain. Overnight rainfall then reduced near-surface velocities such that the groundwave and reflected phases recorded on subsequent days were delayed, creating the artificial variation in amplitudes along the \( y \)-axis of the time slice; at \( y = 55 \) m amplitudes flip from negative to positive and then gradually decrease toward \( y = 140 \) m (Figure 5.14a).

Repeat CMPs were recorded at our control point, before and after the overnight rainfall event (Figures 5.2b and 5.15). They revealed a change in the groundwave phase velocity of 0.005 m/ns and differences in reflection phase arrivals of \( \sim 7 \) ns. Nevertheless, normal-moveout velocities measured from reflection phases were identical for each, indicating that the velocity changes had mainly occurred near the surface (0.07 m/ns; Figures 5.15b and 5.15d).

A plot of each days mean trace (calculated from all traces recorded on each survey day) illustrates the variation in arrival times of the groundwave phase; a sudden delay resulting from the rainwater infiltration followed by progressively earlier arrival times as the ground moisture content returns to pre-rainfall levels (Figure 5.16). We measured the groundwave delay time for each day by cross-correlation of the mean traces. Using these time lags, a single static shift was applied to the ungridded traces collected on each survey day. On the corrected time slice the linear discontinuity at \( y = 55 \) m has disappeared along with the progressive decrease in amplitudes from \( y = 55 \) m to \( y = 140 \) m (Figure 5.14b).
Figure 5.14: Time slice extracted at 116.0 ns (~2.7 m depth) from data volumes (a) before and (b) after the application of static corrections that account for very near-surface velocity variations on each survey day.
5.12. Appendix A: Application of day-to-day static corrections

\[ V_g = 0.080 \text{ m/ns} \]

\[ V_g = 0.075 \text{ m/ns} \]

Figure 5.15: (a) CMP measured at the control point shown in Figure 5.2b. Solid line is the picked groundwave phase \((V_g)\) and dashed lines are picked reflections. (b) Corresponding semblance plot for which warmer colors represent high values. White crosses define the NMO velocities for the two picked reflections shown in (a). (c) and (d) are as for (a) and (b) but for data recorded after a period of rainfall.
Figure 5.16: Plot of mean traces calculated for each survey day showing the variation of groundwave arrival times. Dashed line shows the arrival time of the groundwave peak for survey day 7. Numbers in boxes are the time lags (in ns) calculated by cross-correlating each mean trace with the mean trace for day 7.

5.13 Appendix B: Huffadine trench restoration

5.13.1 Introduction

Here, we present details on the retro-deformation of stratigraphy exposed on the northeastern wall of the Huffadine paleoseismic trench (Figure 5.17a). Restoration of the southwestern wall is not described here, but mean values of cumulative displacement from both walls are listed in Table 5.3 of the manuscript. Retro-deformation is achieved by treating the tops of paleosols (or the bases of the non-weathered tephras above the paleosols) as ancient ground surfaces. By progressively restoring these surfaces, we realign common stratigraphic units to a time before they were faulted. The ages of the well-dated tephras bracket the movement ages. In several instances, the restoration of one prehistoric ground surface aligns several older tephra-paleosol pairs, suggesting that there were no fault ruptures within those intervals.

The total dip-slip offset required to restore each prehistoric ground surface is interpreted as a single-event displacement. We estimate that most displacements measured from the restoration are accurate to $\pm 0.1$ m. In some cases, we have assigned larger uncertainties (see below).

Fault planes observed in the Huffadine trench are very steep with variable dips, which results in:
Figure 5.17: (Continued on the following two pages) (a) Detailed log of the Huffadine northeastern trench wall (a simplified version is shown in Figure 5.6b). Unit ages (in kyr) are listed in brackets; they are based on radiocarbon dating reported by Alloway et al. [2007]. (b) to (h) Results of removing the effects of successively older slip events (i.e., 7 single-event displacements). Information on the restorations is provided in the text of Appendix B in section 5.13. Grid coordinates are in meters.
Restoration 2: Base of Whakatane

Restoration 3: Base of Rotoma paleosol

Restoration 4: Base of Rotorua
(f) Restoration 5: base of paleosol on tephric loess

(g) Restoration 6: base of tephric loess

(h) Restoration 7: Base of Te Rere
1. rotation of blocks between the fault splays,

2. local exaggeration of fault throw between some fault splays (responding to the creation of open areas on the fault planes),

3. toppling of parts of the over-steepened footwall block that rotates the fault plane and creates the geometry of an "apparent reverse fault" [Villamor et al., 2006].

Restorations that correctly account for these effects can be difficult. In some cases, we have ignored the smaller block displacements and restored the layer to fit the shape of the original fault scarp at that particular time.

One of the main characteristics of volcanic airfall deposits is that they mantle the landscape, such that the shape of a volcanic layer is influenced by the pre-existing fault scarp. In addition, unconsolidated tephra deposits may be dragged along the fault plane during displacement. As a consequence, the final shape of layers across a fault zone may represent a combination of the shape of the original scarp and the results of drag folding. Drag folding can be identified for some curved layers if the tops and bottoms of a tephra-paleosol pair are misaligned following restoration. Since the curved parts of horizons that are >1.5 m from the fault planes are not influenced by drag folding, we can use these parts to estimate the original scarp shapes. Furthermore, we take into account other possible fault-related features that may influence our estimates of timing and displacement, including local unconformities, colluvial wedges derived from fault-scarp erosion after earthquakes, fissures, and termination of faults at different layer boundaries.

For safety reasons, the trench was opened in two batters separated by a 1-m-wide bench. Because volcanic layers mantle the landscape, they may not have the same topographic elevation along fault strike. Moreover, normal faults cutting through unconsolidated volcanic layers tend to merge and splay over short distances, thus causing problems in the restoration of layers that transect both batters. For example, several faults observed on one batter may merge into fewer faults on the second batter. Where possible, we do not restore layers that transect different batters. For the restoration of layers older than the Tephric loess (unit 6b), we have to use layers that transect different batters. For these cases, we incorporate a larger uncertainty (± 0.2 m) in our displacement estimates.

In Figures 5.17a - 5.17h, we present the Huffadine trench logs with successive restorations of the stratigraphic sequence that were produced by back-stripping the fault displacements. Note, that several fault strands exposed on the trench walls (s1a-s1c in Figure 5.17a) are imaged as the single fault strand s1 in the GPR data. For simplicity, we only show diagrams of the pre-rupture stage for each event. Calibrated radiocarbon ages in "cal. kyr BP" for stratigraphic units are from Alloway et al. [2007]. In the following, the fault-movement ages are mostly constrained to lie between the ages of two tephra layers. For our slip-rate estimates, we have assumed
that the age of the fault event equals the average of the bounding ages plus/minus
the difference between the two values and the average.

5.13.2 Event 1

We first restore the base of the Taupo tephra (units 2b and 2c; Figure 5.17b). Fault
terminations for this event were not clear in the field. Although Figure 5.17a shows
the terminations (s1b(nw) and s1b(se)) that could be clearly identified, it is possible
that they extend to the base of the top soil unit (unit 1). To restore stratigraphy offset
by event 1, we align unit 2b across splays s1b(nw) and s1b(se), and close the fissure
on s1b(se). This also required the removal of some curvature of both splays of s1b
and rotation of units 3a to 5a in the block between the splays. From restoration 1,
we estimate a total dip-slip displacement of 0.16 ± 0.10 m at the trench site.

The displacement required to realign Taupo tephra does not bring deeper lay-
ers into alignment. In view of this, event 1 likely occurred after deposition of the
Taupo tephra at 1.7 cal. kyr BP and before widespread European settlement in New
Zealand, which began at ~1850 AD (0.1 cal. kyr BP). Accordingly, event 1 has an
assigned age of 0.9 ± 0.8 cal. kyr BP. Our GPR data suggest that this event could
have been large, with offsets as great as 1.7 m. Such displacements are usually
associated with major large destructive earthquakes.

5.13.3 Event 2

Because we were not able to distinguish differences in faulting between the Whaka-
tane tephra and its paleosol (units 3a and 3b, Figure 5.17c), restoration 2 was
achieved by aligning the base of the non-weathered Whakatane tephra (unit 3b). Restor-
ation involved aligning Whakatane tephra across both splays of s1b. Realign-
ment of the Whakatane tephra across s1b(nw) moved a block of unit 5b (wedged
between s1b(nw) and a short northwestern splay of s1b(nw)) to a position higher
than its piercing point on the footwall, suggesting that displacement was probably
partitioned between s1b(nw) and its northwestern splay (although this event did not
appear to rupture to the base of unit 3). We removed the toppling of s1b(nw) and
some rotation of layers close to it. A total dip slip of 0.42 ± 0.10 m was restored for
event 2, which occurred between deposition of the Whakatane tephra 5.5 cal. kyr
BP and deposition of the Taupo tephra 1.7 cal. kyr BP. Based on this information,
we assigned an age of 3.6 ± 1.9 cal. kyr BP to event 2.

5.13.4 Event 3

Restoration of the Rotoma paleosol (unit 4a) was achieved with a displacement of
0.06 ± 0.10 m on the northwestern splay of s1b(se) (Figure 5.17d). Timing of event
3 is bracketed between the ages of the Rotoma tephra (9.5 cal. kyr BP) and the
Whakatane tephra (5.5 cal. kyr BP), which yields an assigned age of 7.5 ± 2.0 cal. kyr BP. Except for a small residual displacement (~0.1 m) on s1b(se), restoration of the Rotoma paleosol realigned most blocks comprising unit 5b. However, because it was difficult to identify subtle lithological boundaries between units 5a, 5b, and 5c to the southeast of s1b(se), we assumed that this layer was fully restored and did not interpret an extra event between events 3 and 4.

5.13.5 Event 4

Realining the base of the Rotorua tephra across strands of s1b (unit 5c; Figure 5.17e) yielded restoration 4. The poorly defined boundary between the Rotorua tephra and its paleosol (unit 5a) southeast of s1b(nw) impeded restoration of the paleosol. For this reason, we restored the Rotorua tephra by aligning the layer southeast of s1c with its continuation northwest of s1b(nw). We were not able to retro-deform in detail the block between splays of s1b, because they were rotated and exhibited exaggerated throws (Figure 5.17e). Restoration 4 realigned most of the bases of the older Tephric loess paleosol (unit 6a), but some residual offset remained across strand s1c (Figure 5.17e). As a result, an earlier event likely displaced the Tephric loess paleosol before the Rotorua tephra was deposited. This requirement did not substantially change the restorations. For the restoration of the base of the Rotorua tephra, we attributed some additional displacement to drag folding. We interpret 0.62 ± 0.10 m of dip slip for restoration 4, which occurred between deposition of the Rotorua (15.4 cal. kyr BP) and Rotoma tephras (9.5 cal. kyr BP) at an assigned age of 12.5 ± 3.0 cal. kyr BP. A fault-scarp-derived colluvial wedge (units 5b) was formed by erosion of the footwall after event 4.

