Modelling geodetic observations of postseismic displacement in the central and southern South Island of New Zealand

By

Pegah Faegh-Lashgary

A thesis
submitted to Victoria University of Wellington
in fulfilment of the requirements for the degree of
Doctor of Philosophy in Geophysics.

Victoria University of Wellington
2016

“If I were again beginning my studies, I would follow the advice of Plato and start with mathematics. ”

Galileo Galilei

Abstract

The last seven years have seen southern New Zealand affected by several large and damaging earthquakes: the moment magnitude (M_W) 7.8 Dusky Sound earthquake on 15 July 2009, the M_W 7.1 Darfield (Canterbury) earthquake on 4 September 2010, and most notably the M_W 6.2 Christchurch earthquake on 22 February 2011 and the protracted aftershock sequence. In this thesis, we address the postseismic displacement produced by these earthquakes using methods of satellite-based geodetic measurement, known as Interferometric Synthetic Aperture Radar (InSAR) and Global Positioning System (GPS), and computational modelling.

We observe several ground displacement features in the Canterbury and Fiordland regions during three periods: 1) Following the Dusky Sound earthquake; 2) Following the Darfield earthquake and prior to the Christchurch earthquake; and 3) Following the Christchurch earthquake until February 2015.

The ground displacement associated with postseismic motion following the Dusky Sound earthquake has been measured by continuous and campaign GPS data acquired in August 2009, in conjunction with Differential Interferometric Synthetic Aperture Radar (DInSAR) observations. We use an afterslip model, estimated by temporal inversion of geodetic data, with combined viscoelastic rebound model to account for the observed spatio-temporal patterns of displacement. The two postseismic processes together induce a significant displacement corresponding to principal extensional and contractual strain rates of the order of 10^{-7} and 10^{-8} yr^{-1} respectively, across most of the southern South Island.

We also analyse observed postseismic displacement following the Dusky Sound earthquake using a new inversion approach in order to describe afterslip in an elastic-viscoelastic medium. We develop a mathematical framework, namely the “Iterative Decoupling of Afterslip and Viscoelastic rebound (IDAV)” method, with which to invert temporally dense and spatially sparse geodetic observations. We examine the IDAV method using both numerical and analytical simulations of Green’s functions.

For the post-Darfield time interval, postseismic signals are measured within approximately one month of the mainshock. The dataset used for the post-Darfield displacement spans the region surrounding previously unrecognised faults that ruptured during the mainshock. Poroelastic rebound in a multi-layered half-space and dilatancy recovery at shallow depths provide a satisfactory fit with the observations.

For the post-Christchurch interval, campaign GPS data acquired in February 2012 to February 2015 in four successive epochs and 66 TerraSAR-X (TSX) SAR acquisitions in descending orbits between March 2011 and May 2014 reveal approximately three years of postseismic displacement. We detect movement away from the satellite of ~ 3 mm/yr in Christchurch and a gradient of displacement of ~ 4 mm/yr across a lineament extending from the westernmost end of the Western Christchurch Fault towards the eastern end of the Greendale East Fault. The postseismic signals following the Christchurch earthquake are mainly accounted for by afterslip models on the subsurface lineament and nearby faults.

Acknowledgements

Three years and three offices later I find myself nearing the completion of my thesis. It hasn't been without its ups and downs, speed bumps and roadblocks. My PhD has certainly been an interesting and pivotal point in my life. I am sure IT Support have been waiting for this moment too, what with the hard drive failure, multiple requests for upgrading hardware, oh and crashing the scientific computing cluster (apparently my code was *over 9000*).

When I was not at the VUW campus, I was at GNS processing the finite element codes. I would have to say the most interesting day at GNS was the great flood of 15th May 2015 - where I was not able to get back into Wellington until very late at night (and not by swimming). There was also that one time where I found myself locked *in* the building.

Not all my time was spent in Wellington. I have been fortunate enough to attend three conferences in three different countries: AGU in San Francisco, U.S.A, EGU in Vienna, Austria, and Summer School in Yamanakako, Japan. Each place was a unique experience of its own and I would like to attend more events in future. Overall, my time with Victoria University of Wellington, and New Zealand in general have been wonderful. Without further ado, I would like to acknowledge a number of key people.

First and foremost, I wish to thank my primary supervisor AProf. John Townend of Victoria University of Wellington for his support, guidance, patience and above all, his knowledge of all things Geophysics. John is one of the most intelligent people I have met and has had a great influence on me. I am very grateful to have had him as my supervisor.

Furthermore, I would like to thank Dr. Charles Williams for his guidance, excellence in programming, and debugging skills. Charles has been a great supervisor and has always been there when I have needed assistance using his code libraries.

I thank Dr. Ian Hamling for reviewing my papers in a timely manner. Ian has without a doubt made my papers read well.

I thank Dr. Mahdi Motagh from GFZ, Potsdam for arranging the opportunity for my PhD in New Zealand. Mahdi has always been prompt to reply to my emails (even at 3am - Berlin time). Mahdi's guidance and managerial finesse helped me keep on track.

At this juncture, I would also like to thank the Earth Quake Commission. Without the funding of EQC, this work would not be possible.

Among other people to thank are my proof readers. My flatmate and good friend Jorge who, while failing School-C English, spent countless hours helping me with corrections. My then-office mate Holly who was an absolute machine with grammar. Dominic, the boy from down the hallway who floats around our office, giving us company and cheering me up during difficult times. My office mate Hubert, who was able to decipher John's handwriting - who knew he had such skills in cryptanalysis?

Last but not least I would like to thank my Mum and Dad over in Iran who I miss very much. They have been very supportive and understanding. Also my husband Nick who has been by my side, keeping me fed, and making me laugh.

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Abbreviations

ABIC Akaike's Bayesian Information Criterion	43
ALOS Advanced Land Observation Satellite	4
BCFZ Brittle Creeping Fault Zone	46
BVP Boundary Value Problem	58
cGPS continuous GPS	43
CSK COSMO-SkyMed	162
DEM Digital Elevation Model	15
FEM Finite Element Method	161
FFT Fast Fourier Transform	35
GPS Global Positioning System	13
ICAIM Independent Component Analysis Inversion Method	161
IDAV Iterative Decoupling of Afterslip and Viscoelastic relaxation	161
InSAR Interferometric Synthetic Aperture Radar	13
LiDAR Light Detecting And Ranging	124
LOS Line Of Sight	131
MCMKF Monte Carlo Mixture Kalman Filter	43
MFZ Marlborough Fault Zone	3
NIF Network Inversion Filter	43
PCAIM Principal Component Analysis-based Inversion Method	76
PDEs Partial Differential Equations	58
PS Permanent Scatterer	19
PS-InSAR Permanent Scatterer-InSAR	16
RMSE Root Mean Square Error	140
SBAS Small BAseLine Subset	129

SDFP Slowly Decorrelating Filtered Phase	18
SISZ South Iceland Seismic Zone	53
SLC Single Look Complex	15
SLS Standard Linear Solid	77
SRTM Shuttle Radar Topography Mission	15
StaMPS Stanford Method for Permanent Scatterers	18
SVD Singular Value Decomposition	99
SSEs Slow Slip Events	161
SAR Synthetic Aperture Radar	14
TEC Total Electron Content	23
TSX TerraSAR-X	162

Chapter 1

Introduction

This thesis contributes to the methods of measuring, modelling and interpreting post-seismic ground displacement following the 15 July 2009 Dusky Sound earthquake and the Canterbury earthquake sequence of 2010–2012 in the central and southern South Island of New Zealand. We study a 400-day time interval following the Dusky Sound earthquake and spanning more than three years of the postseismic period after the 22 February 2011 Christchurch earthquake. The interval studied has seen tens of thousands of reported $M_W > 2$ earthquakes¹ with the most significant events being the Dusky Sound earthquake and the Canterbury earthquake sequence of 2010–2012. In this chapter, we describe the tectonic and geological setting and the recent seismicity in the South Island of New Zealand as an introduction to this research. We provide a brief introduction to study areas and the underlying motivations, objectives and outline of this thesis.

1.1 Geological and tectonic setting of the South Island of New Zealand

New Zealand was contiguous with Australia and Antarctica before the Late Cretaceous (Mortimer, 2004). Accretion onto the Gondwana plate margin during the Paleozoic and Mesozoic formed the New Zealand fragment (Mortimer, 2004; Cox and Sutherland, 2007). Extension during the Cretaceous to Paleogene separated the fragment from Australia and Antarctica as a thinned sub-continental crust (Mortimer, 2004; Cox and Sutherland, 2007).

The basement of the South Island of New Zealand is presently divided into Eastern and Western Province Terranes (Bishop, 1985; Bradshaw, 1989). The dominant feature of the Western Province, extending from Fiordland to west Nelson, is the metamor-

¹<http://quakesearch.geonet.org.nz/>

Figure 1.1: a) Basement geology map of New Zealand (Cox and Sutherland, 2007) adapted from (Mortimer, 2004) b) Regional geological map, data source: GNS Science c) Australian-Pacific plate boundary.

phosed Buller Terrane (Nathan, 1976), which amalgamated with Takaka Terrane in the earliest Devonian (Figure 1.1a; Cooper and Tulloch, 1992). The Buller Terrane consists mainly of Late Cambrian to Late Ordovician quartzose sediments (Nathan, 1976). Metamorphosed Paleozoic sedimentary and volcanic rocks constitute the Takaka Terrane (Cooper, 1989). The Eastern Province Terranes, extending from Otago to East Cape, containing Permian to Jurassic sedimentary rocks, are dominated by the Rakaia and Pahau Terranes (Figure 1.1a; Mortimer, 2004). The Rakaia Terrane was over-printed by metamorphism during the Jurassic to Early Cretaceous forming the Haast Schist. The Pahau Terrane contains younger Late Jurassic to Early Cretaceous Torlesse quartzofeldspathic greywacke that underwent later metamorphism (Figure 1.1a and b; MacKinnon, 1983; Wandres et al., 2004, 2005).

The Australian-Pacific plate boundary, with current high rates of motion, was formed during the Palaeogene and Neogene (Cox and Sutherland, 2007). We summarise the main features of the present New Zealand plate tectonic from northernmost to southernmost as follows (Figure 1.1c):

- Subduction of the Pacific plate along the Hikurangi Margin beneath the North Island (Barnes and Lépinay, 1997; Barnes et al., 1998; Reyners, 1998; Eberhart-Phillips and Reyners, 1999; Wallace et al., 2004; Barker et al., 2009);
- Transition to strike-slip along the Alpine Fault (Van Dissen and Yeats, 1991; Holt and Haines, 1995; Beavan et al., 1999; Norris and Cooper, 2001); and
- Eastward subduction of the Australian plate beneath the Pacific plate along the Puysegur Trench (Norris et al., 1990).

The principal fault structure in the South Island is the Alpine Fault, which accommodates $\sim 70\%$ of relative motion between the Australian and Pacific plates (Berryman et al., 1992; Sutherland and Norris, 1995; Beavan et al., 1999) and along with the Marlborough Fault System connects the northern and southern subduction zones (Berryman et al., 1992). Sutherland et al. (2000) proposed that the continental collision zone is a result of strike-slip motion since ~ 25 Ma that has localised shear strain near the Alpine Fault. The link between the northern segment of the Alpine Fault and the Hikurangi Margin is in the Marlborough Fault Zone (MFZ) (Walcott, 1998). The southern Alpine Fault continues offshore and merges with the Puysegur Trench (Figure 1.1c; Delteil et al., 1996; Lebrun et al., 2000.) Lebrun et al. (2000) interpreted the offshore extension of the Alpine Fault from Milford to George Sound in the northern Fiordland Margin from sea floor morphology. The offshore segment of the fault, inferred from bathymetric and seismic reflection, is geometrically more complex than the onshore segment (Barnes et al., 2005; Delteil et al., 1996). In the central Fiordland Margin, the Alpine Fault splits into the West branch linking the northern end of the Puysegur Trench, and the East branch. In the southern Fiordland Margin, the expression of the West and East branches can be seen from the sea floor morphology and seismic reflection data (Lebrun et al., 2000).

1.2 Recent seismicity in the South Island of New Zealand

As a result of ongoing collision between the Pacific and Australian tectonic plates, New Zealand is highly seismically active. This seismic activity has manifested itself as tens of thousands of $M_W > 2$ reported earthquakes over the last five years. We briefly summarise the main characteristics of the most recent significant earthquakes in the South Island, namely the 15 July 2009 Dusky Sound, 4 September 2010 Darfield and 22 February 2011 Christchurch earthquakes and the Cook Strait earthquake sequence of July-August 2013.

The largest earthquake to strike southern New Zealand since the M_W 7.8 Murchison earthquake of 1929 is the M_W 7.8 Dusky Sound earthquake (focal mechanism shown in Figure 1.4). Based on analysis of strong-motion records, Fry et al. (2010) inferred that rupture initiated at a depth of 38 km and proceeded upwards and southwestwards at a velocity of ~ 2.6 km/s. Beavan et al. (2010a) modelled the fault geometry and slip distribution of the Dusky Sound earthquake using campaign and continuous Global Positioning System (GPS) data as well as ascending and descending tracks of the Advanced Land Observation Satellite (ALOS) (Figure 1.2a is a reproduction of Figure 8 from Beavan et al. 2010a). Their results suggest a maximum slip of ~ 6 m occurring on a curve-shaped patch of approximately 4000 km², equivalent to a scalar moment of $\sim 4.0 \times 10^{20}$ N·m.