5.13.6 Event 5

Restoration 5 involved aligning the base of two blocks comprising the Tephric loess paleosol (unit 6a) across the southeastern splay of s1c (Figure 5.17f). The block of mixed Te Rere tephra between the two splays of s1c is reworked material. A change in dip of the fault plane created an open space, which was subsequently filled with Te Rere units (units 8a to 8d in Figure 5.17f; gray colored blocks show our interpretation of the original shape of these units). Restoration of these Tephric loess paleosol blocks required 0.12 ± 0.10 m of dip-slip displacement that did not realign older layers. This single-event displacement occurred before deposition of the Rotorua tephra at 15.4 cal. kyr BP and after the formation of the Tephric loess paleosol (unit 6a), which developed well after the deposition of the Okareka tephra at ~21.8 cal. kyr BP. Based on this information, we assign an age of 18.6 ± 3.2 cal. kyr BP to event 5.
5.13. Event 6

For restoration 6, the base of the Tephric loess (unit 6b) was aligned across the entire width of the exposed fault zone (Figure 5.17g). It required rotation of the Okareka tephra blocks (unit 7), closure of the fissure, and removal of Tephric loess drag folding (unit 6b) close to the faults. Dip-slip displacement resulting from restoration 6 was $0.71 \pm 0.20$ m; a larger uncertainty was assigned because we realigned layers across s1b that transected different trench batters. This single-event displacement predates the formation of the Tephric loess and its paleosol, which could not be assigned an age. Consequently, we bracketed this event between formation of the Rotorua (15.4 cal. kyr BP) and Okareka (~21.8 cal. kyr BP) tephras, thus yielding an assigned age of $18.6 \pm 3.2$ cal. kyr BP. Although we have assigned the same age to events 5 and 6, it is clear that event 5 occurred some time after event 6 (i.e. sufficient time for the deposition of the Tephric loess and formation of its paleosol). Restoration 6 realigned Te Rere paleosol (unit 8a), but not the older Te Rere tephra (units 8b, 8c, and 8d).

5.13.8 Event 7

Realignment of Te Rere tephra across s1b and the various splays of s1a (Figure 5.17h) yielded restoration 7. Several blocks had to be rotated, contributing to a total dip-slip displacement of $0.35 \pm 0.20$ m; a larger uncertainty was assigned to the realignment of the Te Rere tephra because it transected different trench batters. Since the GPR data indicated that some faulted blocks of the Te Rere tephra were not exposed in the Huffadine trench, we considered this value to be a minimum estimate. Our restoration revealed that displacement occurred during deposition of the Te Rere tephra unit. Because this unit was a mixture of the Te Rere tephra and younger loess deposits, it was possible that some of the displacement occurred after the end of the Te Rere eruptive episode. Nevertheless, we assigned an age of $24.4 \pm 0.8$ cal. kyr BP, which was constrained by radiocarbon dating of Te Rere tephra at other locations within the Taupo Rift [Alloway et al., 2007]. Restoration 7 did not bring older layers into alignment, suggesting that the alluvium was faulted before deposition of the Te Rere tephra. We determined that the surface of the alluvium was horizontal when the Te Rere Tephra was deposited (Figure 5.17h), which indicated that fluvial activity had eroded the scarp produced by earlier faulting.
Chapter 6

Assessing the contribution of off-fault deformation to slip-rate estimates within the Taupo Rift, New Zealand using 3-D GPR surveying and trenching

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submitted to Geology

6.1 Abstract

Active fault slip rates can be difficult to determine using conventional paleoseismic techniques, especially at locations where deformation extends many meters from the fault. We combine interpretations of 3-D ground-penetrating radar (GPR) and paleoseismic trench data to assess the contribution of off-fault deformation to displacement measurements made along and across a normal fault within the Taupo Rift, New Zealand. A GPR reflection horizon correlates with a fault-displaced 24.8 ± 0.4 cal. kyr BP alluvial surface observed in the trench. By measuring the geometries of the undeformed parts of this horizon away from the trench excavation, we show that drag folding and horizontal-axis rotations of the hanging-wall and footwall accommodate ~50% of the total extension across the fault zone. Such broad and complex deformation is difficult to observe at the surface or within trenches, which may explain why geologically determined fault slip rates for the central and southern Taupo Rift are anomalously low when compared to geodetic estimates. We suggest that a combination of GPR surveying and paleoseismic trenching may help
resolve differences between geodetically and geologically determined strain rates observed across active extensional regimes worldwide.

6.2 Introduction

To understand the seismic hazard posed by active faults, accurate estimates of their past slip rates are required. Unfortunately, fault-slip-rate data derived from different sources may be sparse, poorly constrained, and/or ambiguous. As a consequence, there are relatively few areas of the earth where fault-slip-rate data are suitable for seismic hazard analyses (see Cowie and Roberts [2001] for a review).

The introduction of dense GPS surveying networks has resulted in several attempts to reconcile decadal slip rates determined from geodetic measurements with millennial slip rates derived from the late Quaternary geological record (e.g., Wallace et al. [2004a]; McCaffrey [2005]; Zhang et al. [2007]). Discrepancies between the two data sources have been attributed to various factors including genuine differences between present-day and late Quaternary slip rates (e.g., Shyu et al. [2006]; Meyer and Le Dortz [2007]), unrecognized slip on hidden faults (e.g., Wallace et al. [2004b]; Nicol and Wallace [2007]), and undocumented errors in the geological estimates (e.g., Wallace et al. [2004a]).

In extensional environments, paleoseismologists can estimate slip rates from fault-displacement observations of well-dated stratigraphic markers within trench excavations (e.g., Pantosti et al. [1996]; Nicol et al. [2006]). Where deformation is not very complex, fault displacements can be accurately determined by measuring the offset of strata close to the fault plane. More complex deformation can be accounted for by retro-deforming the geological trench logs (e.g., Villamor et al. [2007]). Since off-fault deformation (e.g., subsidiary faults, drag folding, and block rotations) can extend across many meters of a fault zone, the total displacement field may not be observed within a typically small trench excavation. In addition, because displacements can vary along strike, a single trench excavation may yield displacement estimates that are not representative of the entire fault.

In the Taupo Rift of New Zealand (Figure 6.1a), geologically determined extension rates are up to 50% smaller than geodetic rates, suggesting that the paleoseismological data are incomplete and/or that some of the extension is accommodated by other processes such as dike intrusion [Villamor and Berryman, 2001; Wallace et al., 2004b; Nicol and Wallace, 2007]. To investigate the effect of off-fault deformation on estimates of recent (<25 ka) slip rates, we have undertaken a combined 3-D ground-penetrating radar (GPR) and trenching study of a normal fault strand within the Maleme fault zone (Figure 6.1b). The Maleme fault zone comprises a series of closely spaced fault strands that spans a ~2.5-km-wide graben within the central part of the Taupo Rift. (Figure 6.1).
Figure 6.1: (a) Plate boundary setting of the Taupo Rift and location of the Maleme fault zone (black dot). (b) Location of the three adjacent 3-D GPR surveys (shaded areas) and paleoseismic trench (circle). Fault trace data are from the interpreted GPR data and the GNS Science Active Faults Database (http://data.gns.cri.nz/af/). The x and y axes define the local coordinate directions of our GPR surveys. Geographic projection is New Zealand Map Grid.
6.3 Methods

Dense GPR data were acquired using a Sensors and Software PulseEKKO GPR unit linked to a self-tracking laser theodolite [Lehmann and Green, 1999]. Our survey covered an area of ~100 x 250 m and consisted of three adjacent regions (Figure 6.1); small gaps between these regions were a consequence of trees and fence lines. We used a broadside acquisition geometry (i.e., parallel antennas with their long axes perpendicular to the acquisition direction) with 1.0-m-offset 100 MHz antennas to record GPR traces (sampling rate 0.5 ns) along approximately parallel lines.

Coordinate assignments, data editing, data corrections, and any required transformations were made during early stages of the processing (cf., Lehmann and Green [1999]). Processing steps included trace binning and interpolation, time-zero corrections, amplitude scaling, and bandpass filtering. Multiple common-midpoint (CMP) surveys provided subsurface velocity information that was used to migrate the data and convert time to depth. We employed a 3-D migration algorithm that accounted for the modest topographic relief across the survey site [Lehmann and Green, 2000; Heincke et al., 2005]. Finally, the airwave and groundwave phases were muted.

At least three prominent reflection horizons H1, H2, and H3 were observed in the migrated 3-D GPR images (Figure 6.2). Fault-perpendicular cross-sections extracted from the 3-D GPR data revealed tilted reflections within the hanging-walls and footwalls of the faults. Near the faults, GPR reflections exhibited curvature over 5-10-m-wide regions, some of which may have been a consequence of fault drag (Figure 6.2).

We semi-automatically picked each horizon within the 3-D GPR volumes. The 3-D perspective of the deepest horizon H3 in Figure 6.2 illustrates the complex pattern of displacement produced by the closely spaced fault strands s1, s2, s5, s6, and s7. A paleoseismic trench excavated across fault strand s1 allowed us to correlate the reflections to a well-defined sequence of alluvium, tephra, and associated paleosol (Figures 6.2 and 6.3). Horizons H1 and H2 originated from two distinct tephra-paleosol boundaries across which there are notable changes in grain size, whereas horizon H3 originated from the prominent lithologic boundary between the alluvium and the Te Rere tephra (Figure 6.3).

6.4 Fault-displacement measurements

We have estimated displacement for each picked reflection horizon on fault-perpendicular GPR cross-sections every 0.5 m along the length of fault strand s1. Total throws for horizons H1-H3 were measured by summing the vertical separations of reflection offsets close to the fault plane (Figure 6.4). The steep dips (~70-80°) of the faults in the near surface and their pure dip-slip nature meant that the throws
Figure 6.2: Perspective view of cross-sections extracted from the migrated GPR volumes and semi-automatically picked reflection horizon H3. Black arrows - reflections H1 and H2; white arrows - fault strands; dotted lines - offset of H3 across fault strand s1. Black lines - projection of the trench excavation on to reflection horizon H3. Vertical exaggeration is 3.
Figure 6.3: Correlation of GPR data with observations from the paleoseismic trench. (a) Geologic log from northeastern wall of paleoseismic trench. Arrows identify lithologic contacts correlated with GPR reflections. (b) Outline of trench log superimposed on nearby GPR cross-section. Yellow lines follow lithologic contacts correlated with GPR reflections (see (a)).
were good approximations for the fault displacements. Limited vertical resolution and measurement errors contributed to displacement uncertainties of ± 0.3 m at each sample point.

The resulting GPR-based displacement profiles are plotted as solid lines in Figure 6.5a. Although the profiles exhibit characteristic fault-displacement patterns (e.g., Walsh and Watterson [1988]; Kim and Sanderson [2005]) with values that are consistent with the trench-derived estimates (see the pairs of triangles, circles, and squares in Figure 6.5a), they do not account for the effects of off-fault deformation. By measuring only the brittle deformation close to the fault plane, we neglect the potentially important influence of drag folding and horizontal-axis rotations of the hanging-wall and footwall in our displacement measurements. Consequently, we consider the solid lines in Figure 6.5a to be minimum values (H1_{min}, H2_{min}, and H3_{min}).

To estimate the total displacement field across the fault, we need to reconstruct the shapes of the stratigraphic surfaces before they were faulted. Because volcanic airfall deposits have a tendency to blanket the landscape, it is not easy to discriminate between tephra layers that have been folded and tilted in association with
Figure 6.5: (a) Minimum displacement measurements for GPR reflection horizons along fault strand s1 made from the GPR data ($H_{1\text{min}}$, $H_{2\text{min}}$, and $H_{3\text{min}}$; solid lines) and from the northeastern and southwestern walls of the paleoseismic trench (symbols); triangles are for $H_1$; circles for $H_2$; squares for $H_3$. $H_{3\text{total}}$ (dotted line) displacement measurements were determined from horizon $H_3$, but include the effects of off-fault deformation by using the technique illustrated in Figure 6.4. (b) The off-fault-displacement component determined by subtracting $H_{3\text{min}}$ from $H_{3\text{total}}$. 
fault movement and layers that have mantled topography produced by previous fault displacements. Accordingly, we cannot attribute curvature or tilting of the tephra-paleosol reflection horizons H1 and H2 to tectonic deformation. In contrast, reflection horizon H3 originates from the surface of an alluvium layer that was deposited during a period of river activity (Figure 6.3). From a detailed restoration of the paleoseismic trench log, we determine that this surface was practically horizontal as fluvial deposition waned. As a consequence, we can assume that all dip and much of the curvature exhibited by reflection horizon H3 are a result of fault displacements after the formation of this surface.