Both the M_W 7.1 Darfield (Canterbury) earthquake of 2010 and the ensuing M_W 6.2 Christchurch earthquake of 2011, which caused 185 fatalities, occurred on previously unrecognised fault planes (focal mechanisms shown in Figure 1.4; Gledhill et al., 2011; Kaiser et al., 2012; Quigley et al., 2012). Gledhill et al. (2011) documented a complex source structure for the Darfield earthquake which seismic and geodetic observations suggest it initiated on a reverse fault and spread onto nearby strike-slip faults. The ~ 30 km long surface rupture and the static stress drop of ~ 13 MPa of the Darfield earthquake were measured using the real-time kinematic and differential GPS and also airborne Light Detecting And Ranging (LiDAR) data by Quigley et al. (2012).

The Christchurch earthquake is notable for the high ground accelerations it produced near the epicentre; Fry et al. (2011) calculated a peak ground acceleration of 2 g. Co-seismic ground displacements and slip distributions estimated by Beavan et al. (2012b) for the Darfield and Christchurch earthquakes are illustrated in Figure 1.2b and c; These figures are reproductions of Figure 2 from Beavan et al. (2012b).

The Cook Strait earthquake sequence of July-August 2013 began with two M_W 5.5 and 5.8 events, followed by a M_W 6.6 event of 21 July 2013, off-shore from Seddon, and an onshore M_W 6.6 earthquake on 16 August 2013, near Lake Grassmere (focal mechanisms shown in Figure 1.4). Notable ground shaking was reported in Wellington and Marlborough with a peak ground acceleration of 0.2 g in parts of Wellington (Holden et al., 2013). Coseismic ground displacements and slip distributions associated with the M_W 6.6 earthquakes were measured by Hamling et al. (2014) using continuous and campaign GPS data along with two interferograms generated from TerraSAR-X (TSX) and RADARSAT-2 acquisitions (Figure 1.2d). Hamling et al. (2014) estimate predominantly dextral strike-slip of up to 0.9 and 2.1 m associated with the first and second events on the fault surface, respectively. Figure 1.2 is a reproduction of Figure 3

Figure 1.2: a) Coseismic slip distribution of the 15 July Dusky Sound earthquake, horizontal and vertical fits to the continuous and campaign GPS data acquired during August 2009 (Figure 8 from Beavan et al. 2010a). b) Coseismic slip distribution of the 4 September Darfield earthquake and horizontal fit to the continuous and campaign GPS data (Figure 2 from Beavan et al. 2012b). c) Coseismic slip distribution and horizontal fit to the continuous and campaign GPS data following the 22 February Christchurch earthquake (Figure 2 from Beavan et al. 2012b; continued on the next page).

from Hamling et al. (2014). We summarise the previous studies on the seismicity in the South Island schematically in this figure. Further details can be found in the papers by Beavan et al. (2010a, 2012b) and Hamling et al. (2014).

Other recent large earthquakes in the South Island include a M_W 5.8 earthquake north-west of Wanaka in the Otago region ² on 4 May 2015, a M_W 6.0 event west of Arthur's pass ³ on 6 January 2015, a M_W 6.2 earthquake on 13 October 2014 in the Fiordland region ⁴ and a M_W 6.2 earthquake on 24 April 2015, south-east of the Saint Arnaud Range ⁵ in the northern South Island.

²<http://geonet.org.nz/quakes/region/newzealand/2015p332712>

³<http://geonet.org.nz/quakes/region/newzealand/2015p012816>

⁴<http://geonet.org.nz/quakes/region/newzealand/2014p770859>

⁵<http://geonet.org.nz/quakes/region/newzealand/2014p770859>

Figure 1.2: d) Observed and simulated RADARSAT-2 and TerraSAR-X interferograms for the Lake Grassmere earthquake (first and second columns). Observed (black) and modelled (red) horizontal and vertical GPS displacements due to the Lake Grassmere earthquake and sum of the horizontal displacements due the Cook Strait and Lake Grassmere earthquakes at five cGPS sites (third column), respectively (Figure 8 from Hamling et al. 2014).

1.3 Motivations and thesis objectives

This project focuses on monitoring and modelling a range of fault-related processes that can be observed using geodetic means. The seismic and aseismic behaviours of faults encompass a wide range of processes with varying temporal and spatial scales (Tullis, 2007). Quasi-static earthquake nucleation, the propagation of dynamic rupture and its arrest, spatio-temporal variations in stress and fault strength and aseismic slip are all active at different places and stages in a fault system. During the phases related to nucleation and propagation of a rupture, stresses can change rapidly in a few seconds, with spatial variation of the order of a metre. Conversely, tectonic loading occurs over hundreds to thousands of years within the fault zone, on scales of kilometres. The major problem with modelling the entire process of the earthquake machine is dealing with this large range of scales. Therefore, as illustrated in Figure 1.3, this study addresses a specific range of spatial and temporal scales. In addition many complexities are encountered in modelling the earthquake cycle. Different assumptions of rheology, changes of stress field produced by aseismic slip, heterogeneity within the fault zone, thermomechanical coupling, spatial variability of constitutive law parameters, and off-fault plasticity are some of the major sources of complexity during modelling. In this project, we only consider how different assumptions of rheology impact our modelling results.

We focus on the analysis and interpretation of geodetic data collected in the southern and central South Island of New Zealand during 2009-2014 and similar temporal and spatial scales highlighted in Figure 1.3. Our three primary research questions are stated below, with subordinate questions noted:

1. What can geodetic data reveal about postseismic displacement in response to multiple earthquakes?
 - (a) What circumstances or conditions are favourable in enhancing the effectiveness of Interferometric Synthetic Aperture Radar (InSAR) for measuring low-strain rate tectonic deformations?
2. What mathematical framework can we use to provide unique solutions of displacement and stress evolution for postseismic time intervals?
 - (a) How do we optimally choose rheology parameters in order to model the displacement and stress evolution?
 - (b) How can the uncertainty of input parameters for models such as fault geometry and slip distribution change the inferred stress changes?

Figure 1.3: Temporal and spatial scales associated with various fault-related processes (modified from Tullis 2007). The processes of interest for this study (outlined in red in the figure) span temporal scales of days to several years and spatial scales from metres to hundreds of kilometres.

(c) Is the inferred model of stress evolution unique?

3. Can we assess the triggering effect of each earthquake on the later events by modelling the stress and strain evolution? Particularly, can we calculate the strain and stress evolution following the Dusky Sound earthquake and examine its effects on the Canterbury earthquake sequence of 2010–2012?

1.4 Thesis outline

The remainder of this thesis is composed of three stand-alone chapters as well as methodology and summary chapters. The three primary Chapters 3, 4 and 5 are written for the purpose of subsequent publications. Therefore, for consistency, the first person plural (“we”) has been used throughout the thesis and in some places contains duplication of background and introduction material. The regions of study for the stand-alone chapters are shown with boxes in Figure 1.4.

Most of the work in this thesis has been undertaken by myself (Pegah Faegh-Lashgary)

with my supervisors' guidance unless explicitly stated. The titles and contents of the chapters are:

Chapter 2: Measuring, modelling and interpreting postseismic displacement

We first present an overview of contemporary space-geodesy techniques including satellite radar interferometry, known as InSAR and GPS with which to measure ground displacements associated with tectonic changes. This chapter next details basic theorems used to describe postseismic displacements in terms of 1) Viscoelastic relaxation; 2) Frictional afterslip; 3) Poroelastic rebound and 4) Dilatancy recovery. The governing equations for simple geometries and homogeneous materials for different mechanisms are discussed in analytical forms. The final part of the chapter describes semi-analytical approaches using propagator matrix methods and Finite Element Methods for fully numerical computation of stress-displacement vectors.

Chapter 3: Postseismic displacement following the M_W 7.8 Dusky Sound earthquake, Fiordland, New Zealand, and its effect on the M_W 7.1 Darfield (Canterbury) earthquake 14 months later

We describe the tectonic setting of southern New Zealand in the transition from the strike-slip Alpine Fault to the subduction interface of the Puysegur Trench in the Fiordland region (Figure 1.4). The postseismic displacement, as measured by continuous GPS in the time interval between the Dusky Sound and Darfield earthquakes along with campaign GPS data acquired during August 2009 and Differential Interferometric Synthetic Aperture Radar (DInSAR) observations, is modelled using an iterative approach. Prof. Euan Smith processed the continuous GPS data. The campaign GPS data acquired during August 2009 was published by Beavan et al. (2010a). We take advantage of the Principal Component Analysis-based Inversion Method (PCAIM) by which time-dependent slip models are estimated by decomposition of the postseismic displacement into space and time functions. Different values of viscosity yield different afterslip models for the fault embedded in an elastic-viscoelastic medium. Strain and Coulomb stress changes in the vicinity of the Darfield earthquake are calculated and discussed in terms of further effects on the Canterbury earthquake sequence of 2010–2012. Chapter 3 has been submitted to *Geophysical Journal International* in

collaboration with John Townend, Charles Williams and Euan Smith.

Chapter 4: Coupling time-dependent afterslip with viscoelastic rebound using analytical and hybrid numerical Green's functions

We introduce a new inversion framework with which to split the postseismic signal into the viscoelastic responses to the coseismic and postseismic stress changes and afterslip. The novelty of the inversion technique is accounting for the corresponding viscoelastic responses to coseismic event and time-dependent afterslip. We develop the Principal Component Analysis-based Inversion Method used in Chapter 3 (Kositsky and Avouac, 2010) to form the Iterative Decoupling of Afterslip and Viscoelastic relaxation (IDAV) method in Chapter 4. In contrast with the conventional PCAIM inversion method, we use elastic and time-dependent elasto-viscoelastic Green's functions rather than instantaneous elastic Green's functions in order to relate a slip vector into displacement terms. The new approach (IDAV) is examined with analytically and numerically simulated displacement fields associated with a simple elastic-viscoelastic medium and an arbitrary geometry representing a subduction earthquake, respectively. We also apply the IDAV method to the observed postseismic displacement following the Dusky Sound earthquake. Chapter 4 has been submitted to *Computers and Geosciences* in collaboration with John Townend and Charles Williams.

Chapter 5: Observations and modelling of postseismic displacement in Canterbury, New Zealand, following the 4 September 2010 Darfield and 22 February 2011 Christchurch earthquakes

We explain the principal ground displacement features observed following the Darfield and Christchurch earthquakes in terms of poroelastic rebound, dilatancy recovery and afterslip. Dr. Sigrún Hreinsdóttir processed the campaign GPS data acquired annually from 2012 to 2015. The afterslip models involve potentially a subsurface lineament manifested as a band of seismicity, the western faults, associated with the Darfield earthquake, and the eastern faults, associated with the Christchurch earthquake as well as its aftershocks. Chapter 5 has been submitted to *Geophysical Journal International* in collaboration with John Townend, Ian Hamling, Mahdi Motagh, Sigrún Hreinsdóttir and Paul Denys.

Figure 1.4: The Fiordland region: case area of Chapter 3 and 4. The Canterbury region: case area of Chapter 5. Focal mechanisms of recent earthquakes in the Fiordland region, namely the 1989 Doubtful Sound, 1993 Secretary Island, 2000 Thompson Sound, 2003 Fiordland and 2009 Dusky Sound earthquakes, in the Canterbury region, namely the 2010 Darfield and 2011 Christchurch earthquakes and the Cook Strait earthquake sequence which were widely felt in South Island of New Zealand.

Chapter 6: Summary and recommendations

Appendix A: Coulomb stress calculation

Appendix A explains the transformation of a stress tensor from the Cartesian coordinate system to shear, normal and Coulomb components of stress changes. For our Coulomb stress calculation in Chapter 3, we assume an isotropic poroelastic rheology for the frictional model (Harris, 1998) with Skempton's and frictional coefficients of 0.5 and 0.6, respectively.

Appendix B: Global compilation of viscosity, rheology and depth of the top viscoelastic layer

In Chapter 2 and Appendix B, we review different viscoelastic rebound models with a broad range of rheology and viscosity values with which other groups have modelled postseismic signals observed globally. In Appendix B, we summarise the region of study used in 88 papers and metadata of different earthquakes, as well as different assumptions about rheology, vertical or lateral heterogeneities and the viscosity for each layer. This appendix is referred to in detail in Chapter 2.

Appendix C: Post-Dusky Sound supplementary files

We illustrate fits to the post-Dusky Sound cGPS displacement time series with a viscosity of 2×10^{18} Pa·s modelled by the IDAV method.

Appendix D: Post-Canterbury supplementary files

We present coseismic displacements associated with the 13 June and 23 December 2011 earthquakes in Christchurch and the standard deviation of the post-Christchurch mean Line Of Sight (LOS) velocity acquired from TSX images.

Chapter 2

Measuring and modelling postseismic displacement

This chapter has two major sections, addressing the measuring and modelling of post-seismic displacement. The detailed mathematical background and methods discussed in these sections compliment the stand-alone result chapters that follow (Chapters 3, 4 and 5). The first section describes the fundamentals of satellite-based geodetic techniques, including Interferometric Synthetic Aperture Radar (InSAR) and the Global Positioning System (GPS) that can be used to measure postseismic displacement. Secondly we discuss analytical and numerical models of observed displacements occurred on various temporal and spatial scales.

2.1 Measuring postseismic displacement

2.1.1 Interferometric Synthetic Aperture Radar (InSAR)

InSAR is a remote sensing technique used to measure displacement of the ground surface. Centimetre-scale movements can be inferred from measurements of the range change between the ground and antenna by differencing the phase from two radar images acquired at different times (e.g. Massonnet et al., 1993; Bamler and Hartl, 1998; Bürgmann et al., 2000; Rosen et al., 2000). The distance between two different sensors or the same sensor at different times is referred to as the baseline (Ferretti et al., 2007). The geometry of imagery for a pixel acquired by master and slave sensors with a baseline distance B is illustrated in Figure 2.1. The variation of phase difference in this geometry from a ground reference point to its neighbour is approximately

$$\Delta\phi = \frac{2\pi\Delta r}{\lambda} = \frac{4\pi}{\lambda} \frac{B_{\perp}q_s}{R} \quad (2.1)$$

where

$$\Delta r = \Delta(\Delta r_0) \quad (2.2)$$

Figure 2.1: Schematic illustration of interferometric geomtry. All measurements are projected on the slant range. The distance between master and slave sensors is referred to as baseline with two perpendicular and parallel components $B_{\perp}$ and $B_{\parallel}$. For each target the perpendicular distance of the target to the slant range is q_s and the projected distance on the slant range is s (Ferretti et al., 2007).

The variation of phase difference $\Delta\phi$ is proportional to the variation of path length difference Δr acquired from two Synthetic Aperture Radar (SAR) sensors between two points (p_0 ; reference and p_1 ; target). Δr_0 is the reference path length difference acquired from master and slave sensors. λ is the wavelength of the signal and $B_{\perp}$ is the perpendicular component of the baseline. All the measurements are projected to the slant range shown by R in Figure 2.1. The altitude of the target, perpendicular distance from the slant range and slant range distance are represented by variables q , q_s and s , respectively. The temporal baseline is the time difference between two sensor acquisitions. Temporal (B_{temp}) and perpendicular ($B_{\perp}$) baselines are the primary factors controlling the accuracy of InSAR processing (Ferretti et al., 2007).

The phase difference variation in Figure 2.1 consists of two components (Ferretti et al., 2007)

$$\Delta\phi = \Delta\phi_{topo} + \Delta\phi_{geom} \quad (2.3)$$

where

$$\Delta\phi_{topo} = -\frac{4\pi}{\lambda} \frac{B_{\perp}q}{R \sin \theta} \quad (2.4)$$

and

$$\Delta\phi_{geom} = -\frac{4\pi}{\lambda} \frac{B_{\perp}s}{R \tan \theta} \quad (2.5)$$

The example shown in Figure 2.1 has a simple interferometric geometry in which we assume there is only one dominant point scatterer within a pixel. Multiple scatterers within a pixel contribute to the phase value of the pixel (Ferretti et al., 2000, 2001). The multiple contributions impose randomness in the phase value of a Single Look Complex (SLC). By subtracting the phase value from two SLCs acquired at two epochs, this random contribution is removed, but it is subject to unchanged back-scattering characteristics of the points within the pixel between two epochs (Ferretti et al., 2000, 2001). Thereafter, the differences between the atmospheric path delay in two epochs ϕ_{atm} , displacement ϕ_{def} , and noise ϕ_{noise} appear on the interferometric phase in addition to the differences between viewing geometry ϕ_{geom} and topography ϕ_{topo} (Ferretti et al., 2000, 2001). The interferometric phase then is written as

$$\Delta\phi = (\Delta\phi_{topo} + \Delta\phi_{geom}) + (\Delta\phi_{atm} + \Delta\phi_{def} + \Delta\phi_{noise}) \quad (2.6)$$

and

$$\Delta\phi_{def} = \frac{4\pi}{\lambda} d \quad (2.7)$$

Figure 2.2 illustrates how the radar interferometric phase changes with ground displacement d . Two SAR images acquired prior to and following a displacement-inducing event measure the phase difference between the two epochs, which is converted to ground displacement using Equation 2.7. However, as shown in Equation 2.6, the phase difference consists of other components in addition to the displacement phase. The interferometric geometry in which the data are acquired, the topography of the ground and atmospheric and orbital errors introduce other components to the phase difference that we must remove to isolate the displacement phase (see Bürgmann et al., 2000; Elachi and Van Zyl, 2006; Ferretti et al., 2007; Dzurisin and Mori, 2007). We remove components related to geometry and topography during InSAR processing. We calculate the effects of viewing geometry using orbital data during the step known as interferogram flattening (Figure 2.3a). The topographic phase is calculated using either a Digital Elevation Model (DEM) spanning the entire region (two-pass differential interferometry) or utilising a pair of images acquired before the event (three-pass differential interferometry; Figure 2.3b). We can use the DEM product of the Shuttle Radar Topography Mission (SRTM) for two pass differential interferometry. The effects of

Figure 2.2: The principle of phase measurement due to a surface displacement in radar interferometry.

atmospheric and orbital errors on the displacement signal are mitigated by applying several post-processing methods (Figure 2.3c; see Bürgmann et al., 2000; Elachi and Van Zyl, 2006; Ferretti et al., 2007; Dzurisin and Mori, 2007).

InSAR time series techniques

InSAR has a unique advantage over other geodetic methods in its ability to measure surface displacement with millimetre- to centimetre-scale accuracy over large areas (on the order of hundreds of square kilometres; e.g. Massonnet et al., 1993; Bamler and Hartl, 1998; Bürgmann et al., 2000; Rosen et al., 2000). Temporal and baseline decorrelation and variations due to atmospheric conditions are major limitations of conventional InSAR techniques (Bürgmann et al., 2000). The effect of decorrelation is reduced by combining multiple SAR observations using multi-temporal processing techniques. Using more than two SAR scenes acquired over the same area leads to redundant measurements that are used with time series methods to produce crustal displacement maps (Ferretti et al., 2001; Kampes, 2005; Berardino et al., 2002).

InSAR time-series techniques are broadly classified into two categories: 1) Permanent Scatterer-InSAR (PS-InSAR); and 2) Small Baseline Subset (SBAS) methods.

Figure 2.3: Interferometry workflow (Dzurisin and Mori, 2007). a) Interferogram flattening b) Differential interferometry c) Spatial filtering. Each fringe in the filtered interferogram represents displacement of the ground surface towards the satellite by an amount equal to half of the wavelength.

PS-InSAR works by identifying image pixels in a stack of interferograms that remain coherent over long time intervals, with little variation of their scattering properties with either imaging geometry or time (Ferretti et al., 2000; Hooper et al., 2004). The first implementation of PS-InSAR worked well in urban areas where man-made structures behave like strong corner reflectors, but fails to perform satisfactorily in vegetated and mountainous terrains (Ferretti et al., 2000, 2001). The Stanford Method for Permanent Scatterers (StaMPS) technique, developed by Hooper et al. (2004) and modified by Hooper (2006), was the first PS-InSAR method designed to estimate displacement signals even in the absence of very bright scatterers, using Slowly Decorrelating Filtered Phase (SDFP) pixels. Filtered phase values of SDFP pixels decorrelate very little over short time intervals. This method was initially used to analyse episodic crustal displacements in non-urban environments, with application to volcanic settings. The SBAS technique (Berardino et al., 2002) minimises the effects of decorrelation on displacement signals by constructing all possible interferograms in the study area with short temporal and spatial baselines. Using SBAS techniques, velocity maps and time series of coherent pixels are produced (Lanari et al., 2007). The time-series analysis in the SBAS method is performed using the least-squares inversion methodology.

Examples of InSAR time series

Here we present examples of ground displacements in Christchurch prior to and following the M_W 7.1 Darfield earthquake on 4 September 2010. This is done using Envisat and CSK acquisitions and PS-InSAR method combined with SBAS. Note that, these results are not used in other chapters for further modelling.

We measure the ground displacement for the first dataset using satellite radar interferometry including 36 Envisat SAR acquisitions in ascending orbits from February 2004 to July 2010. The second dataset spans the time interval between October 2010 and February 2011 and includes 12 CSK acquisitions in ascending orbits. We process both datasets in the same way. After reading and cropping the SLCs, we extract the satellite orbits for all images and apply DEM offset corrections using the LINZ¹ 40 m DEM. All interferograms for each dataset are coregistered and resampled and all master-slave interferograms are produced using the DORIS software package (Kampes and Usai, 1999). The SBAS networks corresponding to the Envisat and CSK datasets consist of 38 and 117 interferograms, respectively. For the Envisat dataset, PS pixels are selected using the default values of PS-selection and weeding parameters of StaMPS (Hooper and Zebker, 2007; Hooper, 2008; Hooper et al., 2010). For the CSK dataset, pixels

¹Land Information New Zealand; <https://data.linz.govt.nz>

with coherence above a threshold of 0.2 are selected as Permanent Scatterer (PS). After 3D phase unwrapping the times series of displacement and the associated velocity are estimated using StaMPS (Hooper and Zebker, 2007; Hooper, 2008; Hooper et al., 2010).

Figure 2.4 (the Envisat dataset) shows a subsidence signal over western Christchurch that occurred over a long period of almost 6 years prior to the Canterbury sequence of 2010–2012. The subsidence occurs immediately west of the transition from confined to unconfined aquifers, in an area where the water table is locally deep (exceeding 6 m), unconfined groundwater is downwards flowing and levels are highly variable (Cox et al., 2012; Van Ballegooy et al., 2013; Figure 2.4a and b).

Figure 2.5 (the CSK dataset) represents accumulated displacement due to a number of aftershocks (red stars) particularly on the December 26th, 2010, which impacted the Central City (Bannister and Gledhill, 2012). Several aftershocks shown in Figure 2.5 have magnitudes larger than M_w 4.0. These results demonstrate how SBAS and PS-InSAR time series technique capture small magnitudes of displacements in Christchurch.

Figure 2.4: a) Mean line of sight velocity measured from ASAR/Envisat acquisitions prior to the 4 September 2010 Darfield earthquake. The colors represent displacement rates in the LOS directions: red colors indicate movement away from satellite and blue colors indicate movement towards the satellite. Transition from confined to unconfined aquifers is inferred from Cox et al. (2012) b) Time series of displacement at point S₁ in Christchurch extracted from the Envisat dataset. Faults are as described in Figure 2.5.

Figure 2.5: Mean line of sight velocity measured from COSMO-SkyMed ascending orbits following the Darfield and prior to the Christchurch earthquakes. Red stars represent aftershocks between October 2010 and February 2011. For permanent scatterers, the colours represent displacement rates in the LOS direction: red indicates movement away from satellite and blue indicates movement towards the satellite. H) Western ENE I) NNE J) Eastern faults related to the 22 February Christchurch earthquake, K) SW-NE L) NW-SE Faults related to the 13 June 2011 aftershock and M) SW-NE Fault related to the 23 December 2011 aftershock.

2.1.2 Global Positioning System (GPS)

In GPS techniques, distances between satellites and receivers are measured to compute point positions on the Earth surface. Similar to trilateration, GPS surveys require known positions of control points (satellite positions). The position of the receiver in the Cartesian coordinate system (u_x, u_y, u_z) and the receiver's clock offset dT_u are four unknown parameters that are estimated from measuring distances between receiver and a minimum of four other satellites (see Remondi, 1985a,b, 1988, 1991; Figure 2.6a).

The two observable variables in GPS surveys are the pseudo-range for lower-accuracy and the carrier phase for high-precision surveys. Pseudo-range p and carrier phase ϕ

Figure 2.6: a) A navigation solution for GPS measurements of the position and receiver clock with a minimum of four other satellites b) Multipath is highlighted by blue dashed lines c) Techniques of single between-receivers indicated by pale blue area, between-satellites by pale green are, double difference by pale pink area and triple difference by the overall area (Van Sickle, 2008).

observations for each satellite-receiver pair with the distance of ρ are

$$p = \rho + d_\rho + c(dt - dT) + d_{ion} + d_{trop} + \epsilon_{m_p} + \epsilon_p \quad (2.8)$$

$$\phi = \rho + d_\rho + c(dt - dT) + \lambda N - d_{ion} + d_{trop} + \epsilon_{m_\phi} + \epsilon_{phi} \quad (2.9)$$

respectively, where N is the the number of full wavelengths λ associated with the modulated carrier. These calculations include several biases in observation equations such as: orbital errors d_ρ , ionospheric d_{ion} and tropospheric effects d_{trop} , receiver noises $\epsilon_p, \epsilon_\phi$, multipath $\epsilon_{m_p}, \epsilon_{m_\phi}$ and satellite and receiver clock errors $(dt - dT)$. ρ is pseudo-range or carrier phase value excluding the biases in observations.

A satellite clock error dt is monitored by the Control Segment which broadcasts clock corrections as part of the navigation message. The navigation message includes all parameters required to perform a positioning service, which is sent back to the users (see Misra and Enge, 2006; Van Sickle, 2008). Effects of refraction and reflection on the direction and speed of the signal while it travels through atmosphere is another source of bias. These errors are larger for a signal travelling from satellites located near the receiver's horizon than its zenith, due to the amount of time that a signal interacts with the ionosphere while it travels from horizon (see Misra and Enge, 2006; Van Sickle, 2008).