Total displacement across the H3 reflection horizon is estimated by projecting the undeformed planar parts of the horizon towards the fault plane (Figure 6.4; cf. Chapman and Meneilly [1991]; Mansfield and Cartwright [1996]). By measuring the vertical separation between the projection points (blue circles in Figure 6.4), we determine along-strike total displacement values for reflection horizon H3 ($H_3^{\text{total}}$; Figure 6.5a). Displacement uncertainties resulting from this method are $\pm 0.5$ m at each sample point. The component of displacement attributable to off-fault deformation is obtained by subtracting $H_3^{\text{min}}$ from $H_3^{\text{total}}$. On average, off-fault deformation produced $\sim 2$ m of displacement that is not visible in the trench excavation (Figure 6.5b).

### 6.5 Age constraints

Although we cannot date the alluvium directly, its age is bracketed by those of stratigraphically younger and older deposits. Within the trench, the alluvium units are immediately overlain by the Te Rere tephra, which has a calibrated radiocarbon age of $24.4 \pm 0.4$ cal. kyr BP [Alloway et al., 2007]. Although not observed in the trench, the alluvium deposits are mapped elsewhere within the Maleme fault zone to overlay a sequence of older lacustrine sediments [Villamor and Berryman, 2001]. At a location a few kilometers to the west of our survey site, Villamor and Berryman [2001] determined a radiocarbon age of $25.1 \pm 0.4$ cal. kyr BP for the lacustrine sediments. Based on the two bounding ages, we assign a tentative mean age of $24.8 \pm 0.4$ cal. kyr BP to the initial displacement of the upper surface of the alluvium.

Even though our two displacement estimates ($H_3^{\text{min}}$ and $H_3^{\text{total}}$) are measured from the same GPR reflection horizon, they may record two different stages of fault displacement. Our estimates of total displacement ($H_3^{\text{total}}$) are the result of initial deformation of the alluvium surface after the cessation of river erosion and sediment deposition. However, the interface forming reflection horizon H3 was not created until after the Te Rere tephra was deposited. Since the fault scarp may have been somewhat eroded by weathering between the initial period of faulting and tephra deposition, the tephra may have mantled a partially eroded fault scarp. A second period of faulting may then have further offset H3. Consequently, the vertical offsets of reflection horizon H3 on our GPR cross-sections may not have
been preserved until after Te Rere tephra deposition and the hypothetical second period of faulting. Based on this information, we assign a minimum age of $24.4 \pm 0.4$ cal. kyr BP to $H_3^{\text{min}}$.

### 6.6 Influence on extension rate estimates

Our two estimates for the initiation of faulting are very similar, yet the amount of displacement recorded by $H_3^{\text{total}}$ is up to twice that for $H_3^{\text{min}}$. As a consequence, the displacements measured using the two techniques yield two very different vertical slip rates (VSR). By taking the average values of displacement along the $H_3^{\text{min}}$ and $H_3^{\text{total}}$ profiles (Figure 6.5a) and the initial displacement ages outlined above, we derive average vertical slip rates of $VSR_{\text{min}} = 0.10 \pm 0.01$ and $VSR_{\text{total}} = 0.18 \pm 0.02$ mm/yr, respectively. Using the average 72° fault dip observed in the migrated GPR data along fault strand s1, $VSR_{\text{min}}$ and $VSR_{\text{total}}$ yield minimum and total extension rates of $0.03 \pm 0.01$ and $0.06 \pm 0.01$ mm/yr, respectively. Thus, off-fault deformation in the form of drag folding and block rotation accommodates $50\%$ of total extension across the fault.

We suggest that such a large component of off-fault deformation is a consequence of the low strength of the near-surface geology and the immaturity of the Maleme fault zone, which forms the focus of recent extension within the central Taupo Rift [Villamor and Berryman, 2001]. The near surface geology of weakly consolidated alluvium, tephra and paleosol (Figure 6.3a) is more likely to accommodate ductile strains (e.g., drag folding) than more consolidated material. In addition, most of the Maleme fault zone strands are separated by just a few tens of meters (Figures 6.1b and 6.2) and exhibit listric geometries on the GPR and trench cross-sections (e.g., Figures 6.3 and 6.4). As a consequence, the blocks between the faults have probably experienced higher rates of rotation than those between more widely distributed faults further from the central rift axis.

### 6.7 Relevance to regional scale extension across the Taupo Rift

Although the relatively high proportion of off-fault deformation represented in our GPR data may only be significant for recent ($\leq 25$ ka) deformation within the Maleme fault zone, it has important implications for our understanding of present-day rates of extension across the entire width of the Taupo Rift. By measuring vertical displacements of dated geomorphic surfaces and stratigraphic horizons logged in trenches within the central Taupo Rift, Villamor and Berryman [2001] determine an extension rate at seismogenic depth of $6.4 \pm 3.8$ mm/yr for the past 50 kyr. Based partly on the work of Villamor and Berryman [2001], Nicol and Wallace [2007] report geological extension rates of $8 \pm 5$ and $4 \pm 3$ mm/yr for the
6.7. Relevance to regional scale extension across the Taupo Rift

central and southern parts of the Taupo Rift for a much longer period of 1-2 Myr. In contrast, geodetic estimates of strain rate across the central and southern Taupo Rift over the past 20 years are 13.4 ± 1.6 and 8.9 ± 1.3 mm/yr, respectively [Nicol and Wallace, 2007]. Consequently, the geodetic rates are 1.7-2.2 times the geological estimates. Nicol and Wallace [2007] provide the following explanations for the relatively lower rates of geological extension across the Taupo Rift:

- overestimates for the commencement age of tectonic extension, thus yielding lower geological rates,
- deformation has been concealed by volcanism or geothermal activity,
- alternative mechanisms of extensional deformation like volcanic processes (e.g., dike intrusion), and
- higher rates of extension at present than in the past.

Based on the results of our study, we suggest that unrecognized off-fault deformation may be responsible for a large portion of the geological extension-rate deficit. Near-surface extensional deformation in the form of drag folding and block rotation is likely to be important in the youngest relatively immature parts of the central and southern Taupo Rift, where there is a dense concentration of relatively short overlapping fault strands that displace relatively weak stratigraphic layers of tephra and alluvium. In contrast, the older, longer, and more isolated faults from parts of the Taupo Rift that are further from active volcanic centers (and consequently displace older, more compacted, and stronger stratigraphic layers) will tend to exhibit deformation that is concentrated closer to the fault.

Small geological exposures provided by trenches may not be large enough to observe off-fault deformation that extends over several meters. In addition, the cumulative effects of long-term (<25 ka) block rotations that are evident in our GPR data, are difficult to observe at the surface, because they have been masked by subsequent tephra deposition. By using GPR to extrapolate the geometries of fault-displaced strata away from the walls of a trench excavation, we demonstrate that it is possible to measure subsurface fault displacement along and across the fault zone. Although we use a dense 3-D GPR survey to track changes in displacement along a single fault, a series of appropriately located 2-D profiles can be used to extend our observations across the entire width of the Maleme fault zone. In addition, similar GPR measurements have the potential to supplement geological observations from the multitude of trenches excavated at other locations within the Taupo Rift (e.g., Nicol et al. [2006]).
6.8 Conclusions

Off-fault deformation accommodates a major component of displacement within the central Taupo Rift. Prominent GPR reflections imaged within a large 3-D data volume correlate with well-dated stratigraphic horizons observed in a paleoseismic trench. Independent cumulative displacement estimates based on offsets of stratigraphic horizons within the trench and vertical offsets of GPR reflection horizons close to the main fault strand are very similar. By measuring vertical offset from the undeformed parts of the reflection horizons several meters away from the main fault strand, we show that drag folding and horizontal-axis block rotations contribute an extra ~50% to the displacement measurements. This deformation, which is difficult to recognize using standard paleoseismic techniques, may explain why geologically determined extension rates are anomalously low relative to geodetic extension rates within the central and southern Taupo Rift.

6.9 Acknowledgements

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Chapter 7

Conclusions and Outlook

The complex shallow subsurface structures of active fault zones are generally difficult to characterize using traditional paleoseismic techniques like surface mapping and trenching. High-resolution 3-D GPR surveying can be used to characterize shallow fault zone structure where interpretations from surface observations may be ambiguous and provides a low-cost non-invasive addition to trenching. Using examples from active faults in New Zealand, I have demonstrated the potential for GPR surveying in paleoseismic investigations. In the following, I summarize the key advances resulting from my work, including the development and application of attribute techniques, how interpretations from the GPR data can complement paleoseismic observations, and new methods for measuring fault displacements and slip rates. I also discuss some limitations of GPR in active fault investigations and make suggestions for further research.

7.1 Visualization of 3-D GPR data using geometric attributes

Although correctly processed and migrated GPR volumes can be rich in structural detail, interpretations from these data are often limited by our inability to delineate and describe in a quantitative fashion the typically diverse reflection patterns. To minimize the subjectivity of my interpretations and to extract maximum structural information, I took advantage of coherence- and texture-based attribute volumes computed from the processed GPR volumes. In Chapters 2 and 4, I demonstrated how these techniques improve interpretations of 3-D GPR data collected across four different active fault zones: the transpressive Alpine fault zone, the strike-slip Wellington fault zone, normal faults of the Maleme fault zone, and reverse faults of the Ostler fault zone.

Many of the structures observed in our attribute volumes were not obvious in the original GPR volumes. The coherence-based attributes provided useful quanti-
ative information on the coherency, azimuth, and dip of reflections. They proved to be an excellent tool for delineating different types of stratigraphic reflection juxtaposed by faulting (e.g., the dipping gravel layers in the Alpine fault zone survey) and for quantifying the geometry of fault-plane reflections (e.g., in the Wellington and Ostler fault zone surveys). Normal fault strands within the Maleme fault zone data set were well resolved on the basis of their low coherency. The texture-based attributes provided information overlapping and complementary to that supplied by the coherence-based attributes and were most useful for delineating distinctive reflection patterns associated with the various geological units.

### 7.2 Using GPR to complement paleoseismic investigations

The results of this thesis have demonstrated the utility of GPR surveying in diverse paleoseismic investigations. Although the locations of the different active fault zones were identified on the basis of their prominent topographic expressions, little was previously known of their subsurface structure. For example, the Alpine fault zone GPR data revealed three steeply dipping overlapping fault strands, whereas previous geomorphic mapping had indicated only a single strand. Similarly, GPR data from the Ostler fault zone imaged three merging and splaying fault strands that were not evident at the surface. Where subsurface geological information on fault structure was available at the Maleme fault zone (trench) and Wellington fault zone (riverbank exposure) sites, the GPR data allowed these observations to be extrapolated for many tens of meters along and across fault strike.

### 7.3 Limitations of GPR in paleoseismic investigations

The case studies presented in this thesis were undertaken at sites where the local geology was favorable for the acquisition of high-quality GPR data. For example, strong electrical contrasts within sequences of alluvium and tephra at the Alpine and Maleme fault zone sites enabled strong GPR reflections to be recorded, which could be correlated to stratigraphic and tectonic structures. In contrast, locations where the shallow subsurface is distinguished by relatively small changes in electrical properties may yield only weak reflections. Moreover, in areas characterized by highly conductive surface and subsurface materials such as clays, electromagnetic waves may be strongly attenuated.

The Alpine, Wellington, Maleme, and Ostler active fault zones were characterized by prominent near-surface deformation structures. As a consequence, GPR data have provided invaluable information on late Quaternary fault activity. In regions where near-surface fault deformation is not so obvious, alternative methods
would need to be employed to identify and characterize active faults. For example, active faulting in the Swiss Alps has been poorly understood, largely because strain rates across this region are low and glacial deposits and erosional processes have removed the surface expression of past fault activity. To overcome these limitations, Ustaszewski et al. [2007] used a combination of different paleoseismic techniques to investigate the long-term (post-Miocene) history of an active fault in the western Swiss Alps. As part of this study, I acquired a 3-D GPR data set across the fault to look for evidence of recent (<10 ka) activity. The results of their study, including my contribution are presented in Appendix A.

7.4 Future uses of GPR in fault zones and other environments

7.4.1 Attribute methods

The geometric attribute techniques shown in Chapters 2 and 4 are applicable to other types of GPR investigation. They would be particularly useful for any study that requires discrimination of different GPR reflection facies. For example, coherence- and texture-based attributes could be used to determine depositional sequences within coastal landforms, fan deltas, or glacial environments.

In addition to the coherence- and texture-based algorithms, new attribute methods are being developed for 3-D seismic interpretation that will have potential application to 3-D GPR data. One of the latest advances is the development of curvature attributes [Al-Dossary and Marfurt, 2006]. Curvature attributes are complementary to the coherency attribute, because they can be used to identify subtle faults within 3-D data that are at or below the limits of resolution (e.g., Figure 7.1). Like the coherence- and texture-based attributes, they are computed from sub-volumes of data surrounding a data point. Thus, it would be relatively simple to adapt my computer codes to take advantage of this technique.

7.4.2 Quantifying strike-slip displacements

One of the major advantages of 3-D GPR in paleoseismic investigations is the capability to image both lateral [Gross et al., 2002] and vertical (e.g., Tronicke et al. [2006]) fault displacements. From the Malene and Ostler fault zone data sets, I showed examples of vertical fault displacements of stratigraphic layers. Unfortunately, none of the data sets imaged laterally offset piercing points, from which I could quantify components of strike-slip displacement.

Because they often contain datable material, buried paleochannels are excellent markers for measuring lateral slip rates. But, they are difficult to locate using costly and invasive trenching techniques (see Weldon et al. [1996] for examples). Ac-
Figure 7.1: Examples of coherence and curvature attributes calculated along a 3-D seismic reflection horizon (reproduced from Chopra and Marfurt [2007]). (a) Time surface. (b) Coherency. (c) Most-positive curvature. (d) Most-negative curvature. Curvature attributes indicate a better focusing of the base and edges of the channels and other features as compared with coherency. Yellow arrows indicate a linear feature on the coherency display, but the same feature appears crisper on the most-positive curvature display. Cyan arrow indicates the well-defined lower leg of the channel on the curvature displays, which is not seen clearly on the coherency display. The edges of the meandering channel are well-defined throughout the most-positive curvature display and the base of the channel can be followed clearly on the most-negative display (data courtesy of Arcis Corporation).
7.5 Future research at surveyed locations

Accordingly, 3-D GPR can be used as an alternative method to find buried paleochannels and identify suitable trenching sites. The identification of laterally displaced paleochannels would be of great benefit to ongoing investigations of late Quaternary strike-slip rates along the Alpine and Wellington fault zones (e.g., Norris and Cooper [2001]; Langridge et al. [2005]).

7.5 Future research at surveyed locations

The Alpine fault zone GPR data successfully imaged the shallow (<15 m) structure of steeply dipping (78 ± 8°) fault strands within late Pleistocene-Holocene river gravels. The dip and fault structure at greater depth at this location is uncertain. Members of the ETH Applied and Environmental Geophysics group have acquired high-resolution 3-D seismic reflection data to image fault zone structure beyond the sedimentary cover into basement rock to depths of ~200 m [Kaiser et al., 2007]. As part of this investigation, an attempt will be made to correlate structures interpreted from the GPR data with the seismic data.

The Ostler fault zone GPR data were acquired at a location where folding and a series of small faults have accommodated displacement in a transfer zone between two thrust segments. A series of high-resolution 2-D seismic lines were acquired near the site of the GPR survey by members of the ETH Applied and Environmental group [Campbell et al., 2007]. The results of this study should provide some insight into how displacement is transferred between the two fault segments.

Although I have only sampled a small fraction of extension across the Maleme fault zone, it would be relatively straightforward to extrapolate my results across the entire width of the fault zone using a series of 2-D GPR profiles. In addition, my method for measuring displacements using a combination of 3-D GPR and trench data could be applied to other trenched faults within the Taupo Rift. Although, the 3-D GPR data from the Maleme fault zone have greatly improved our understanding of how extensional deformation is accommodated in the shallow subsurface, further research should include an investigation into how this deformation is accommodated at greater depths. The normal faults imaged by our GPR data are separated by a few tens of meters and they exhibit listric geometries. By investigating how and at what depth these faults merge, we may gain insight into how fault displacements produced by earthquakes at seismogenic depths are transferred to the near-surface strands. In 2003, members of the ETH Applied and Environmental Geophysics group acquired a high-resolution seismic reflection profile across the Maleme fault zone in an attempt to image the fault structures to depths >100 m. Processing of this data set is ongoing.
7.6 Measuring fault displacements and slip-rates using 3-D GPR data

Where they can be correlated to trench observations, 3-D GPR data can be used to estimate slip rates on active faults. Previous studies have shown that although fault slip rates can exhibit spatial and temporal variability over short time scales, after a characteristic time period these faults begin to exhibit uniform slip-rate patterns. Using a combination of GPR surveying and trenching, I was able to determine slip-accumulation patterns and rates along and across several normal fault strands within the Maleme fault zone for the past 24.4 kyr. By calculating average cumulative displacements with time for five practically complete fault strands, I was able to obtain robust slip-rate estimates. The slip rates were variable for time intervals \( \leq 12.5 \) kyr long, which suggested that at least four earthquakes are required for these faults to exhibit uniform slip rates characteristic of their long-term behavior.

From the Maleme fault zone GPR data set I also imaged drag folding and horizontal-axis rotations within the hanging-wall and footwall of a fault strand. By measuring the geometries of a GPR reflection horizon several meters from the fault, I showed that off-fault deformation had accommodated around 50\% of recent (\( \leq 24.4 \) kyr) extension across the fault. This type of deformation was difficult to recognize using standard paleoseismic techniques, which may explain why geologically determined extension rates were anomalously low relative to geodetic extension rates within the central and southern Taupo Rift.

The techniques used to estimate displacements and slip rates from the Maleme fault zone GPR data could be applied to other active fault zones. Such a study requires: 1) the offset, and subsequent preservation of, datable stratigraphic horizons deposited synchronously with faulting and 2) that these stratigraphic horizons can be imaged using GPR.
Appendix A

Unravelling the evolution of an Alpine to post-glacially active fault in the Swiss Alps

Michaela Ustaszewski, Marco Herwegh, Alastair F. McClymont, O. Adrian Pfiffner, Robyn Pickering, Frank Preusser

A.1 Abstract

The Gemmi fault is a prominent NW-SE striking lineament that crosses the Gemmi Pass in the central Swiss Alps. A multi-disciplinary investigation of this structure that included geological mapping, joint profiling, cathodoluminescence and scanning electron microscopy, stable isotope measurements, luminescence- and U-Th-dating, 3-D ground penetrating radar (GPR) surveying and trenching reveals a history of fault movements from the late Pliocene to the Holocene. The main fault zone comprises a 0.5-3 m thick calcite cataclasite formed during several cycles of veining and brittle deformation since <2.5 Ma. Displaced Cretaceous rock layers show an apparent dextral slip of ~10 meters along the fault. A detailed study of a small sediment-filled depression that crosses the fault provides evidence for a post-glacial reactivation of the fault. A trench excavated across the fault exposed a Late-Glacial-age loess layer and Late Holocene colluvial-like slope-wash deposits that showed evidence for fault displacement of a few centimetres, indicating a recent strike-slip reactivation of the fault. Focal mechanisms of recent instrumentally recorded earthquakes are consistent with our findings that show that the fault at the Gemmi Pass, together with other parallel faults in this area, may be reactivated in todays stress field. Taking together all the observations of its ancient and recent
activity, the Gemmi fault can be viewed as a window through geological space and time.

A.2 Introduction

In the central Alps, the collision of the Adriatic and the European continental margins initiated crustal thickening during the Cenozoic [Schmid et al., 1997; Escher et al., 1997; Pfiffner et al., 2002]. GPS measurements show that the convergence of the two plates is ongoing at a rate of <2 mm/year [Calais et al., 2002], which is likely to have caused recent rock uplift and seismicity [Sue and Tricart, 2003; Persaud and Pfiffner, 2004; Ustaszewski and Pfiffner, in press].

The Alps have been studied for over 150 years and their evolution has been recorded in great detail (e.g. Trümpy [1998, 2003]; Pfiffner et al. [1997]). In contrast, little is known about their recent history. This is surprising, because active and potentially seismogenic faults represent a significant hazard to life and constructions. Knowledge of their neotectonic history (<15 ka) is therefore a key issue and a pre-requisite for any assessment of seismic hazard.

Few previous researchers have touched on the topic of recent tectonic activity in the Alps. High-precision levelling measurements have been conducted since 1917 [Kahle et al., 1997; Funk and Gubler, 1980; Gubler et al., 1981; Schlatter and Marti, 2002], and the Swiss Seismologic Survey publishes an annual report of the seismic activity in Switzerland (e.g. Baer et al. [2005]; Deichmann et al. [2007]). Todays stress regime, inferred from focal mechanisms of instrumentally recorded earthquakes shows a contrast between the Penninic units south of the Rhône river, where N-S extension predominates, and the Helvetic units north of the Rhône river, which are characterised by a dextral strike-slip regime with a NW-SE oriented $\sigma_1$-axis and a NE-SW oriented $\sigma_3$-axis [Maurer and Deichmann, 1995; Maurer et al., 1997; Deichmann et al., 2002; Kastrup et al., 2004].

Sue [1998] identified an active fault in the Western Alps (Briançonnais Alps). Delacou et al. [2005] investigated large earthquakes in the Chablais area (south of Lake Geneva), and found a correlation between their focal mechanisms and known fault systems in the region. Neotectonic activity in eastern Switzerland was studied by Persaud [2002] and Persaud and Pfiffner [2004], and in western and central Swiss Alps by Ustaszewski [2007] and Ustaszewski and Pfiffner [in press]. Several authors [Eckardt et al., 1983; Jäckli, 1965; Renner, 1982] have studied post-glacially active faults in the central Swiss Alps. Persaud and Pfiffner [2004] showed that a gravitational component plays an important role in the reactivation of these faults, whereas Ustaszewski and Pfiffner [in press] and Ustaszewski et al. [in press] attributed a more complex interplay of differential uplift, large-scale gravitational movements and presumably also crustal tectonics in the formation of these faults.

Active faults often exhibit very subtle morphologic expressions of neotectonic deformation, which could be missed by a standard geological mapping survey. In
A.3 Geological Setting

In this study, aerial photographs of the central and western Swiss Alps were searched for linear features as a first step to finding potentially post-glacially active faults. A number of these lineaments were subsequently visited in the field to assess their origin. Three different types of lineaments were distinguished: a) lineaments caused purely by tectonic processes, b) lineaments caused purely by gravitational processes, and c) composite lineaments caused by gravitational and tectonic processes [Ustaszewski et al., in press]. To determine the age of past activity of the purely tectonic lineaments, indicative surface expression, morphology and especially deformation of the surrounding Quaternary sediments were studied. Few purely tectonic faults indicated post-glacial movements [Ustaszewski and Pfiffner, in press]. In the course of these investigations, one fault showing unequivocal evidence for post-glacial activity was detected in the Gemmi Pass area (Figure A.1).

The aim of this study was to document and understand the evolution of the fault from Alpine time (Tertiary) to the present. In this sense, the geometry of the fault and its fabric from the km down to the µm scale, and the timing of different deformation episodes were investigated in detail with a multidisciplinary approach that encompassed 3D ground penetrating radar (GPR), trenching, qualitative fabric analysis using cathodoluminescence microscopy and scanning electron microscopy, stable isotope measurements, and age determination (U-Th-dating and optically stimulated luminescence dating). This approach allowed us to unravel the complete structural record starting from (i) the first initiation of the fault, (ii) later repetitive overprints by brittle deformation and fluid flow, to (iii) recent activity.