The magnitude of ionospheric delay can be determined from the Total Electron Content (TEC) in the ionosphere, which has a daily cycle (see Coco, 1991; Borries et al., 2007). The ionospheric effect is dispersive, meaning that it is correlated with the frequency of the signal. This effect is therefore modelled and removed for dual-frequency receivers using L1 (1575.42 MHz) and L2 (1227.60 MHz) frequencies. For single-frequency receivers, ionospheric model parameters in the navigation message are useful to partially (but not completely) remove ionospheric refraction effects.

Satellite positions are obtainable from broadcast ephemerides or final orbits generated by analysis centres a few weeks after the observation based on tracking satellites. A precise satellite position leads to a precise estimation of receiver position (e.g. Malys et al., 1997; Van Sickle, 2008). Associated errors in the estimation of satellite position remain as orbital errors d_ρ in the final position measurement.

In contrast to the ionospheric effects, tropospheric errors are non-dispersive with similar effects on L1 and L2 frequencies. The pressure-correlated dry component contributes more to the total tropospheric errors and it is easier to be estimated. The wet component of the tropospheric effects requires information of water vapour content in the troposphere (see Dodson et al., 1996; Van Sickle, 2008).

Errors due to multipath occur when signals are reflected from nearby objects and inter-

ferre with the direct signal from the satellite (Figure 2.6b). Some receivers identify and reject weaker and more diffuse reflected signals or reverse circular polarization. Minimising the effect of multipath is also possible via GPS antenna design (see Van Sickle, 2008).

Differencing

Single, double and triple differences are the techniques most frequently used to mitigate or remove different position measurement biases. Satellite clock and orbital errors are assumed to be very similar in measurements from the same satellite but different receivers when using the technique called between-receiver single difference (Figure 2.6c, pale blue polygon). In a single between-satellite difference technique, receiver clock errors are eliminated by differencing simultaneous measurements from two satellites (Figure 2.6c, pale green polygon). Atmospheric effects also remain quite similar for satellites with a common receiver, so this technique somewhat mitigates the effects of ionospheric and tropospheric errors. With a double difference technique using a combination of both between-satellite and between-receiver difference techniques, we remove both satellite and receiver clock errors (Figure 2.6c, pale pink polygon). Two double difference techniques combined at two different epochs are referred to as a triple difference by which the integer cycle ambiguity N (Equation 2.9) is quantifiable (see Parkinson and Enge, 1996; Misra and Enge, 2006; Van Sickle, 2008; Figure 2.6c).

2.2 Modelling postseismic displacement

Postseismic displacements following large earthquakes have generally been modelled using four mechanisms: 1) long-term viscoelastic relaxation in the lower crust/upper mantle (e.g. Elsasser, 1967; Savage and Prescott, 1978; Pollitz et al., 2000; Pollitz, 2003b; Rousset et al., 2012a); 2) short-term afterslip in the upper/lower crust on the mainshock fault surface (e.g. Perfettini and Avouac, 2007; Ryder et al., 2010b; Johnson et al., 2012; Lubis et al., 2013); 3) poroelastic rebound (e.g. Booker, 1974; Rice and Cleary, 1976; Peltzer et al., 1998; Wang and Kümpel, 2003; Hu et al., 2014); and 4) shallow fault zone dilatancy recovery (e.g. Fielding et al., 2009). We interpret the observed postseismic displacements following the 15 July 2009 Dusky Sound earthquake and the Canterbury earthquake sequence of 2010–2012 in terms of these four mechanisms. In the following sections, we focus on the principles employed to derive the governing equations for each postseismic mechanism. In order to derive analytical solutions for stress-displacement vectors, we often assume a simple geometry and homogeneous

properties. The components of stress-displacement vectors are combinations of stress and displacement. We discuss forward analytical models in detail.

In addition, we need inversion techniques to interpret geodetic observations. Observations of time-dependent postseismic displacement using satellite-geodesy techniques enable us to analyse time-dependent mechanisms. In this chapter, we describe methods of time-dependent inversion of postseismic displacement. More details are presented in Chapters 3 and 4 of this thesis.

More realistic postseismic models require consideration of complex properties, such as an irregular non-planar medium (e.g. Pollitz, 1992), multiple relaxation time scales (Hetland and Hager, 2005), lateral heterogeneity (Pollitz, 2003a), vertical layering (Hetland and Hager, 2005) and the effects of coupling postseismic mechanisms with gravity (Pollitz, 1997; Rundle, 1980; Iwasaki, 1985). Deriving analytical solutions for complex problems is often impossible. Consequently, semi-analytical approaches and numerical models are extensively used to approximate complex mathematically-defined models (e.g. Haskell, 1953; Schapery, 1967; Peltier, 1985; Wang and Kumpel, 2003; Fukahata and Matsu'ura, 2006; Aagaard et al., 2009, 2013). We review the basic ideas behind the semi-analytical and numerical methods employed in this thesis. In particular, the Finite Element Method (FEM) using the numerical code PyLith (Aagaard et al., 2009, 2013, 2015) and the propagator matrix method of Fukahata and Matsu'ura (2006) are used to compute numerical and semi-analytical viscoelastic Green's functions in Chapters 3 and 4. In Chapter 5, we adopt the framework developed by Wang and Kumpel (2003) for computing the effects of multi-layered poroelastic rebound using an extension of Haskell's propagator method and the technique used by Fielding et al. (2009) for estimating shallow fault zone dilatancy recovery process following the Canterbury 2010–2012 earthquake sequence.

2.2.1 Viscoelastic relaxation of ductile materials

Viscoelastic effects during the postseismic time interval relax the deviatoric stress induced coseismically (e.g. Freed and Lin, 1998, 2001; Rydelek and Sacks, 2001; Freed, 2005; Pollitz et al., 2006a; Freed et al., 2007; He et al., 2011; Decriem and Árnadóttir, 2012).

The brittle-ductile transition of rocks occurs with increasing temperature at depths. The ductile behaviour of rocks in the lower crust and upper mantle leads to diffusion and dislocation creep processes (see Karato and Wu, 1993; Fliervoet et al., 1999; van Hunen et al., 2005). Mantle convection, which controls the cooling of the earth, is a result

Figure 2.7: Schematic illustration of atomic displacement and lattice deformation from a) Edge dislocation creep. Compression, tension and shear zones are presented in boxes. An extra half-plane occurs in the compression zone. b) Purely shear screw dislocation creep.

of these processes. The creep processes also alter the stress induced by earthquakes, producing postseismic viscoelastic relaxation (see Turcotte and Schubert, 2014).

Creep processes are defined based on Fick's law (Fick, 1855). Atoms tend to move from high to low density, in order to reach the most stable state with evenly distributed vacancies in crystals. At lower stresses, diffusion creep dominates and rocks behave as a Newtonian fluid, with linear pressure- and temperature-dependent viscosities (Kundu and Cohen, 1990; Batchelor, 2000; Kirby, 2010). At higher stresses, edge and screw dislocations are major forms of dislocation creep that cause a power-law or non-Newtonian behaviour of fluids (Astarita and Marrucci, 1974). Dislocations cause atomic-scale deformations in crystals. Edge dislocations in crystals yield an extra half-plane in the lattice (Figure 2.7a, ellipse). Furthermore, edge dislocations produce a region of compression where the extra half-plane occurs, a tension field and a shear region (Figure. 2.7a). In contrast, screw dislocations cause an offset in the lattice and a purely shear dislocation (Figure. 2.7b) region in crystals. If the dislocation creep persists, the viscosity follows a stress-dependent power-law rheology. Laboratory studies of olivine, the primary mineral of the mantle, show that dislocation creep is the dominant process (see Karato et al., 1986; Karato and Wu, 1993; Turcotte and Schubert, 2014). However, mantle strain rates are much lower than laboratory strain rates (see Paterson, 1987; Turcotte and Schubert, 2014). It is not inferred from laboratory studies whether the dominant process is the dislocation creep, nor whether the mantle behaves as a non-Newtonian fluid (see Turcotte and Schubert, 2014).

The general viscosity η for either dislocation or diffusion creep is expressed as

$$\eta = C \left(\frac{\sigma}{\mu} \right)^{1-n} \exp \left(\frac{E_a + pV_a}{RT} \right) \quad (2.10)$$

where

$$C = \frac{1}{2} \left(\frac{\mu}{A} \right) \left(\frac{h}{b} \right)^m \quad (2.11)$$

Here C is the pre-exponential factor, n is a power, μ is the shear modulus, σ is the stress magnitude, E_a and V_a are the activation energy and activation volume per mole, R is the universal gas constant and p and T are pressure and absolute temperature (Kirby and Kronenberg, 1987; Carter and Tsenn, 1987). In Equation 2.11, b and h are the lattice spacing and grain size, respectively, m is a power and A is a constant. For diffusion creep, $n = 1$ and $m = 2.5$ are appropriate powers. In the particular case where dislocation creep is independent of the grain size and the lattice spacing, $m = 0$ and $n = 3.5$. We infer from Equation 2.10 that regardless of the dominant creep mechanism, the viscosity is exponentially proportional to the pressure and inverse of absolute temperature. Furthermore, if $n \neq 1$ the equation implies stress dependent viscosity for the mechanism of dominant dislocation creep (see Turcotte and Schubert, 2014).

Elsasser (1969) considered the viscous behaviour of the asthenosphere due to diffusion and dislocation creep processes with a simple one-dimensional model, assuming an elastic layer overlying a viscous channel. In this model, the elastic behaviour of the asthenosphere at short time scales was not considered. Several researchers have since attempted to model the viscoelastic relaxation of ductile materials, considering the elastic behaviour of the asthenosphere, with an elastic layer overlying a viscoelastic half-space. This two-layer model is the most widely used representation of the brittle-ductile transition (e.g. Flügge, 1975; Iwasaki, 1985; Pollitz, 1997; Hearn et al., 2002b; Hu et al., 2004; Pollitz, 2003a; Fukahata and Matsu'ura, 2006). We use this representation in Chapters 3 and 4, where we only consider linear viscoelastic materials (Newtonian fluids) for the viscoelastic half-space with dominant diffusion creep, and $n = 1$.

Correspondence Principle

We use the Correspondence Principle to achieve analytical solutions to viscoelastic problems (e.g. Nur and Mavko, 1974; Savage and Prescott, 1978). Stresses arising from a Newtonian fluid flow are linearly proportional to the strain rate (Kundu and Cohen, 1990; Batchelor, 2000; Kirby, 2010). The Laplace transform of this linear constitutive law is expressed as a linear relation between transformed stress and strain. Conse-

quently, the governing equations of viscoelasticity after applying the Laplace transform, become similar to the corresponding elastic governing equations. This statement is known as the Correspondence Principle. The derivation of this principle is as follows

The constitutive law between shear stress σ and shear strain ϵ for linear viscoelastic media is expressed as

$$\Phi\sigma = \Psi\epsilon \quad (2.12)$$

Depending on the rheology of the media

$$\Phi = \sum_{k=0}^{q_\phi} \phi_k \frac{d^k}{dt^k} \quad (2.13)$$

and

$$\Psi = \sum_{k=0}^{q_\psi} \psi_k \frac{d^k}{dt^k} \quad (2.14)$$

where ϕ_k and ψ_k are the stress and strain coefficients and determined by their rheological properties (Flügge, 1975; Findley et al., 1977). The Laplace transform of equations 2.13 and 2.14 reveals the Correspondence Principle

$$\hat{\Phi} = \hat{\Phi}(s) = \sum_{k=0}^{q_\phi} \phi_k s^k \quad (2.15)$$

$$\hat{\Psi} = \hat{\Psi}(s) = \sum_{k=0}^{q_\psi} \psi_k s^k \quad (2.16)$$

where the Laplace transform of a function $f(t)$ is the function $F(s)$

$$F(s) = \int_0^\infty \exp(-st) f(t) dt \quad (2.17)$$

The time derivatives of equations 2.13 and 2.14 are transformed to linear combinations of the Laplacian variable s in equations 2.15 and 2.16.

Rheology of the viscoelastic half-space

The stress and strain coefficients ϕ_k and ψ_k in Equations 2.13 and 2.14 are defined by the rheology of the viscoelastic half-space (see Hetland and Hager, 2005; Segall, 2010). The Newtonian rheology of the viscoelastic half-space is represented with different configurations of springs (elasticity) and dashpots (viscosity). A spring in series with a dashpot represents a simple univiscous Maxwell rheology (Figure 2.8a), which reproduces a non-recoverable creep. In contrast, a spring in parallel with a dashpot (Figure 2.8b) represents a recoverable Kelvin rheology.

The simplest model that describes both Kelvin and Maxwell materials is a Standard

Figure 2.8: Schematic illustration of different rheology models: a) Maxwell rheology: a spring in series with a dashpot b) Kelvin rheology: a spring in parallel with a dashpot.

Linear Solid (SLS) model (Table 2.1; Wang et al., 2006; Hetland and Hager, 2005). A Maxwell element in series with a Kelvin element makes a bi-viscous Burger's body with two phases of viscoelastic creep (Table 2.1; Hetland and Hager, 2005; Pollitz, 2003b). Either by increasing the number of vertical layers or using a Burger's body, postseismic mechanisms with two mechanical time scales or more can be reproduced. However, heterogeneous rheology in which a weaker material is embedded within a stronger medium can also cause multiphases in postseismic signals (Hetland and Hager, 2005).

We summarise mechanical analogue models in terms of springs and dashpots and corresponding stress-strain constitutive laws using equations presented by Hetland and Hager (2005) and Segall (2010) in Table 2.1, for Maxwell and SLS materials, and for a Burger's body. For instance, the constitutive law for a Burger's body is written as

$$\sigma + a_1 \dot{\sigma} + a_2 \ddot{\sigma} = a_3 \epsilon + a_4 \dot{\epsilon} \quad (2.18)$$

The material behaviour of a Burger's body depends on the stress, the first and second derivatives of the stress, strain, and the first derivative of the strain. The coefficients a_1 , a_2 , a_3 and a_4 are functions of shear modulus and viscosity.

We adopt a univiscous Maxwell rheology and a Burger's body in Chapter 3 and 4, to analyse postseismic displacement following the Dusky Sound earthquake.

Analytical viscoelastic solutions

The Correspondence Principle simplifies solutions for a viscoelastic model, represented with an elastic layer overlying a viscoelastic half-space, to a two-layer elastic model. This elastic model consists of upper and lower layers with different elastic properties. Replacing the elastic moduli with the Laplace transformed moduli introduces a Lapla-

Table 2.1: Mechanical analogue model and material behaviour for viscoelastic materials

	Mechanical analogue model	Material behaviour
Maxwell		$\dot{\epsilon} = \left(\frac{\dot{\sigma}}{2\mu} \right) + \left(\frac{\sigma}{2\eta} \right)$
Standard Linear Solid		$\begin{aligned} \sigma + \frac{\eta}{\mu_1} \dot{\sigma} \\ = \\ 2\mu_2 \epsilon + 2\eta \left(1 + \frac{\mu_2}{\mu_1} \right) \dot{\epsilon} \end{aligned}$
Burger's		$\begin{aligned} \sigma + a_1 \dot{\sigma} + a_2 \ddot{\sigma} \\ = \\ a_3 \epsilon + a_4 \dot{\epsilon} \end{aligned}$

cian variable s in the corresponding viscoelastic solutions (Equation 2.22). Taking the inverse Laplace transform, converts the Laplacian variable s to a time variable t . Therefore, in contrast with instantaneous elastic solutions, viscoelastic solutions are time-dependent.

For a model consisting of elastic upper and lower layers with different elastic properties Chou (1966) and Rybicki (1971) derived Equation 2.19, for the one-dimensional displacement field $d(x)$ due to strike-slip motion on a fault.

$$d(x) = \left(\frac{\Delta u}{\pi} \right) \left(\pm \left(\frac{\pi}{2} \right) - \tan^{-1} \left(\frac{x}{D} \right) + \sum_{n=1}^{\infty} \left(\frac{\mu_c - \mu}{\mu_c + \mu} \right)^n \tan^{-1} \left(\frac{2Dx}{4n^2 H^2 - D^2 + x^2} \right) \right) \quad (2.19)$$

where Δu is slip on the fault with a maximum depth D , embedded in the upper elastic layer with thickness H , and $\pm$ represents the sign of the position x . μ_c and μ are the shear moduli of the layers. For a material with a linear viscosity, the Laplace transform of Equation 2.19 is expressed as Equation 2.21. For different rheological assumptions, $\hat{\gamma}(s)$ (Equation 2.20) is first calculated and then the inverse Laplace transform of Equation 2.21 is taken. The analytical solutions obtained with this method are calculated for specific cases of a strike-slip fault within an elastic layer overlying a

Table 2.2: Stress and strain coefficients in equations 2.13 and 2.14, specific forms of $\hat{\gamma}^n(s)$ and the corresponding inversely transformed forms of $\gamma^n(t)$ for Maxwell and SLS materials.

	Maxwell	SLS
Laplace coefficients	$\phi_0 = 1; \psi_0 = 0$ $\phi_1 = \left(\frac{\eta_m}{\mu_m}\right) = \tau_M; \psi_1 = \eta_m$	$\phi_0 = \frac{\mu_v + \mu_e}{\mu_v}; \psi_0 = \mu_e$ $\phi_1 = \left(\frac{\eta_v}{\mu_v}\right) = \tau_K; \psi_1 = \mu_e \tau_K$
Specific form of $\hat{\gamma}^n(s)$	$\hat{\gamma}^n(s) = \left(\frac{1}{2\tau_m}\right)^n \left(\frac{1}{s+1/2\tau_m}\right)^n$	$\hat{\gamma}^n(s) = \left(\frac{\mu}{2\mu_v \tau_K}\right)^n \left(\frac{1}{s + \frac{2\mu_v + \mu}{2\mu_v \tau_K}}\right)^n$
Specific form of $\gamma^n(t)$	$\gamma^n(t) = \left(\frac{1}{2\tau_m}\right)^n \frac{1}{(n-1)!} t^{n-1} \exp(-t/2\tau_m)$	$\gamma^n(t) = \left(\frac{1}{2\tau_K}\right)^n \frac{1}{(n-1)!} t^{n-1} \exp(-t/2\tau_K)$

linear viscoelastic half-space (Equation 2.22).

$$\hat{\gamma}(s) = \begin{pmatrix} \hat{\mu}_c - \hat{\mu} \\ \hat{\mu}_c + \hat{\mu} \end{pmatrix} \quad (2.20)$$

$$\begin{aligned} \hat{d}(x, s) = \left(\frac{\Delta \hat{u}(s)}{\pi}\right) \left(\left(\frac{\pi}{2}\right) - \tan^{-1}\left(\frac{x}{D}\right)\right) + \\ \left(\frac{1}{\pi}\right) \left(\sum_{n=1}^{\infty} \Delta \hat{u}(s) \hat{\gamma}^n(s) \tan^{-1}\left(\frac{2Dx}{4n^2 H^2 - D^2 + x^2}\right)\right) \end{aligned} \quad (2.21)$$

Finally, taking the inverse Laplace transform of Equation 2.21 leads to

$$d(x) = \left(\frac{\Delta u}{\pi}\right) \left(\frac{1}{2} - \tan^{-1}\left(\frac{x}{D}\right)\right) + \left(\frac{1}{\pi}\right) \sum_{n=1}^{\infty} \Delta u * \gamma^n(t) \tan^{-1}\left(\frac{2Dx}{4n^2 H^2 - D^2 + x^2}\right) \quad (2.22)$$

where $*$ represents convolution in time.

We summarise the stress and strain coefficients in equations 2.13 and 2.14, specific forms of $\hat{\gamma}^n(s)$ and the corresponding inversely transformed forms of $\gamma^n(t)$ for Maxwell and SLS materials in Table 2.2. The material behaviour of a Burger's body follows Equation 2.18 but due to the complexity of the equations and their derivations, we do not include the corresponding $\hat{\gamma}^n(s)$ and $\gamma^n(t)$ in that case. For the complete derivations of Equation 2.20 and its inverse form see Hetland and Hager (2005).

More realistic viscoelastic models

We have previously discussed the simplest viscoelastic solution for a case of a strike-slip fault embedded in an elastic layer overlying a viscoelastic half-space. However, developing analytical forms of viscoelastic solutions is not generally possible unless the effect of heterogeneity within the media, gravity or other complexities associated with more realistic viscoelastic models can be discounted. Accordingly, in order to invert complex forms of viscoelastic solutions in the frequency domain into the time domain, time-consuming numerical methods such as the Haskell propagator algorithm (Haskell,

Figure 2.9: Schematic illustration of a propagator matrix being used to transform information throughout horizontally stratified layers.

1953) and the Hankel transform have been developed.

Semi-analytical algorithms are applied to reduce the computational time, using the collocation technique or full normal mode analysis and were the earliest tools for solving the Laplace transform inversion of viscoelastic solutions (Schapery, 1967; Peltier, 1985). Although the collocation technique is less time-consuming than full normal mode analysis, it requires preliminary information about the relaxation times (see Mitrovica and Peltier, 1992).

Several studies have attempted to derive semi-analytical solutions using the propagator matrix method for the layered viscoelastic representations. The gravitational layered earth models of Rundle (1980), Iwasaki (1985) and Fukahata and Matsu'ura (2006) and spherical geometry models without (Pollitz, 1992) and with considering the effects of gravity by Pollitz (1997) are examples of such semi-analytical solutions.

For layered elastic and viscoelastic models, the propagator matrix method attempts to simplify the boundary-value-problem to linear matrix equations (Gilbert, 1983). If $f(z)$ is a vector of N variables, the transformation from the layer z_0 to z_n is written in the form of propagator matrices M as follows (Figure 2.9):

$$f(z_n) = M(z_n, z_{n-1}) \cdots M(z_2, z_1) M(z_1, z_0) f(z_0) \quad (2.23)$$

The roles of propagator matrices for analysing elastic or viscoelastic responses to dislocation sources in stratified earth models are: 1) To take stress-displacement vectors and propagate them to top or bottom layers; and 2) To assure that traction-free boundary conditions are met (Gilbert, 1983).

For instance, the homogeneous governing equation derived from the equilibrium equation for the stress-displacement vector $v = [d_3, \sigma_{23}]$ is expressed in the general form

$$\frac{dv(z)}{dz} = Av(z) \quad (2.24)$$

where z is the vertical component. Matrix A consists of 4×4 elements as functions of the elastic constants λ and μ and also a wavenumber k . The wavenumber is introduced as the result of applying the Fourier transform to the governing equation in order to reduce the partial differential equation to an ordinary differential equation.

$$A = \begin{bmatrix} 0 & k & 1/\mu & 0 \\ \frac{-k\lambda}{\lambda+2\mu} & 0 & 0 & \frac{1}{\lambda+2\mu} \\ \frac{4\mu k^2(\lambda+\mu)}{\lambda+2\mu} & 0 & 0 & \frac{k\lambda}{\lambda+2\mu} \\ 0 & 0 & -k & 0 \end{bmatrix} \quad (2.25)$$

The propagator matrix $M(z, z_0)$ for propagating from z_0 to z in which $v(z) = M(z, z_0)v(z_0)$ is given by

$$M(z, z_0) = I + \int_{z_0}^z A(\xi_1) d\xi_1 + \int_{z_0}^z A(\xi_1) \left[\int_{z_0}^{\xi_1} A(\xi_2) d\xi_2 \right] d\xi_1 + \cdots, \quad (2.26)$$

Similarly, the inhomogeneous form of an equilibrium equation equivalent to Equation 2.24 is written as

$$\frac{dv(z)}{dz} = Av(z) + h(z) \quad (2.27)$$

where h represents the source term. Using the propagator matrix $M(z, z_0)$, the general solution of the inhomogeneous equilibrium equation is expressed in the form

$$v(z) = M(z, z_0)v(z_0) + \int_{z_0}^z M(z, \xi)h(\xi)d\xi \quad (2.28)$$

(e.g. Aki and Richards, 1980). We do not intend to discuss the equilibrium equations here and their derivations and thus all equations are presented in their general forms. Further details can be found in Segall (2010).

Singh (1970) and Sato (1971) proposed two types of propagator matrices, namely up-going and down-going, respectively. The up-going and down-going propagator matrices are numerically stable below and above the source, respectively. Therefore, the down-going propagator matrix produces stable solutions for the surface displacement and stress field. We must consider that unless unified up-going and down-going propagator matrices are used, the internal displacement is numerically unstable (Fukahata and Matsu'ura, 2005). To address this, Fukahata and Matsu'ura (2005) derived general expressions for the internal displacement field for a multi-layered elastic half-space and Fukahata and Matsu'ura (2006) for the viscoelastic half-space, using the Correspondence Principle. In Chapter 4, we apply the semi-analytical method of Fukahata

Figure 2.10: Dislocation source at a depth d within a multi-layered viscoelastic half-space and viscoelastic slip response at point (r, ϕ, z) (Fukahata and Matsu'ura, 2006)

and Matsu'ura (2006) in order to calculate viscoelastic Green's functions (viscoelastic responses to unit slip). The viscoelastic slip-response function estimated by Fukahata and Matsu'ura (2006) consists of instantaneous elastic and viscoelastic transient terms. The elastic term is expressed by the Heaviside step function $H(t)$ and the transient term by the superposition of different decay time functions (Equation 2.31). We only present equations for the z component of displacement d_z in a cylindrical coordinate system (r, ϕ, z) in a medium with elastic constants μ_l and γ_l (Figure 2.10). The arbitrary point and the dislocation source are located in layer j and k , respectively.