A.3 Geological Setting

The lineament studied is a prominent NW-SE striking vertical fault located at the Gemmi Pass near Leukerbad in the Canton Valais (Figures A.1 and A.2). The fault is oriented 225/90 and is perpendicular to the regional NE-SW striking fold axes. It can be followed over a horizontal distance of 2.6 km (probably 3.6 km as parts might be covered by slope talus) and an altitude range of 1800 to 2700 m. The fault cuts through the Helvetic nappe stack, which in this area consists of the Doldenhorn, the Gellihorn and the Wildhorn nappe. The affected rock types are Jurassic to Eocene carbonates, sandstones and shales [Pfiffner, 1993] that were thrusted and folded during Oligocene-Miocene times [Pfiffner et al., 2002]. The fault investigated, hereafter referred to as "Gemmi fault", belongs to an array of parallel faults, showing a spacing of several tens to a few hundred meters. The location of the Gemmi fault at the bottom of a large high-elevation valley suggests that the fault is unlikely to have been gravitationally reactivated. The fault is located close to the village of Leukerbad, where thermal springs show temperatures of up to 51°C [Mu\-ralt and Vuataz, 1993]. The absence of tritium in the pure thermal water component...
Figure A.1: (a) Overview map of the study area, geological units after Furrer et al. [1956]. Location of joint profiles PI and PIII of Figure A.3 and GPR and trenching location of Figure A.7 are indicated. Coordinate numbers correspond to the Swiss national km grid. (b) Non-rectified aerial photograph of approximately the same area as in Figure A.1a. Arrows show the distinctive fault trace of the Gemmi fault.
indicates that the water must have a long underground residence time of at least 50 years [Muralt and Vuataz, 1993]. A Tertiary hydrothermal-telemagmatic dyke is reported at the Trubelnstock, 3 km south of the Gemmi fault [Furrer and Hügi, 1952], suggesting that elevated geothermal gradients have been present since that time.

A.4 Fault expression in bedrock

A.4.1 Characterization of fault zone

The fault exhibits a pronounced geomorphologic expression marked by a broad and, in places, deep (~3 m) incision in the landscape (Figures A.2a and A.2b). It cuts through predominantly Jurassic to Eocene limestone and, in a few places, sandstone beds. The fault zone has a ~20 m wide damage zone and the core zone itself varies in width between 0.5 and 3 meters (Figure A.3a). By following a weakly preserved lineation on the fault plane (fault plane: 250/55, striation: 335/05), an apparent dextral displacement of ~10 meters can be calculated from the offset host rock layers (bedding: ~305/12). This is a minimum estimate of the slip along the fault since its initiation because multiple reactivations of the fault with changing sense of slip cannot be ruled out. The offset along the fault is too small to be observed in the aerial photograph of Figure A.1b.

The host rocks are incorporated into the edges of the damage zone, where the limestones exhibit a cataclastic texture with white calcitic veins. The latter become more abundant towards the centre of the fault zone (Figures A.3b and A.3e). The core zone is dominated by white cataclasite consisting of broken calcite joint filling. In those parts of the damage zone that have not undergone extensive brittle deformation, large calcite rhombohedra, with side lengths of up to 10 cm, are observed together with idiomorphic calcite crystals in open fissures (Figures A.3b, A.3c and A.3d).

Joints are pervasive through out the entire bedrock, forming a dense, parallel and predominantly NW-SE oriented pattern (see rose diagrams in Figure A.3a). Joint density measurements revealed an increasing density of fault parallel joints toward the main deformation zone, indicating a concentration of deformation towards the centre of the fault zone (Figures A.3a and A.3b). Closed, open and reactivated joints were recorded in two ~40 m long profiles across the fault zone. Open and reactivated joints appear in a broad 15-to-20-m-wide damage zone (Figure A.3a). The core zone itself is characterised by a very low number of visible joints. This anomaly is caused by worsening outcrop conditions and debris cover because of the strong disintegration of the fault rock.
Figure A.2: (a) Overview photograph of the Gemmi fault (indicated by black arrows), showing its pronounced morphologic expression. Location of joint profile PIII of Figure A.3 is indicated. (b) Overview photograph of the Gemmi fault (indicated by black arrows). Locations of joint profile PI of Figure A.3, GPR-survey area (Figure A.7c) and trenching site (Figures A.7a and A.7b) are indicated. (c) Subtle micromorphological observations along the Gemmi fault. Here, the fault trace crosses a slope and is discernible within displaced moraine scree.
A.4. Fault expression in bedrock

Figure A.3: (a) Joint distribution along profile PI and PIII, profile locations are shown in Figures A.1 and A.2. Rose diagrams show strike and dip angle of joints. (b) Profile across the core zone of PIII showing zones of different degrees of brittle overprint. (c) Calcite rhombohedrons with length of up to 10 cm indicate large amounts of fluids have percolated through the fault zone. (d) Idiomorphic calcite crystals grown in open voids. (e) Incorporation of the carbonate host rock (dark grey) into the fault zone at the border of the damage zone indicated by the increasing proportion of white calcite material.
A.4.2 Different generations of veins and cataclasite

Thin sections of the fault rock and the joints were investigated using cathodoluminescence microscopy in order to characterize the different phases of fluid flow and brittle reactivation of the fault zone. Variations in cathodoluminescence colours and intensities represent variations in the amounts of Mn and Fe, provoking and quenching cathodoluminescence, respectively. These different cathodo-microfacies allow us to decipher the geologic evolution of the sample, in particular, the interaction of brittle deformation, fluid flow and calcite precipitation.

Repetitive cycles of fluid flow and brittle reactivation of the fault zone can be discriminated by cross-cutting relationships and variations in cathodo-microfacies, where older generations generally appear in darker colours than the younger generations (Figures A.4a and A.4b). The samples of the fault rock display at least six phases of fluid flow and brittle reactivation, four of them are shown in Figure A.4a.

A representative sample taken from the centre of the fault zone can be subdi-
vided using textural criteria into three zones that parallel the fault plane (zones 1, 2 and 3 in Figure A.5a). Zone 1 is the oldest and comprises a proto-cataclasite with clasts of calcite crystals as large as 6 cm in a fine grained calcitic matrix, forming a component-supported fabric. This zone reveals abundant brown-red, iron-rich solution seams. Zone 2 consists of <1 cm large clasts embedded in a layered gouge-matrix (Figures A.5b and A.5c). These clasts are composed of fine-grained cataclastic material and are rimmed by partly idiomorphic calcite crystals forming circumgranular crust cements and, in places, pendant seams. The matrix consists of fine-grained calcite with a grain size varying from 5 µm to 0.5 mm. Scanning electron microscopy (SEM) measurements of the matrix show a homogenous composition of almost 100% calcite. Trace concentrations of Mg were found in just a few locations. Bands of partly idiomorphic calcite crystals, with growth-directions towards the fault plane, pervade the matrix and have spacings of ~1 mm. Zone 3 is closest to the fault plane and contains a component-supported proto-cataclasite with few voids, where calcite crystals grew into an open fissure. Cataclasite clasts in this zone are as large as 1 cm.

A.4.3 UV-light

Observations of the sample under ultra-violet light revealed a greenish luminescence of zone 2. An investigation for bio-markers uncovered only recent organic material, precluding the action of micro-organisms in the formation of the layering in the fault gouge. Therefore, microbiological activity during the formation of zone 2 is unlikely (pers. comm. V. Thiele and J. Reitner, University of Göttingen, Germany). A possible explanation for the luminescence is that organic material, originating from nearby black shales, was infiltrated into the fault zone by fluids.

A.4.4 Calcite twins

The nature of calcite twins gives information about the temperature conditions during their formation [Burkhard, 1993]. Calcite grains of zone 1 are dominated by curved thick twins and twinned twins of type III, indicating formation temperatures of >200°C and intense deformation. In zone 2, the growth rims of the clasts and the differently sized matrix grains show scarce evidence for twinning (Figure A.6a). Thin and straight type I twins are observed at only a few locations (Figure A.6b). The lack of twinning indicates formation temperatures of <200°C and very little deformation. The grain boundaries of matrix calcite grains are lobate (Figure A.6b), indicating grain boundary migration under the presence of fluids. However, cataclasite clasts in zone 2 are densely twinned and show thick, patchy twins with sutured boundaries, curved thick twins and twinned twins (Figures A.6a, A.6c and A.6d). These type IV and type III twins indicate intense deformation, dynamic recrystallisation and intracrystalline deformation mechanisms at temperatures above 250°C.
Figure A.5: Different observations made from one sample of Gemmifault rock. (a) Zone 1: medium-grained fine-grained calcite matrix with thin comminuted calcite rims. (b) Magnification of the area highlighted in (a). (c) CL-image of zone 2 illustrating different phases of fluid flow and brittle reactivation. Mitre of calcite matrix with newly grown crystals. (d) δ¹⁸O isotope profile across the three zones (locations of measurements indicated in (a)).
A.4. Fault expression in bedrock

Figure A.6: Ultra thin sections of zone 2 cataclasite. (a) Example of a large clast (on the right) with a calcite rim in calcite matrix with different grain size. (b) Example of matrix grains showing lobate grain boundaries but no twinning. (c) Example clast shows bent twins (twin Type III) and twin boundary migration (twin Type IV). (d) Example clast shows twinned twins (twin Type III) and sutured twin boundaries (twin Type IV).

and 200°C, respectively. Twin boundary migration post-dating the twinning is observed in Figures A.6c and A.6d. Calcite grains of zone 3 show only a few type I twins.

A.4.5 Isotope profile

Oxygen ($\delta^{18}$O) and carbon ($\delta^{13}$C) isotopic ratios were measured from a fault rock sample displaying macroscopic layering in the cataclasite (Figure A.5d). All calcite values are reported relative to V-PDB (Vienna Pee Dee Belemnite Standard), the reproducibility of standard materials is ± 0.08 (V-PDB). A short description of the methodological aspects of the isotope measurements is given in Appendix 1 (section A.9). The carbon isotope ratios range from 0.62 to 0.96‰ V-PDB in zone 1, and from -0.07 to 0.16‰ V-PDB in zones 2 and 3. The oxygen isotope ratios decrease linearly with increasing distance from the fault plane from -20.08 to -26.11‰ V-PDB. A reference sample of the host rock limestone showed $\delta^{13}$C values of 1.47 to 1.72‰ V-PDB and $\delta^{18}$O values of 9.86 to 6.47‰ V-PDB.
A.5 Fault expression in Quaternary deposits

Quaternary sediments cover the fault at several locations. Rupture of the fault in post-glacial times would have disrupted the cover sediments providing evidence for recent fault activity. In the western part of the Gemmi fault, a subtle indication for recent reactivation was discernible at the ground surface, where the fault trace crosses a slope covered by moraine scree (Figure A.2c) and displaces these young sediments. At another location along the fault a small ~60 x 30 m depression in the limestone has been filled in with glacial and post-glacial sediments. A 3-D ground penetrating radar (GPR) survey was undertaken over this small basin to investigate potential disruption of the sediments (Figure A.2b). Results from the survey were used to identify an appropriate trench location, where the stratigraphic sequence could be investigated in more detail.

A.5.1 GPR data

A detailed description of the methodological aspects of the GPR-survey is given in Appendix 1 (section A.9). The processed data show a characteristically chaotic reflection pattern from the karstic limestone and more continuous, horizontal reflections from within the layered sediments (Figure A.7a). The chaotic pattern of reflections from the limestone precluded identification of the fault zone within the basement. A comprehensive analysis of trace attributes, including reflection strength and instantaneous phase, identified subtle truncations of some of the horizontal reflections within the sediments that could be attributed to strands of the fault. After carefully interpreting cross-sections of the data volume we could not positively identify fault displacement but could not discount past vertical movements on the order of <20 cm, the limit of the resolution of the data. The sharp contrast in reflection patterns between the bedrock and the young sediments allowed us to accurately determine the 3-D geometry of the basin revealing a deepening of the basin bottom to more than 4 meters towards the north.