For an elastic multi-layered half-space the vertical component of elastic displacement (d_z^e) is written as

$$d_z^e(r, \phi, z, t, \mu_l, \gamma_l) = H(t) \frac{\Delta u}{4\pi} \int_0^\infty \xi Y_2^e(z, \xi, j, \mu_l, \lambda_l) J(r, \phi, \xi) d\xi \quad (2.29)$$

where J is calculated from Bessel functions (see Watson, 1995) of order 0-2 (details in Fukahata and Matsu'ura 2006). More details are given for calculating Y_k^e by Fukahata and Matsu'ura (2006). The elastic components are denoted by superscript e . For the viscoelastic multi-layered half-space using the Correspondence Principle, the Laplace transform of the z component of displacement is expressed as

$$\tilde{d}_z(r, \phi, z, s) = \frac{\Delta u}{4\pi} \int_0^\infty \xi \tilde{Y}_2(z, s, j) J(r, \phi) d\xi \quad (2.30)$$

After operating the inverse Laplace transform and obtaining Y_k , $\tilde{Y}_k$ is replaced by Equation 2.31 in Equation 2.30 and the viscoelastic displacement components are resolved.

$$Y_k(z, t; j) = H(t) \left[Y_k^e(z, j, \mu_l, \lambda_l) - \sum_{i=1}^M a_i [1 - \exp(b_i t)] \right] \quad (2.31)$$

The transient creep is expressed by superposition of M exponential decaying functions with M values of coefficients a_i and b_i (see Fukahata and Matsu'ura 2006 for more details of calculating the coefficients).

In addition to vertical heterogeneity of viscoelastic properties, the mechanical behaviour of the ductile material within the lower crust and upper mantle vary with lateral variations of viscosity, temperature and elastic thickness. Pollitz (2003a) considered the relaxation of the heterogeneous spherical viscoelastic model and introduced semi-analytical solutions which also account for the coupling of viscoelastic rebound and gravity.

Although the semi-analytical inversion methods were first proposed to reduce computational time, purely numerical methods are no longer impossible tasks. Wang et al. (2006) presented a numerical method using the Riemann-Mellin contour integral and the Fast Fourier Transform (FFT) technique to compute Green's functions in a viscoelastic-gravitational half-space. The use of FEM (e.g. Aagaard et al., 2009, 2013, 2015) to compute the viscoelastic solutions taking into consideration vertical and lateral heterogeneity, coupling with gravity and more complex geometries (e.g. slab geometry for subduction earthquakes) is another example of a fully-numerical viscoelastic methods. In Chapter 4, we consider lateral heterogeneity and the geometry of a slab in a simulated subduction earthquake in our simulations using PyLith (Aagaard et al., 2009, 2013, 2015).

Coupling with gravity

Fault movements perturb gravitational equilibrium (Segall, 2010). Over short temporal and small spatial scales, the perturbation of gravitational forces and potential is negligible. However, for long-term viscoelastic effects, especially after large earthquakes, the effect of the gravity field must be considered as it may change the static displacement by up to 10% (e.g. Fernández et al., 1996, 1999).

Rundle (1982) solved for static displacement in a self-gravitating earth model with gravity g as a constant body force and density ρ . In this case, Equation 2.25 therefore

becomes

$$A = \begin{bmatrix} 0 & k & 1/\mu & 0 \\ \frac{-k\lambda}{\lambda+2\mu} & 0 & 0 & \frac{1}{\lambda+2\mu} \\ \frac{4\mu k^2(\lambda+\mu)}{\lambda+2\mu} & k\rho g & 0 & \frac{k\lambda}{\lambda+2\mu} \\ k\rho g & 0 & -k & 0 \end{bmatrix} \quad (2.32)$$

The eigenvalues of Equation 2.25 without considering the effects of gravity, are $\pm|k|$. With gravity we have eigenvalues $\omega = \pm|k| \left[1 \pm \frac{k_g}{k}\right]^{1/2}$ for matrix A (Equation 2.32) where the gravitational wavenumber is

$$k_g = \frac{\rho g}{\sqrt{\mu(\lambda + 2\mu)}} \quad (2.33)$$

If $k \leq k_g$, the eigenvalues are imaginary and the boundary value problem has no solution. At long time scales, when the effective shear modulus is small, the gravitational wavenumber becomes large enough to make the boundary value problem unstable (Wang, 2005; Wang et al., 2006). Wang (2005) corrected this issue by using the Adams-Williamson condition in the governing equations

$$\frac{d\rho}{dz} = \frac{\rho^2 g}{K} \quad (2.34)$$

where $K = \lambda + 2\mu/3$ is the bulk modulus. The modified governing equations lead to the modified gravitational wavenumber and modified eigenvalues (equations 2.35 and 2.36, respectively):

$$k_g = \frac{\rho g}{\sqrt{\mu(\lambda + 2\mu)}} \implies \tilde{k}_g = \frac{\rho g}{\sqrt{4/3 K(\lambda + 2\mu)}} \quad (2.35)$$

$$\omega = \pm|k| \sqrt{1 \pm \frac{k_g}{k}} \implies \omega = \pm \sqrt{k^2 + 2/3 \left(\tilde{k}_g^2 \pm \tilde{k}_g \sqrt{\tilde{k}_g^2 - k^2} \right)} \quad (2.36)$$

Additionally, using numerical methods such as the FEM, the singularity due to gravity can also be artificially resolved by assuming a finite thickness of the half-space and imposing constraints on the bottom interface of the layer (see Wu, 1992; Zhong et al., 2003; Wu, 2004).

Multiphase postseismic creep and layering

Several researchers have shown that the postseismic signal can reveal single or multiple phases of relaxations. Savage et al. (2003) fit exponential functions to the postseismic relaxation after the 1992 M_w 7.3 Landers and 1999 M_w 7.1 Hector Mine earthquakes, California, of the form $u_{exp} = A[1 - \exp(-\frac{t}{\tau_{exp}})]$, representing a single process with a

relaxation time of τ_{exp} . Herring (2003) analysed the postseismic signal measured by the Southern California Integrated GPS network after the 1994 Northridge earthquake and described the observed displacement with logarithmic functions of the form $u_{ln} = B[1 + \ln(-\frac{t}{\tau_{eln}})]$. Savage et al. (2005) explained postseismic displacements in the epicentral area, after the 1992 M_w 7.3 Landers, 1999 M_w 7.6 Chi-Chi (central Taiwan), 1999 M_w 7.1 Hector Mine, 2002 M_w 7.9 Denali (Alaska), 2003 M_w 6.5 San Simeon and 2004 M_w 6.0 Parkfield earthquakes (California), in terms of logarithmic functions.

Postseismic relaxation with two phases can be well-described by logarithmic functions. Hetland and Hager (2005) examined how layering and different rheologies affect viscoelastic relaxation. In addition to the viscoelastic half-space, they considered a viscoelastic middle layer between the upper elastic layer and the lower half-space in three-layer models. In doing so, Hetland and Hager (2005) showed that a model consisting of an elastic layer overlying a Burger's half-space (elastic-Burger's) imposes two mechanical time scales with logarithmic relaxation. The weaker the Kelvin element in the Burger's body, the larger the difference between the resulting time scales are.

The elastic-Maxwell-Maxwell model causes large-wavelength displacement if the relaxation time associated with the middle layer is long. Both exponential and logarithmic functions can describe the temporal displacement related to the elastic-Maxwell-Maxwell model.

A three-layer model with a Maxwell rheology in the middle layer and a Burger's body in the half-space behaves differently near the fault if the Maxwell element is relatively weak with respect to the half-space. The greater the distance from the fault, the greater the effect the Burger's body has. If the Maxwell middle layer is strong, two phases of relaxation at any distance from the fault are produced and the model approaches the solutions of the elastic-Burger's model. Generally, this three-layer model exhibits up to three time-scales in the postseismic relaxation. Interestingly, swapping the rheologies of the middle layer and the half-space results in four mechanical time scales from the coupling of the two upper layers and the two lower layers. The number of the mechanical time scales increases to up to five by associating a Burger's body to the middle layer and a Burger's Body to the lower half-space. In Chapter 3, we only examine the effects of a bi-viscous Burger's half-space on the multiphase postseismic displacement following the Dusky Sound earthquake.

Literature review of viscoelastic models used in previous studies

Having reviewed different viscoelastic models with a broad range of rheologies and corresponding viscosities, it is beneficial to compare the different assumptions upon which models of postseismic signals have been based. A summary of some of the parameters used in these studies is presented in Figure 2.11. For more details, we classify the information extracted from 88 studies in Tables B.1, B.2 and B.3 (Appendix B). From this literature review, key considerations are highlighted:

Non-uniqueness of postseismic solutions: Rousset et al. (2012a) demonstrated the non-uniqueness of postseismic solutions for the 1999 Chi-Chi earthquake based on a sensitivity test. It was shown that surface postseismic displacement is not sensitive to the rheology of the upper mantle. As Rousset et al. (2012a) explained, understanding the subsurface structure of the earth using geodesy can often present non-unique solutions. Different studies infer different mechanisms of postseismic displacements. For instance, the postseismic displacement observed in the period 2007-2011 after the Kuril earthquake doublet (Russia), can be explained either by a Maxwell rheology with varying viscosity from 2×10^{17} to 10^{18} Pa·s within four years or by a power law creep (Kogan et al., 2013). In a similar way, Rollins et al. (2015) considered several models to explain the postseismic displacement following the the 2010 M_W 7.2 El Mayor-Cucapah earthquake including: 1) Afterslip within the lower crust and assuming a zone with a low viscosity and Newtonian asthenosphere; 2) Assuming a zone with a low viscosity and mantle with a power-law rheology and Newtonian asthenosphere.

Improving the uniqueness of postseismic solutions: Using other observations from sources such as tomography, thermo-kinematics models, seismicity and the local hydro-isostatic adjustment associated with the mantle flow can improve the uniqueness of geodetic problems. Hoechner et al. (2011) demonstrated that the viscoelastic effect of a Maxwell rheology combined with an afterslip model underestimates the postseismic geoid changes observed by the GRACE mission following the M_W 9.1 2004 Sumatra earthquake. However, a Burger's body with a transient viscosity of 10^{18} Pa·s and steady-state viscosity of 10^{19} Pa·s within the viscoelastic layer underlain by the elastic layer, 40 km thick, agrees with the geoid changes. Hoechner et al. (2011) suggested that installing ocean bottom pressure gauges, measuring geoid changes by the GRACE mission, in addition to space-based geodetic observations are useful tools to distinguish between bi-viscous and afterslip as the main postseismic mechanisms. Furthermore, Panet et al. (2010) suggested that, as GPS data are less sensitive to deeper processes, gravity signals observed by the GRACE mission when used as complementary data overcome the small-scale and sparsity of GPS measurements. They

showed how the estimated mantle and lower-mantle viscosity can be improved using large-scale gravity signals. The other advantage of constraining viscoelastic relaxation models with gravity observations is the sensitivity of gravity to the vertical component of displacement (Panet et al., 2010; Hoechner et al., 2011).

Slab geometry of subduction earthquakes Assuming a realistic slab geometry or a simple elastic over viscoelastic plane earth model leads to different postseismic displacements. Masterlark (2003), Hu et al. (2004) and Pollitz et al. (2008) explained how the damping effect of the slab can be mitigated by either reducing the viscosity of the mantle on the continental side, or the thickness of the elastic layer.

Lateral heterogeneity of viscoelastic properties Yamasaki et al. (2014) considered a localised weak zone with an exponentially decreasing viscosity at depth, to explain GPS observations within a six month period following the 1999 İzmit earthquake (North Anatolian Fault Zone). Le Pourhiet et al. (2014) and Yamasaki et al. (2014) associated the presence of low-viscosity zones below the elastic layer with either shear heating, enhanced fluid content, fabric development or grain size reduction. They suggested that the lateral variations of rheology sometimes explain the geodetic observations more efficiently than the stratified layers.

2.2.2 Transient creep

Transient creep causes logarithmic decay of ground displacement following large earthquakes. Benioff (1951) interpreted postseismic relaxation in terms of a mechanism similar to the time-varying creep recovery observed in low-temperature laboratory experiments. Perfettini and Avouac (2004) inferred logarithmic creep decay from geodetic observations at a rate similar to the decay of the cumulative number of aftershocks following the Chi-Chi earthquake. Additionally, Savage et al. (2005) showed that GPS observations following multiple large earthquakes reveal logarithmic decay with time in the epicentral area during the early postseismic stages, which is primarily explained by two rheologies including frictional afterslip and a bi-viscous Burger's body.

Perfettini and Avouac (2004) considered rate-and-state frictional law in order to link the logarithmic decay of aftershocks to geodetic observations. Several authors postulate that the transient creep process is a result of stress-driven afterslip combined with viscoelastic flow (e.g. Johnson et al., 2009; Wang et al., 2009; Panet et al., 2010; Yamagiwa et al., 2015; Rollins et al., 2015). If so, the rapid logarithmic decay at the early stage of postseismic time period can be explained by afterslip being driven by elastic and viscoelastic stress changes induced by earthquakes. Johnson et al. (2009) devel-

Figure 2.11: Different viscosity values and start depth of viscoelastic layers extracted from a review of 88 papers. Red area of the circles correlates with values of viscosity and the size with the start depth of viscoelastic layers. The papers discussed in the current section are shown in the map.

oped a numerical coupled afterslip and viscoelastic relaxation model in order to model postseismic displacement following the Denali, Alaska, earthquake. They assumed that afterslip obeys the rate-strengthening friction law and solved for the numerical integral in Equation 2.37 in order to estimate slip (s) at time t

$$\tau_i(t) = \tau_i^c(t) + \int_0^t G_i(t - t')v(t')dt' \quad (2.37)$$

where $\tau_i^c(t)$ is the shear stress induced by coseismic elastic and viscoelastic responses on each patch of the fault (patch i). G_i is the time-dependent viscoelastic Green's functions (Fukahata and Matsu'ura, 2006), and v is the time-dependent slip-rate.

Barbot and Fialko (2010a) presented a framework in which a unified representation of postseismic mechanisms is considered by defining viscoelastoplastic rheology and different relaxed quantities. These quantities are deviatoric stress for viscoelastic relaxation, shear stress for creep and the trace of stress tensor for poroelastic rebound.

In addition to frictional afterslip, the Burger's rheology for the upper mantle with two transient and steady state viscosities has been shown to account for transient creep following several large earthquakes. Transient rapid decays observed in central Alaska following the Denali earthquake and beneath the Mojave Desert following the Hector Mine earthquake were the first such examples of using the Burger's rheology (Pollitz, 2005, 2003b). The bi-viscous rheology has since been used in several other studies to model the early rapid decay in the postseismic period (e.g. Árnadóttir et al., 2005; Pollitz et al., 2006b; Johnson et al., 2007; Pollitz et al., 2008; Hearn et al., 2009; Panet et al., 2010; Hoechner et al., 2011; Sun and Wang, 2015).

Short-term rapid decay of postseismic displacement is often explained in terms of very low values of transient viscosity ($< 10^{17}$ Pa·s) and short relaxation times (e.g. Palano et al., 2009). Very short relaxation times are also consistent with the characteristics timescales of afterslip (a few months to a few years; Pritchard and Simons, 2006; Wang et al., 2012). Therefore, in Chapters 3 and 4, we interpret the logarithmic transient displacement following the Dusky Sound earthquake in terms of afterslip and viscoelastic responses to both coseismic stress changes and afterslip mechanism.

Fault frictional afterslip

A rate-strengthening rheology is commonly assumed to be the dominant type of friction during the postseismic time period inducing afterslip (Lapusta et al., 2000; Rice et al., 2001; Barbot et al., 2009). This form is derived from the general rate-and-state-

dependent friction law of Ruina (1983) and Dieterich (1994)

$$f = f_0 + a \ln\left(\frac{v}{v_0}\right) + b \ln \theta$$

$$f \equiv f_0 + a \sinh^{-1}\left(\frac{v}{2v_0} \theta^{\frac{b}{a}}\right) \quad (2.38)$$

where f_0 is the coefficient of friction at a reference time and reference slip rate v_0 . θ is a state variable, v is slip rate and a and b are frictional parameters. If $\frac{b}{a} > 1$, the work-hardening parameter b/a is rate-weakening, whereas if $\frac{b}{a} < 1$, it is rate-strengthening.

Assuming, $\tau = f\sigma'$, where τ is shear stress in the direction of sliding on a fault and σ' is the effective normal stress, the following extra assumptions have been made in order to derive the rate-strengthening form of the friction law (Lapusta et al., 2000; Rice et al., 2001; Barbot et al., 2009):

- Normal stress changes are negligible with respect to lithostatic and tectonic stress components;
- There are no weakening processes;
- Friction is steady-state; and
- Coulomb stress changes induced by earthquakes are the dominant stress changes inducing further afterslip.

Accordingly, the slip rate induced by coseismic stress changes is derived from Equation 2.38 in the form of Equation 2.39

$$v = 2v_0 \sinh^{-1} \left[\frac{\Delta\tau}{(a-b)\sigma} \right] \quad (2.39)$$

where $\Delta\tau$ is shear stress changes and $(a-b)\sigma$ is the parameter of the model associated with the background normal stress and the frictional parameter $(a-b)$; e.g. Lapusta et al., 2000; Rice et al., 2001; Barbot et al., 2009).

Assuming the component of earthquake-induced traction changes (in the direction of sliding on the fault surface) drives afterslip, the forward relation between the slip rate and the traction changes is defined by Equation 2.39. Alternatively, afterslip can be estimated using conventional methods by inverting the observed surface displacement (e.g. Freed, 2007; Kositsky and Avouac, 2010; Evans and Meade, 2012). Using forward modelling methods, we need to know the coseismic slip distribution, the geometry of the afterslip surface, the sliding direction or rake angle and the appropriate frictional properties. In contrast with the forward method, inverting the surface displacement data does not consider the traction changes on the fault surface induced by the earthquake; Slip occurs either within or beyond the coseismic rupture patches. Moreover, in Equation 2.39, if $(a-b)\sigma \ll \Delta\tau$, the rate-strengthening afterslip model possesses a singularity. In addition, the general form of the rate-and-state dependent

friction law requires knowledge of several properties such as state variable θ and frictional parameters a and b .

In Chapter 5, we employ forward model of frictional afterslip from Equation 2.39 as a first-order approximation in order to understand the main postseismic mechanisms following the Canterbury earthquake sequence of 2010–2012. In Chapter 3 and 4, we discuss time-dependent inversion methods with which to estimate afterslip following the Dusky Sound earthquake in an elastic-viscoelastic medium.

Time-dependent inversion of geodetic data

Daily measurements of continuous GPS (cGPS) sites provide high-resolution crustal information with a horizontal accuracy of a few millimetres and poorer vertical accuracy (approximately three times larger, Heflin et al., 1992; King et al., 1995). Time-dependent inversion of temporally dense observations measured from cGPS data enables estimation of transient slip associated with postseismic displacements. Therefore, providing a framework to invert temporal geodetic observations is of great use. Segall and Matthews (1997) developed the Network Inversion Filter (NIF) method to approximate slip using spatial basis functions $B_k(\xi)$ and temporal coefficients $c_k(t)$ of the form

$$s(\xi, t) = \sum_{k=1}^M c_k(t) B_k(\xi) \quad (2.40)$$

Slip through time is predicted in this method from the sum of steady-state slip and random walks. Segall and Matthews (1997) computed Kalman gain (see Anderson and Moore 2012) to estimate the update for each parameter and associated covariance matrix. The temporal smoothness of slip was controlled by a constant scaling parameter. Fukuda et al. (2004) subsequently modified the NIF method by applying a Monte Carlo Mixture Kalman Filter (MCMKF) in which the temporal smoothness varies as a stochastic variable. Consequently, if slip changes abruptly, MCMKF is capable of capturing the slip event. However, Fukuda et al. (2004) assumed spatially uniform fault slip in their method, a limitation later addressed by Fukuda et al. (2008) with combining a hierarchical Bayesian approach (MacKay, 1992) and state-space Kalman filtering (Kalman, 1960; Segall and Matthews, 1997; Kitagawa and Gersch, 2012). The effect of viscoelastic relaxation is not taken into account in these methods.

Estimating time-dependent slip distribution was also considered by Fukahata et al. (2004) using Akaike’s Bayesian Information Criterion (ABIC) method (Akaike, 1977, 1980; Yabuki and Matsu’Ura, 1992; Ide et al., 1996). The advantage of this approach

Figure 2.12: Schematic diagrams for a) Spatial basis b) Temporal basis functions used to model slip rates throughout ABIC method.

is that they could consider the effects of viscoelastic relaxation in the mantle. Slip rate is thus written as the superposition of temporal $T_l(t)$ and spatial basis functions $X_k(\xi)$ (Figure 2.12)

$$v(\xi, t) = \sum_{k=1}^K \sum_{l=1}^L a_{kl} X_k(\xi) T_l(t) \quad (2.41)$$

where a_{kl} are coefficients determined from the inversion of geodetic observations using the ABIC method. The Bayesian framework used in the ABIC method incorporates the smoothness of the fault slip in time and space as a priori information by introducing two hyperparameters, α^2 and β^2 . Coefficients a_{kl} and the hyperparameters are optimised in order to minimise the misfit with geodetic observations as well as the spatial and temporal roughnesses of fault slip.

In all the above methods, inversion of time-dependent geodetic observations is performed by the expansion of slip in terms of spatial and temporal basis functions. Kositsky and Avouac (2010) inverted geodetic time series with the inverse of elastic Green's functions multiplied by Singular Value Decomposition (SVD) of ground observations D . This method does not require use of a temporal hyperparameter. Temporal information is directly obtained from decomposition of temporally dense geodetic observations

into spatial (U) and temporal functions (V), via

$$D_r = U_r \Lambda_r V_r' \quad (2.42)$$

$$\mathbf{S} = G_e^{-1} U_r (\Lambda_r V_r') \quad (2.43)$$

where $\mathbf{S}$ is time-dependent slip matrix and G_e is the matrix of elastic Green's functions, D_r is a filtered form of the full displacement matrix D with n observed components at T epochs, of the dimensions $n \times T$; U_r is of dimension $n \times r$, representing the first r columns of U ; Λ_r is of dimension $r \times r$, formed from the first r rows and columns of Λ , which contains singular values of D ; and V_r is of dimension $T \times r$ and represents the first r columns of V . r is the number of retained singular values and the optimal value of r is selected by minimising the misfit between the displacement signal D and the filtered signal D_r (Kositsky and Avouac, 2010). If the single values are sorted from higher to lower values, we can perform filtering by single value truncation. Throughout this thesis, matrix transpose is denoted by prime ($'$) symbol.

Viscoelastic responses in the lower crust/upper mantle using the Principal Component Analysis-based Inversion Method (PCAIM) are not considered here. In Chapter 4, we propose an approach using semi-analytical (Fukahata and Matsu'ura, 2006) or hybrid numerical viscoelastic Green's functions (Aagaard et al., 2009, 2013, 2015) in order to invert the geodetic time series assuming an elastic-viscoelastic medium.

2.2.3 Viscoelastic cycle models

The ultimate aim of Chapter 4 of this thesis is to provide an inversion framework in which to constrain postseismic displacement mechanisms in an elastic-viscoelastic medium. This framework is useful to model long-term ground displacements measured from geodetic observations within a viscoelastic cycle. Studying a viscoelastic cycle encompassing repeated earthquakes, with corresponding postseismic mechanisms, and the longer-term interseismic period is a subject of ongoing research. The materials described in this section are not directly used in the following chapters, although different rheological models here compliment the assumptions in Chapters 3 and 4.

The concept of viscoelastic cycle models of periodic earthquakes is that the asthenosphere relaxes the stress imposed by the periodic earthquakes on the upper part of the fault, and aseismic creep on the lower segment. Postseismic relaxation within the asthenosphere thereafter transfers stress to the elastic layer. The idealised earthquake cycle with periodic earthquakes on the upper part of the fault has been assumed for viscoelastic cycle models in several studies under different rheological assumptions.

Perfettini and Avouac (2004) proposed the Brittle Creeping Fault Zone (BCFZ) rheology for the region connecting the seismogenic zone (at temperatures $< 250^\circ\text{C}$) to the region with temperatures $> 400^\circ\text{C}$ (Figure. 2.13a). The rheology for the BCFZ is assumed to be stable rate-strengthening, which does not produce earthquakes. During the interseismic period, the upper part of the fault is locked. The stress drop induced by the earthquake is transferred to the creeping part of the fault below the coseismic rupture zone, relaxing the stress produced by the upper part of the fault. Perfettini and Avouac (2004) assumed that microseismicity over the region below the locked fault is triggered by interseismic strain accumulation. However, the strain released by microseismicity is a fraction of the measured strain by the geodetic data (Perfettini and Avouac, 2004). Thus, rate-strengthening creep is suggested to be the primary mechanism explaining the geodetic observations.

An alternative mechanism is that, grain size reduction after repeated earthquakes localises the ductile shear zone displacement down-dip of the seismogenic zone (Montési and Hirth, 2003). Ductile transient creep is associated with power-law or Newtonian rheologies. For a ductile shear zone of thickness h , the slip rate depends on the power n of the stress. For a Newtonian fluid with a linear rheology ($n = 1$) and a pre-exponential factor A

$$v = Ah\sigma^n \quad (2.44)$$

Equilibrium requires that

$$\frac{d\sigma}{dt} = K(v - v_p) \quad (2.45)$$

A spring-slider system with tectonic- and earthquake-induced stresses is assumed, represented by a loading point velocity v and a shear zone velocity v_p , respectively (Figure 2.14). K is the stiffness of the spring and represents stress sources and shear zone geometry.

Assuming either rate-strengthening friction (Equation 2.39) or a power-law rheology (Equation 2.44) yields the differential equation (Equation 2.45) to be solved and thus an estimation of the fault slip history. Fault slip due to the rate-strengthening rheology evolves in time in the form of $\ln(1 + \frac{t}{\tau_R})$ with τ_R a material constant. For the power-law creep and a linear rheology, evolution of fault slip has the form $n(1 - [1 + (1 - \frac{1}{n})\frac{t}{\tau_R}]^{\frac{1}{1-n}})$ and $(1 - \exp(-\frac{t}{\tau_R}))$, respectively (Montési, 2004).

Savage (2007) developed an analytical model of postseismic slip on the BCFZ with the rheology of Lomnitz solids. Based on the experiments of Lomnitz 1956, materials having the transient creep response are referred to as Lomnitz solids (Savage, 2007). In contrast with the model of Perfettini and Avouac (2004), the BCFZ extends to the

Figure 2.13: a) The BCFZ rheology and triggered microseismicity during the interseismic period, BCFZ is located in the region with temperatures between 250 and 400 below the coseismic rupture patch. b) BCFZ rheology extending to the viscoelastic subsurface. c) Viscoelastic cycle model with constant creep on the creeping part of the fault d) Viscoelastic cycle model over a zero shear stress zone. e) Viscoelastic cycle model over a viscous shear zone.