A.5.2 Trench data

The Quaternary sediments were investigated in a large trench (15.4 m long and 2 m wide) that was excavated across the suspected fault (Figures A.1 and A.2b). The trench site was selected in the narrowest part of the basin, where GPR data had revealed the limestone basement at a maximum depth of ~2 m, allowing us to reach the cataclasite beneath the sediments. The uncovered bedrock formed an asymmetric basin with a maximum depth of >2.2 m at the northeastern end of the trench (x in Figure A.7b). The near vertical bedrock wall on the northeastern side of the trench exhibits a steep-dipping to vertical polished fault plane.

The trench log in Figure A.7b and A.7c illustrates the composition of the Quaternary sediments filling the depression. The base of the fill consists of a dark
A.5. Fault expression in Quaternary deposits

Figure A.7: (a) GPR cross section coincident with trench wall. (b) Trench log. The disrupted zone within the loess layer is shown in Figure A.8. (c) Photograph of trench wall.
brown diamictic layer of up to 1.5 m in thickness. At the bottom, well consolidated diamictic material (g in Figure A.7b) is overlain by weakly consolidated diamictic material (e and f in Figure A.7b). They both bear clasts with a variety of lithologies that are exotic to this part of the Alps. We interpret these deposits as glacial sediments, i.e. basal till and reworked till, respectively. Large boulders of up to 1 m in diameter were found in the till material. These glacially transported blocks consist of limestone, sandstone and black shales. The larger blocks are imbricated and dip towards the southwest, indicating a northeast paleo-iceflow direction that originated from the nearby Wildstrubel glacier towards the northeast.

We interpreted a relatively constant 20 to 30 cm thick, fine-grained (silt to fine sand fraction), yellow layer deposited on top of the moraine (d in Figure A.7b) as a loess layer. It has a sharp upper contact, whereas the basal contact to the diamictic material is irregular. The loess layer mantles the underlying till unit.

An up to 1.5 m thick layer of grey-brown, colluvial-like slope-wash deposits overlays the yellow loess layer (b in Figure A.7b). It consists of brown fine-grained silt material intercalated with numerous sand and gravel lenses. We also observed thin (~5 cm), horizontal clay bands that extend for up to 7 m within this unit (c in Figure A.7b). These bands onlap the loess at both sides of the basin.

The uppermost 5 to 15 cm of the trench wall contains the top soil layer (a in Figure A.7b). A cataclastic fault zone disrupts the partly karstified limestone bedrock from meter 6.4 to meter 6.8. This 40 cm wide zone is split in the middle by an open joint or fault plane with a possible fault gouge filling. Neither striations on the fault planes nor vertical displacement of the bedrock surface were observed at this location.

In order to check for potential post-glacial activity at the fault, the unconsolidated sediments directly above the cataclastic bedrock were studied in detail. The ~50 cm thick moraine layer showed no evidence of deformation but small-scale disturbance within this very heterogeneous moraine material is difficult to detect. However, the otherwise continuous yellow loess layer is heavily disrupted at this location, incorporating moraine material from below. It displays flame-like structures and the upper surface is vertically displaced by at least 5 cm across the disrupted zone (Figure A.8). We attribute most of these structures to fault movement because they occur over a short interval immediately above the cataclasite zone in the bedrock. Only the lower part of the overlaying slope wash deposits are affected by the deformation. A continuous clay band approximately 30 cm above the loess sediment is not displaced (Figure A.8). We suggest that this layer was deposited some time after the last fault movement.
A.5. Fault expression in Quaternary deposits

Figure A.8: Line drawing and close-up photograph of disrupted loess layer. Bold arrows show apparent offset.
A.6 Age of faulting

A.6.1 Dating of veining and cataclastic overprint

In order to constrain the timing of the fault activity, we attempted to date the carbonate fault rock by Uranium-Thorium disequilibrium dating. Because of the vast difference in the solubility of U and Th, calcite precipitates from groundwater with negligible amounts of Th. The resulting disequilibrium in the Uranium decay series and subsequent \textit{in situ} decay of $^{234}\text{U}$ to $^{230}\text{Th}$ can be used as a dating tool [Ivanovich \textit{et al.}, 1992]. We measured five calcite samples from the fault rock sample of Figure A.5: two from zones 1 and 2 (samples taken from calcite rims), and one from zone 3. Samples were pre-screened using a Fujifilm BAS-1800 $\beta$-scanner and areas of higher radioactivity were selected and sub-samples mechanically prepared using a hand-held diamond-wheel saw. Analytical procedures for the separation and concentration of U and Th were undertaken as described by Ivanovich and Harmon [1992]. Isotopes of U ($^{248}\text{U}, ^{238}\text{U}$) and Th ($^{230}\text{Th}, ^{232}\text{Th}$) were measured separately on a Nu Instruments MC-ICPMS at the University of Bern, following the protocol described by Fleitmann \textit{et al.} [2007]. Results are summarized in Table A.1.

The $^{230}\text{Th}/^{234}\text{U}$ activity ratios fall between 0.999 and 1.131 (Table A.1) and are all close to equilibrium, i.e. 1. The U-Th dating technique is based on the disequilibrium between $^{230}\text{Th}$ and $^{234}\text{U}$ at the time of formation of the mineral; equilibrium is re-established after some 500 kyr, forming the upper limit of this dating technique. The $^{230}\text{Th}/^{234}\text{U}$ activity ratios reported here indicate that all the calcite material associated with the fault formed over 0.5 Ma ago.

However, some age information can be obtained from the $^{234}\text{U}/^{238}\text{U}$ values. Groundwater is characterised by ($^{234}\text{U}/^{238}\text{U}$) activity ratios greater than 1, most likely due to the preferential release of $^{234}\text{U}$ during weathering as a result of alpha recoil [Gascoyne, 1992]. Secondary calcites, forming from these waters, therefore inherit their ($^{234}\text{U}/^{238}\text{U}$) activity ratio and these decay to unity with time. By $\sim$2.5 Ma secular equilibrium (where the rate of decay of each intermediate product equals the rate of its production) between $^{234}\text{U}$ and $^{238}\text{U}$ is reached, producing activity ratios of 1. The $^{234}\text{U}/^{238}\text{U}$ activity ratios reported here, ranging from 1.0006 to 1.0782 (Table A.1), i.e. above 1, thus broadly indicate an age of less than 2.5 Ma for the formation of the calcites.

If the original activity ratio is known or can be reconstructed, an age can be determined from the remaining excess of $^{234}\text{U}$. Unfortunately, in this case, the initial $^{230}\text{Th}/^{234}\text{U}$ activity ratios could not be measured or estimated, so no direct ages can be calculated. However, by combining information from the $^{230}\text{Th}/^{234}\text{U}$ and $^{234}\text{U}/^{238}\text{U}$ rations, we can limit the time of formation of the cataclasite and the period of fluid flow to between 0.5 and 2.5 Ma.
A.6. Age of faulting

<table>
<thead>
<tr>
<th>Sample (location)</th>
<th>U (ppb) ±</th>
<th>Th (ppb) ±</th>
<th>$^{234}$U/$^{238}$U</th>
<th>$^{230}$Th/$^{232}$Th</th>
<th>$^{238}$Th/$^{234}$U</th>
</tr>
</thead>
<tbody>
<tr>
<td>MU 1 (zone 1)</td>
<td>65.1 ± 0.1</td>
<td>21.84 ± 0.10</td>
<td>1.0194 ± 0.0020</td>
<td>9.72 ± 0.12</td>
<td>1.060 ± 0.012</td>
</tr>
<tr>
<td>MU 2 (zone 1)</td>
<td>65.1 ± 0.1</td>
<td>15.18 ± 0.07</td>
<td>1.0782 ± 0.0014</td>
<td>15.77 ± 0.19</td>
<td>1.131 ± 0.012</td>
</tr>
<tr>
<td>MU 3 (zone 2)</td>
<td>1248.2 ± 1.8</td>
<td>7.40 ± 0.04</td>
<td>1.0006 ± 0.0007</td>
<td>520.29 ± 4.60</td>
<td>1.018 ± 0.008</td>
</tr>
<tr>
<td>MU 4 (zone 2)</td>
<td>678.7 ± 0.9</td>
<td>38.13 ± 0.19</td>
<td>1.0107 ± 0.0006</td>
<td>56.03 ± 0.47</td>
<td>1.032 ± 0.007</td>
</tr>
<tr>
<td>MU 5 (zone 3)</td>
<td>1871.6 ± 2.5</td>
<td>100.71 ± 0.33</td>
<td>1.0074 ± 0.0006</td>
<td>56.46 ± 0.40</td>
<td>0.999 ± 0.006</td>
</tr>
</tbody>
</table>

Table A.1: Uranium-Thorium dating results, including concentrations in parts per billion of U and Th and isotopic ratios used in age determinations, i.e. $^{230}$Th/$^{234}$U and $^{234}$U/$^{238}$U. Activity ratios were calculated using the decay constants described in Cheng et al. [2000]. All three zones of the cataclasite were sampled; samples are numbered MU1 (Zone I, oldest) to MU5 (Zone III, youngest). The $^{230}$Th/$^{234}$U ratios are close to equilibrium indicating the calcites formed over 0.5 Ma ago. The $^{234}$U/$^{238}$U ratios, close to but not in equilibrium, indicate that none of the calcites are older than 2.5 Ma, thus constraining the period of their formation to between 0.5 and 2.5 Ma.

- $^a$ Measured activity ratio, $(^{234}$U/$^{238}$U) = $(^{234}$U/$^{238}$U)$_{ab}$/$\lambda$($^{238}$U/$^{234}$U), where $(^{234}$U/$^{238}$U)$_{ab}$ is the abundance ratio of $^{234}$U/$^{238}$U and $\lambda$ are the decay constants for $^{238}$U and $^{234}$U, respectively.

- $^b$ Measured activity ratio, $(^{230}$Th/$^{234}$U) = $(^{230}$Th/$^{234}$U)$_{ab}$/$\lambda$($^{234}$U/$^{230}$Th), where $(^{230}$Th/$^{234}$U)$_{ab}$ is the abundance ratio of $^{230}$Th/$^{234}$U and $\lambda$ are the decay constants for $^{234}$U and $^{230}$Th, respectively.
A.6.2 Dating of most recent fault activity

Pollen analysis

An analysis of pollen in the slope wash material did not yield a result as the pollen content was too small and poorly preserved, prohibiting specific age determination. It is possible that the material was exposed at the surface for a long period, such leaving the pollen grains largely decomposed by microbial activity (pers. comm. R. Drescher-Schneider, Graz, Austria).

Luminescence dating

The last daylight exposure of sediment grains, i.e. sediment deposition ages, can be determined from luminescence methods (cf. Aitken [1998]; Stokes [1999]; Preusser [2004a]; Lian and Roberts [2006]). The method analyses a light sensitive signal in quartz and feldspar grains that is erased during light exposure and rises during burial, when the grains are sealed from daylight. The signal is induced by the interaction of radioactive rays with the crystal lattice. To obtain accurate dates, the samples must be transferred to the laboratory without exposing them to light and all preparation and measurements should be carried out under low-energy red-light illumination. The amount of energy absorbed by the minerals is quantified in the laboratory by comparing the natural luminescence intensity of a sample with its response to known given doses (equivalent dose, De). In addition, it is necessary to measure the radioactivity of the sediment, and the luminescence age is then calculated as the ratio of De and dose rate (D) (the radioactive dose acting on the sample for a certain amount of time). Here, three samples from the Gemmi trench have been dated to further constrain the age of the observed fault deformation. Two samples (Gemmi1 and Gemmi2) were taken from the loess layer and one (Gemmi3) from the slope-wash deposits in the upper part of the sequence. For sample location in the trench see Figure A.7b.

A detailed description of the methodological aspects is given in Appendix 1 (section A.9). A detailed list of data used for calculating luminescence ages is summarised in Table A.2. IRSL ages of 13.5 ± 1.6 ka (Gemmi1) and 14.7 ± 2.1 ka (Gemmi2) were determined for the two samples from the loess sediments. The aeolian nature of the sediment and the good performance of the measurement protocol imply that the two ages should be considered reliable and puts the deposits of the loess sediments into the transition between the Last Glaciation of the Alps and the Holocene, i.e. into the Late Glacial period (cf. Preusser [2004b]).

The age of the samples from the slope-wash deposits (Gemmi3) are difficult to interpret. For the K-rich feldspar fraction, 34 of 36 aliquots gave De values between 0 and 6.5 Gray (Gy) and only two values are above 20 Gy (Figure A.9). We consider these two values to represent outliers resulting from partial bleaching of IRSL prior to deposition and those are not considered in further calculations. The large scatter
<table>
<thead>
<tr>
<th>Sample (material)</th>
<th>Grain size (µm)</th>
<th>n</th>
<th>K (%)</th>
<th>Th (ppm)</th>
<th>U (ppm)</th>
<th>W (%)</th>
<th>Depth (m)</th>
<th>D$_{\cos}$ (mGy ka$^{-1}$)</th>
<th>D (Gy ka$^{-1}$)</th>
<th>Recycling ratio</th>
<th>D$_e$ (Gy)</th>
<th>Age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gemmi1 (loess)</td>
<td>4-11</td>
<td>7</td>
<td>0.72 ± 0.02</td>
<td>5.74 ± 0.39</td>
<td>2.72 ± 0.10</td>
<td>24.7</td>
<td>1.20 ± 0.10</td>
<td>281 ± 0.28</td>
<td>2.36 ± 0.09</td>
<td>0.99 ± 0.09</td>
<td>31.9 ± 1.0</td>
<td>13.5 ± 1.6</td>
</tr>
<tr>
<td>Gemmi2 (loess)</td>
<td>4-11</td>
<td>7</td>
<td>0.67 ± 0.01</td>
<td>5.90 ± 0.34</td>
<td>2.75 ± 0.11</td>
<td>23.2</td>
<td>0.70 ± 0.10</td>
<td>302 ± 0.27</td>
<td>2.40 ± 0.06</td>
<td>0.99 ± 0.06</td>
<td>35.3 ± 3.1</td>
<td>14.7 ± 2.1</td>
</tr>
<tr>
<td>Gemmi3F (colluv.)</td>
<td>100-200</td>
<td>36</td>
<td>0.73 ± 0.02</td>
<td>5.22 ± 0.27</td>
<td>2.20 ± 0.10</td>
<td>27.9</td>
<td>0.90 ± 0.10</td>
<td>240 ± 0.24</td>
<td>2.03 ± 0.08</td>
<td>1.08 ± 0.08</td>
<td>2.03 ± 0.28</td>
<td>1.0 ± 0.2</td>
</tr>
<tr>
<td>Gemmi3Q (colluv.)</td>
<td>100-200</td>
<td>31</td>
<td>0.73 ± 0.02</td>
<td>5.22 ± 0.27</td>
<td>2.20 ± 0.10</td>
<td>27.9</td>
<td>0.90 ± 0.10</td>
<td>240 ± 0.24</td>
<td>1.53 ± 0.09</td>
<td>1.05 ± 0.09</td>
<td>3.67 ± 0.43</td>
<td>2.4 ± 0.5</td>
</tr>
</tbody>
</table>

Table A.2: Summary data of luminescence dating. Given for each sample is the analysed grains size, number of measured aliquots (n), the concentration dose rate relevant elements (K, Th, U) with uncertainty resulting from scatter of values for different nuclides and general reproducibility of measurements as given by Preusser and Kasper [2001], present day moisture content (W, given as percentage of the dry sample), cosmic dose rate with uncertainty (D$_{\cos}$), total dose rate (D), the average recycling ratio, equivalent dose (D$_e$) with standard error (see text for details) and resulting age estimate with standard deviation (1σ). (colluv. colluvial-like slope-wash deposits)

$^a$ Assuming that spread of individual D$_e$ values is due to dose rate inhomogeneity gives a mean D$_e$ of 13.3 ± 1.7 Gy and a resultant OSL age of 8.7 ± 2.0 ka.
of the remaining 34 aliquots for very low $D_e$ values can be explained by the low signal/background ratio of the IRSL signal. Mean $D_e$ calculated from these values is $2.03 \pm 0.28$ Gy, which represents an IRSL age of $1.0 \pm 0.2$ ka.

For the quartz fraction an even larger spread of individual $D_e$ is observed, with values ranging from 2.8 to 41.0 Gy (Figure A.10). The source of this large spread of values is not clear. The observed standard variation of 72 % is much higher than values of about 20 % due to dose rate inhomogeneity reported in the literature (cf. Preusser et al. [2007]).

However, it is possible that the sediment under consideration here is highly inhomogeneous with respect to dose rate (e.g. the carbonate may have some hot spots of radioactivity). Using the mean $D_e$ value of $13.3 \pm 1.7$ Gy of all measured aliquots gives an OSL age of $8.7 \pm 2.0$ ka. However, some influence of partial bleaching on the $D_e$ distribution cannot be ruled out. Following Preusser et al. [2007] a mean $D_e$ was calculated assuming a natural spread of individual $D_e$ 20 % values due to dose rate inhomogeneity. We obtained a mean value of $3.67 \pm 0.43$ Gy and an OSL age of $2.4 \pm 0.5$ ka. The likely age of deposition is therefore between $8.7 \pm 2.0$ ka and $2.4 \pm 0.5$ ka. One problem remaining with the dating of sample Gemmi3 is the difference between feldspar IRSL and quartz OSL ages. The most likely explanation, underestimation of feldspar IRSL due to fading, is not supported by experimental evidence. Conversely, we did not find any indication for thermal transfer of quartz OSL, which would result in overestimated quartz ages. Nevertheless, we can conclude that the slope-wash deposits are of Holocene age and are most likely no older than a few thousand years (Late Holocene).
A.7 Discussion and conclusion

Results from the many disciplines used in this study allow us to synthesize a history of the evolution of the Gemmi fault. The Gemmi fault was initiated as an open a-c-joint oriented perpendicular to the regional NE-SW trending fold axes at a late stage of Alpine nappe emplacement and related deformation. It belongs to a group of veins that are pervasive in the area of the Gemmi Pass, with extension directions parallel to the regional fold axes [Dietrich et al., 1983].

Exhumation rates of 0.83 mm/yr estimated from Figure 10 in Herwegh and Pfiffner [2005] allow us to estimate that the exposed fault rock initially formed at about 8-12 km depth at around 10-15 Ma (Figure A.11). A similar exhumation rate for this area can be calculated using typical recent rock uplift rates of 1.2 to 1.3 mm/yr [Kahle et al., 1997; England and Molnar, 1990] and denudation rates between about 0.27 ± 0.14 and 0.9 ± 0.3 mm/yr [Wittmann et al., 2007; Norton et al., 2008].

The Gemmi fault lies within a 20 m wide dense array of brittle joints oriented parallel to the major fault zone. We assume that with progressive strain, some of the joints interconnected to form a 0.5 to 3 m wide zone, which defines the core zone of the present day fault. During this episode new parallel open joints formed and the contact between older vein filling and host rock partly opened. The bimodal distribution of joints in profile PIII (Figure A.3a) may indicate that a second ~20 m wide zone of weakness SW of the main fault has not evolved into a fault zone. Considerable amounts of fluids percolated through the joints, precipitating idiomorphic calcite crystals and blocky calcite cement. As deformation concentrated in the core
Figure A.11: Recapitulatory time-depth-graph. (Depths were calculated using exhumation rates derived from Herwegh and PfiFFner [2005]). Wi: Wildhorn nappe; P/AA: Penninic and Austroalpine lid. Dark grey arrows point to areas of ongoing cycles of brittle deformation and fluid pulses in depth.
A.7. Discussion and conclusion

zone, strain and strain rates became high enough for the local formation of cataclasites. These fracturing and calcite precipitation phenomena observed in the fault rock indicate dissolution precipitation processes under the presence of fluids and brittle faulting.

The joint fill of the core zone was later overprinted by repeated brittle deformation. Several phases of fluid flow and brittle reactivation are registered in the fault rock, as indicated by different cathodo-microfacies at the micro-scale and by a zonation of the fault rock at the macro-scale. Differences in cathodo-microfacies occur either due to variations in saturation of the fluids (source-controlled) or due to deviations in pressure conditions (deformation-controlled). For example, a decrease in pressure caused by an increase in porosity may have led to precipitation.

The layering in zone 2 of the fault rock sample can be explained by a gravitational settling of fault gouge material in a dilative compartment of the fault zone. Small clasts of a more highly deformed cataclasite falling into this void formed a component-supported framework, in which fluid flow was possible. From these fluids calcite crystals grew idiomorphically and formed pendant calcite-crusts around the clasts. The fine-grained gouge material then settled in laminae, intercalated with thin layers of calcite crystals, which were precipitated during new fluid pulses. As only proto-cataclasites occur in the fault rock, shear strain must have been limited. This is confirmed by the relatively low total displacement of about 10 m of the entire fault.

Temperatures during the early stage of faulting can be estimated from microstructures and twin types in the calcite cataclasite. Herwegh and Pfiffner [2005] report peak metamorphic conditions of 260 to 270°C for a region encompassing the Gemmi fault. This is in close agreement with earlier temperature estimates of about 290°C by Burkhard [1993]. These values represent peak temperatures for the host rock of the Gemmi fault, which is in congruence with mylonitic textures and intense dynamic recrystallisation observed in the host rock limestone. Recent studies on carbonate mylonites of the Helvetic Alps have revealed that temperatures higher than ~220°C are required for dynamic recrystallisation in carbonate rocks (e.g. Herwegh and Pfiffner [2005]; Herwegh et al. [2005]; Ebert [2006]).

Observations of the calcite twins indicate that, for the zoned fault rock, temperatures decreased towards the fault plane (Figure A.5a). Using the geothermometer of Burkhard [1993], which is based on the temperature-dependant occurrence of different types of calcite twins, the type III twins within the calcitic joint infill of zone 1 of the fault rock suggest temperatures of >200°C. However, type IV twins were found in the cataclastic clasts of zone 2, revealing temperatures of >250°C. As no evidence for dynamic recrystallisation was found in the fault rock itself, these clasts must originate from a different, presumably older and more highly deformed cataclasite. The calcite grains of the matrix and the rims in zones 2 and 3 show scarce type I twins, which suggests temperatures <200°C.

Using age constraints from the U-Th measurements of the zoned fault rock (0.5
to 2.5 Ma) and exhumation rates calculated from Herwegh and Pfiffner [2005], we can deduce that the fault rock was ~2 km beneath today's surface during its formation (Figure A.11). At this depth, strain rates and temperatures required for the formation of type III twins cannot be reached under normal conditions. Nevertheless, there are several possible sources of additional heat input. Firstly, episodic earthquake deformation and an additional input of shear heating could explain the formation of these twins. And secondly, hot hydrothermal fluids might have been present as indicated by recent hydrothermal springs in Leukerbad. These fluids may have raised the local geothermal gradient both vertically and laterally within the host rock.

The δ¹⁸O values measured from the zoned fault rock sample show very low values (-26.11 to -20.08‰ V-PDB). In contrast, the bedrock carbonates have typical values for marine limestone of 9.86 to 6.47‰ V-PDB [Hoefs, 1997]. Thus, they can be excluded as the primary source of fluids in the fault rock. More likely is the influence of either meteoric water, which has very negative values between -25 to -69‰ V-PDB (i.e. 5 and -40‰ δ¹⁸O SMOW, Hoefs [1997]) or fluids from granitic rocks, which have typical values between -35 to -16‰ δ¹⁸O V-PDB (i.e. -5 to 14‰ δ¹⁸O SMOW, Hoefs [1997]). Recent weathering has not changed the original isotope trend as indicated by the values of the samples of zone 3. These are still higher than those of zone 1, although zone 3 was located closer to the present day surface and therefore should have more negative values if weathering had played a major role.

A significant trend in the δ¹⁸O signal of the fault rock from lower values in zone 1 to higher values in zone 3 is observed. This trend represents either the weaker influence of fluids with negative δ¹⁸O values towards the fault plane, or a temperature decrease. We prefer the latter interpretation because it is consistent with calcite twin observations (Figures A.5a and A.5d).

Burkhard and Kerrich [1988] and Dietrich et al. [1983] studied stable isotope data of veins and faults of the Helvetic nappes. Most of these show equilibrium values between vein and the host rock. Only two samples - a late tensile vein perpendicular to the fold axes and a large late normal fault - from Burkhard and Kerrich [1988] show negative δ¹⁸O values comparable to those from our Gemmi fault, which would point to an external fluid source.

The Uranium-series isotopic ratios also provide some insight into the source of the fluids from which the calcitic fault rock formed. A distinct difference exists in the U-concentrations between the samples of zone 1, and zones 2 and 3 (Table A.1). Zone 1 has a much lower U-concentration of 65.1 ppb, while the other two zones record substantially higher U-concentrations of 678.7 and 1871.6 ppb respectively. This difference indicates different fluid sources for the successive generations of calcite precipitation.

A similar example of a fault bounded hydrothermally mineralized breccia with a multistage deformation history is reported by Hofmann et al. [2004] from the
Grimsel area, about 60 km to the ENE of the Gemmi Pass. Unlike the Gemmi fault examined in our study, this fault is located in crystalline rocks of the Aar massif. The breccia shows remarkably similar fluid flow indicators, crystal precipitation and brittle reactivation features. $^{39}$Ar/$^{40}$Ar dating provides a middle Pliocene (3.30 ± 0.06 Ma) age of formation of the fault breccia. This age is similar to the 0.5 and 2.5 Ma age range for the mineralization of the Gemmi fault reported here.

In a trench excavated across the Gemmi fault we found deformed post-glacial sediments directly above cataclasite fault rock formed within limestone basement. A purely cryogenetic cause for these deformation structures can be ruled out for the following reasons. First, the vertical displacement of the loess occurs only above the fault zone. Second, the displacement of the loess layer can be followed continuously along the strike of the fault and over a distance of 2.5 m across the trench. And third, small water influxes were observed at several places along the floor of the trench but not from the fault zone cataclasite. Therefore, any seasonal freeze-thaw deformation is more likely to have occurred away from the fault zone.

Instead, we argue that the observations of deformed Holocene sediments in the trench indicate a post-glacial reactivation of the fault. The small vertical displacement and the rather chaotic deformation structures suggest strike-slip movement along the fault, consistent with observations of pre-glacial fault displacement at other locations. The OSL dating of the disrupted loess layer and the partly disturbed slope-wash deposits constrains the age of the last fault rupture. Latest activity on the Gemmi fault occurred after the deposition of the loess layer (i.e. after the Late Glacial period). Parts of the colluvial-like slope wash deposits are also deformed, indicating rupture during more recent times. OSL sample Gemmi3 was taken at the same level as the deformed slope wash deposits. Thus its age (between 8.7 ± 2.0 ka and 2.4 ± 0.5 ka) can be taken as the upper limit for the last time of reactivation. We therefore conclude that the last prominent reactivation of the Gemmi fault most likely occurred during the late Holocene. This result should, despite a scarcity of evidence from past studies, encourage further investigation of other faults for evidence of recent activity. We suggest that the present lack of evidence is probably a consequence of several factors: high erosion rates in the Alps, human-induced modification of the landscape, and a paucity of dateable Quaternary deposits, without which the exact timing of fault movements can not be determined.

The strike of the Gemmi fault and the observed displacement along the Gemmi fault correlate with the present-day stress-field, which shows a NW-SE trending compressive $\sigma_1$-axis and a NE-SW extensive $\sigma_3$-axis [Kastrup, 2002; Kastrup et al., 2004]. Furthermore, the Gemmi fault is located at the northeastern end of an elongated earthquake cluster in the Valais north of the Rhône river in the Wildhorn-Rawil-Martigny area, which is directly to the west of this study area [Maurer and Deichmann, 1995; Maurer et al., 1997; Deichmann et al., 2002]. Analyses of focal mechanisms of earthquakes that occurred between 1961 and 1998 [Kastrup et al., 2004] show the predominance of strike-slip earthquakes and reveal two main strike
orientations of possibly seismogenic faults in the area of the Rawil depression: NE-SW and NW-SE. Therefore, NW-SE trending faults like the Gemmi fault are prone to reactivation in today's stress field.

Using correlations between moment magnitude and surface rupture length, between maximum displacement and surface rupture length (after Wells and Coppersmith [1994]) and assuming a rupture over the complete length of the Gemmi fault (2.6-3.6 km), we estimate that the Gemmi fault experienced maximum Magnitudes of 5.5 to 5.8, and displacements of 2 to 5 cm. These values are plausible in the context of the historical earthquake catalogue of Switzerland, where the two largest historically recorded earthquakes in the Valais had moment magnitudes of 6.4 (1524 at Ardon and 1855 at Visp, Earthquake Catalogue of Switzerland).

The Gemmi fault has been active since its formation during Alpine nappe emplacement. Thus, from a single outcrop we can compare active and ancient deformation processes occurring at shallow and intermediate depth, a rare opportunity in the geological record. We suggest that the episodic cycles of brittle deformation and fluid pulses that formed the veins and cataclasites, i.e. the older structures presently exposed at surface, were ongoing at a few kilometres depth during the time of post-glacial fault activity. Given the regional seismicity pattern together with evidence for Holocene and historical activity, we conclude that episodes of veining and cataclasite formation are still occurring at depth within the crystalline basement (Figure A.11). At depth, structures similar to those observed at the Grimsel Pass [Hofmann et al., 2004] are likely to form now. In this way, the Gemmi fault provides a window through geological space and time.

A.8 Acknowledgements

Financial support was provided by the Canton of Bern and the Swiss Geophysical Commission (project Swiss Seismotectonic Atlas). We thank Björn Heincke, Mark Blome, Simone Herzig and Kamil Ustaszewski for their help conducting the GPR-survey. Our special thanks go to Sabine Brodhag, Sybille Fenner, Irène Herwegh, Simon Herzig, Ueli Jörin, Ulrich Linden, and Mario Schneider for their help digging the trench at the Gemmi Pass. We are very grateful to Ruth Drescher-Schneider, Graz, Austria (pollen analysis of the trench samples), Karl Ramseyer, University of Bern, Switzerland (stable isotopes of the fault rock samples), and Joachim Reitner and Volker Thiel, University of Göttingen, Germany (biomarker investigations of the fault rock). RP wishes to thank Ingeburga Hebeisen for her instruction in the preparation of the U-Th samples and Jan Kramers for the U-series discussion. We thank J.P. McCalpin for his challenging review and an anonymous reviewer for the detailed comments, which improved the clarity and quality of the manuscript.
A.9 Appendix 1: Methodological aspects of the different techniques

A.9.1 Isotope measurements

Samples (0.2-0.5 mg) for the stable isotope analysis were drilled with a 0.5 mm diameter conical dentist drill. δ\(^{18}\)O and δ\(^{13}\)C isotopic compositions were measured using a VG Isocarb system attached to a VG Prism II isotope mass spectrometer at the Institute of Geological Sciences of the University of Bern. The phosphoric acid extraction was made at 90°C.

A.9.2 GPR measurements

A Sensors and Software PulsEKKO GPR unit integrated with a self-tracking laser theodolite [Lehmann and Green, 1999] was used to survey the surface area of the basin (Figure A.2b). Stacked radar data were acquired on a dense (18 x 18 cm) grid over the depression using 0.5-m-offset 200 MHz antennas. From additional common-midpoint (CMP) surveys we could determine a representative velocity for the subsurface sediments of 0.063 m/ns. The scaled and bandpass filtered data revealed reflections from both the sediment and the limestone basement to depths of ~5 m. Processing included topographic migration, which was required to collapse diffractions and reposition dipping reflectors while accounting for topographic gradients within the survey area [Lehmann and Green, 2000]. A top mute was used to mask the air- and ground-waves.

A.9.3 Luminescence dating

All samples were extracted by forcing opaque plastic tubes into the sediment, which were removed and sealed in lightproof bags. The outer part of the sediment core was removed in the laboratory to ensure that none of the material could have been exposed to light during sampling. Some portion of the inner part was used for preparation and was chemically pre-treated (20 % HCl, 30 % H\(_2\)O\(_2\)) to remove carbonates and organic matter. For samples Gemmi1 and Gemmi2, the fine grain fraction 4-11 µm of the suspension was enriched by settling using Stokes’ law. A solution of ~2 mg sediment per 1 ml acetone was then pipetted on a stainless steal disc and the sample material settled on the disc after evaporation of the acetone. For sample Gemmi3, which is much coarser, the fraction 100-200 µm was collected by sieving and a quartz and a K-feldspar fraction, respectively, were separated using heavy liquids (sodium polytungstate solutions with densities of 2.70 g cm\(^{-3}\) and 2.58 g cm\(^{-3}\)). The quartz fraction was additionally treated by etching for 40 min in 40 % HF to remove any remaining feldspar contamination and the outer part of the quartz grains, which is affected by alpha irradiation.
All luminescence measurements were carried out on an automated TL/OSL-DA20 reader by Riso National Laboratory at the Institute of Geological Sciences of the University of Bern. Infrared stimulated luminescence (IRSL) of polymineral fine grains and K-feldspar separates were measured at 50°C using a combination of Schott BG39 and L.O.T. Oriel Interference detection filter (410 nm). For the quartz separates, optically stimulated luminescence (OSL) by blue diodes was detected using a Hoya U340 filter. The purity of quartz samples was confirmed by measuring the response of the sample to IR stimulation and all quartz OSL measurements were carried out at 125°C. The luminescence intensity of the samples is generally low (Figure A.12).

Prior to dating a series of experiments was carried out to test the suitability of the measured material and verify the measurement parameters. A common feature observed for quartz that has a relatively young sedimentary history is thermal transfer of the luminescence signal caused by pre-heating the mineral after laboratory irradiation [Rhodes and Bailey, 1997; Rhodes, 2000]. We tested this for the Gemmi3 quartz sample (Gemmi3Q in Table A.2) but found no evidence for thermal transfer, similar to results from Alpine quartz from other areas [Klasen et al., 2006; Preusser et al., 2007]. There is also no evidence for any thermal transfer in the polymineral
fine grain samples. The quartz and feldspar/polymineral fine grains were preheated for 10 s at 230°C and 290°C, respectively. Dose recovery tests revealed that a known laboratory dose was recovered within the precision of the method (dose recovery test are within 10 % of given dose). Storage tests (four weeks at room temperature) were carried out for samples Gemmi1 and Gemmi3F to test if the feldspars are affected by fading, which will result in underestimated ages (e.g. Wallinga et al. [2000]; Auclair et al. [2003]). These tests did not reveal any indication for fading of the investigated feldspars.

The concentration of dose rate relevant elements was determined in the laboratory by high-resolution gamma spectrometry (cf. Preusser and Kasper [2001]). No indication for radioactive disequilibria in the Uranium decay chain was observed (cf. Zander et al. [2007]). In situ sediment moisture was used for dose rate calculation. The geological setting and the relatively young age of the sediment discount any major changes of the hydrological setting during residence time. The contribution of cosmic radiation to the total dose rate is estimated using present day sample depth and considering the high elevation (2370 m) of the sample site (ADELE software).
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