Figure 2.14: Spring-slider system

viscoelastic subsurface (Figure. 2.13b). The constitutive law for the Lomnitz solids is

$$\sigma = \frac{\mu}{(1 + q \ln(1 + \frac{t}{\tau_R}))} \epsilon \quad (2.46)$$

where σ is the imposed stress, ϵ is the strain response, μ , q and τ_R are material constants.

Figure. 2.13.(c-e) are illustrations for three different analytical cycle models that Johnson and Segall (2004) proposed for modelling creep along the San Andreas Fault System. All three viscoelastic cycle models consist of a vertical fault within an elastic layer, overlying a Maxwell viscoelastic half-space under the following assumptions: 1) constant creep on the lower part of the creeping fault (Figure 2.13c); 2) creep within a zero shear stress zone (Figure 2.13d); and 3) creep within a linear viscous shear zone (Figure 2.13f).

2.2.4 Poroelastic rebound of a porous medium

Understanding the role of fluids at different times and positions in an earthquake cycle is also important in order to interpret postseismic displacement. An instantaneous change of the confining stress due to an earthquake causes pore pressure changes and the formation of dilational and compressional zones (e.g. Peltzer et al., 1996, 1998). Later in the postseismic period, the fluid flows between these zones in order to equilibrate the pore pressure changes. This postseismic process is known as poroelastic rebound (e.g. Roeloffs, 1996; Peltzer et al., 1996, 1998).

In this section, we discuss the following governing equations of poroelastic rebound

$$(\lambda + 2\mu) \nabla (\nabla \cdot \vec{d}) - \mu \nabla \times (\nabla \times \vec{d}) - \alpha \nabla P_p = \vec{f} \quad (2.47)$$

$$Q^{-1} \frac{\partial P_p}{\partial t} - \chi \nabla^2 P_p + \alpha \frac{\partial}{\partial t} \nabla \cdot \vec{d} = 0 \quad (2.48)$$

where λ and μ are the Lamé parameters, d is displacement signal induced by poroelastic rebound, P_p is pore pressure, χ and Q^{-1} are Darcy conductivity and bulk compressibility, respectively, and f is a body force. In order to derive these governing equations, we require several basic definitions.

For a cube of porous rock, an instantaneous change in confining stress yields a decrease of the mean normal stress by an amount $\Delta\sigma_{kk}/3$. Consequently, a cube with undrained bulk modulus K_u compacts isotropically by an amount $\Delta\sigma_{kk}/3K_u$. As a result, pore pressure increases by an amount $-B\delta\sigma_{kk}/3$, where B is Skempton's coefficient (e.g. Biot, 1941; Rice and Cleary, 1976; Roeloffs, 1988; Wang, 2000). The coseismic and subsequent postseismic pressure changes are often referred to as undrained and drained states with values of Poisson's ratio ν_u and ν , respectively (e.g. Peltzer et al., 1998; Jonsson et al., 2003; Freed et al., 2006). In an undrained state (the coseismic period), forces are applied rapidly and the process is not in equilibrium. In contrast, if the porous solid is deformed slowly (during the postseismic period) fluids can flow and maintain equilibrium (Segall, 2010).

The constitutive behaviour of a porous elastic rock saturated with fluid was first described in detail by Biot (1962). Terzaghi (1943) had previously shown that for the first time that pore pressure (P_p) has impact on the normal components of the stress tensor ($S_{ij}; i = j$) but not on the shear components ($S_{ij}; i \neq j$) according to equation

$$\sigma'_{ij} = S_{ij} - \delta_{ij} P_p \quad (2.49)$$

where σ'_{ij} is the effective normal stress (the difference between applied stress and pore pressure) and δ_{ij} is the Kronecker delta function. Nur and Byerlee (1971) modified the Terzaghi equation to obtain the exact effective stress law

$$\sigma'_{ij} = S_{ij} - \delta_{ij} \alpha P_p \quad (2.50)$$

where α is Biot coefficient

$$\alpha = 1 - \frac{K_b}{K_g} \quad (2.51)$$

and K_b and K_g are the rock bulk modulus and the individual solid grains' bulk modulus.

For an elastic solid (with Lamé parameters λ and μ) and neglecting poroelasticity, the

Figure 2.15: Fluid flow through porous medium formulated by Darcy's law.

stress and strain components are linearly proportional, such that

$$\sigma_{ij} = \lambda \epsilon_{kk} \delta_{ij} + 2\mu \epsilon_{ij} \quad (2.52)$$

when poroelasticity is considered, Equation 2.52 is modified to

$$\sigma_{ij} = \lambda \epsilon_{kk} \delta_{ij} + 2\mu \epsilon_{ij} - \alpha \delta_{ij} P_0 \quad (2.53)$$

where

$$P_0 = \frac{1 - 2\nu}{1 + \nu} P_p \quad (2.54)$$

and ν is Poisson's ratio (Nur and Byerlee, 1971; Zoback and Byerlee, 1975; Walls and Nur, 1979).

The strain-displacement relation and the equilibrium equations, respectively, are defined as follows

$$\epsilon_{ij} = \frac{1}{2} \left(\frac{\partial d_i}{\partial x_j} + \frac{\partial d_j}{\partial x_i} \right) \quad (2.55)$$

$$\frac{\partial \sigma_{ij}}{\partial x_j} = f_i \quad (2.56)$$

where f_i and d_i are the i^{th} components of body force and displacement at position x .

Based on Darcy's law, the discharge rate Q is proportional to the pressure drop between two points, the permeability k and surface area A , and is inversely proportional to the distance L between two points and the viscosity η (Figure 2.15).

Accordingly, the volumetric flux density q_i in the i^{th} direction through a porous medium with permeability tensor k_{ij} and density ρ_0 is obtained from Darcy's law

$$q_i = -\rho_0 \frac{k_{ij}}{\eta} \left(\frac{\partial P_p}{\partial x_j} - \rho_0 g \delta_{3j} \right) \quad (2.57)$$

where g is gravity, η viscosity and x_j the j^{th} component of position (see Whitaker, 1986; Segall, 2010).

Conservation of fluid mass (m) requires that

$$\frac{\partial q_i}{\partial x_j} = \frac{\partial m}{\partial t} \quad (2.58)$$

The strain-displacement relation (Equation 2.55), equilibrium conditions (Equation 2.56), Darcy's law (Equation 2.57), and conservation of fluid mass relation (Equation 2.58) are used along with the constitutive behaviour of a porous elastic solid (Equation 2.53) to derive the governing equations (for further details see Segall 2010 and Wang and Kämpel 2003).

$$\mu \nabla^2 d_i + \frac{\mu}{1-2\nu} \frac{\partial^2 d_j}{\partial x_i \partial x_j} - \alpha \frac{\partial P_p}{\partial x_j} = f_i \quad (2.59)$$

$$Q^{-1} \frac{\partial P_p}{\partial t} - \chi \nabla^2 P_p + \alpha \frac{\partial}{\partial t} \nabla \cdot d = 0 \quad (2.60)$$

where bulk compressibility and Darcy conductivity are expressed as

$$Q^{-1} = \frac{9}{2} \frac{(1-2\nu_u)(\nu_u - \nu)}{(1-2\nu)(1+\nu_u)^2 \mu B^2} \quad (2.61)$$

$$\chi = \frac{9}{2} \frac{(1-\nu_u)(\nu_u - \nu) D}{(1-\nu)(1+\nu_u)^2 \mu B^2} \quad (2.62)$$

Here D is hydraulic diffusivity and ν_u and ν are the undrained and drained Poisson's ratios, respectively. The governing equations are expressed in vector form using the curl of the curl identity

$$\nabla \times (\nabla \times d) = \nabla (\nabla \cdot d) - \nabla^2 d \quad (2.63)$$

Thus

$$(\lambda + 2\mu) \nabla (\nabla \cdot \vec{d}) - \mu \nabla \times (\nabla \times \vec{d}) - \alpha \nabla P_p = \vec{f} \quad (2.64)$$

$$Q^{-1} \frac{\partial P_p}{\partial t} - \chi \nabla^2 P_p + \alpha \frac{\partial}{\partial t} \nabla \cdot \vec{d} = 0 \quad (2.65)$$

where $\lambda = \frac{2\mu\nu}{1-2\nu}$.

Poroelasticity's solutions

Wang and Kämpel (2003) suggested a method for numerical computation of the governing equations 2.64 and 2.65 in a multi-layered poroelastic medium, using the Laplace-Hankel transform of the solution vectors (Equation 2.66). For displacements of interior interfaces, the pore pressure and traction on both sides of the layer are equal. The top and bottom of the domain are assumed to be free surfaces with zero pressure changes and traction; the boundary conditions are satisfied in the numerical solutions. Using

cylindrical surface harmonics, the governing equations are transformed to a system of ordinary differential equations. The Hankel transform for displacement is expressed in terms of cylindrical surface harmonics by

$$d(z, r, \theta, s) = \sum_m \int_0^\infty [U_m(z, s)Z_k^m(r, \theta) + V_m(z, s)R_k^m(r, \theta) + W_m(z, s)T_k^m(r, \theta)]kdk \quad (2.66)$$

where the cylindrical surface harmonics are

$$Z_k^m(r, \theta) = e_z Y_k^m(r, \theta) \quad (2.67)$$

$$R_k^m(r, \theta) = e_r \frac{1}{k} \frac{\partial}{\partial r} Y_k^m(r, \theta) + e_\theta \frac{1}{kr} \frac{\partial}{\partial \theta} Y_k^m(r, \theta) \quad (2.68)$$

$$T_k^m(r, \theta) = e_r \frac{1}{kr} \frac{\partial}{\partial \theta} Y_k^m(r, \theta) - e_\theta \frac{1}{k} \frac{\partial}{\partial r} Y_k^m(r, \theta) \quad (2.69)$$

Here Y_k^m represents a Bessel functions of degree m multiplied by a trigonometric vector (Equation 2.70), (z, r, θ) are the components of the cylindrical coordinate system, s is the Laplace parameter, U_m , V_m and W_m are the coefficients in approximating the solution vectors.

$$Y_k^m(r, \theta) = J_m(kr) \begin{pmatrix} \cos m\theta \\ \sin m\theta \end{pmatrix} \quad (2.70)$$

Using cylindrical harmonics, the problem of poroelasticity is simplified to optimising the coefficients U_m , V_m and W_m . For more details see Wang and Kmpel (2003).

Literature review of poroelastic rebound models

A gradual change in Poisson's ratio from undrained (coseismic) to drained (postseismic) values is the conventional way of explaining observed poroelastic rebound following earthquakes (e.g. Peltzer et al., 1998; Jonsson et al., 2003; Freed et al., 2006). The relations between undrained ν_u and drained ν values for Poisson's ratio are written as (Wang, 2000)

$$\nu = \frac{3\nu_u - \alpha B(1 + \nu_u)}{3 - 2\alpha B(1 + \nu_u)} \quad (2.71)$$

$$\nu_u = \frac{3\nu + \alpha B(1 - 2\nu)}{3 - \alpha B(1 - 2\nu)} \quad (2.72)$$

where α and B are Biot and Skempton's coefficients.

Peltzer et al. (1998) used values of 0.31 and 0.27 for the undrained and drained Poisson's ratios to explain displacement observed across the Emerson Fault in a combination of afterslip and poroelastic rebound following the Landers earthquake. They showed that the poroelastic rebound effects with higher undrained Poisson's ratio of 0.35 account

for the observed uplift signal in Homestead Valley.

Jonsson et al. (2003) observed subsidence within coseismic compressional zones and uplift within dilational zones correlated with water level changes. The signal associated with poroelastic rebound vanished within two months of the June 2000 sequence in the South Iceland Seismic Zone (SISZ). Freed et al. (2006) assessed the contribution of poroelastic rebound in the postseismic signal measured after the Denali, Alaska earthquake with undrained and drained values of 0.26 and 0.22. They found largest contribution of poroelastic rebound near the Denali/Totschunda junction.

Assuming Poisson's ratio to be an isotopic material property is inappropriate in some cases, particularly along subduction zones due to the contrast between stiff oceanic crust and compliant continental crust (Masterlark, 2003; Eberhart-Phillips and Stuart, 1992; Wald and Graves, 2001; Masterlark et al., 2001). Masterlark (2003) explained the delayed triggering effect from the M_w 8.0 1995 Jalisco-Colima Mexico subduction earthquake on a nearby M_w 6.3 rupture 63 days later due to a transient pore fluid pressure recovery. In their study, a 3D Finite Element Model was used to simulate the poroelastic postseismic displacement (see Section 2.2.6). They assigned different values of permeability to the oceanic and the continental crusts in order to account for the lag between the two events. In a similar study, Hughes et al. (2010) investigated stress-coupling between the M_w 9.2 Sumatra-Andaman and the M_w 8.7 Nias earthquakes in terms of transient fluid flow. Their model produced separate regimes of poroelastic rebound in the shallow forearc and volcanic arc and in the oceanic slab. The former dissipated within a month after the event. The migration of fluids within the oceanic crust led to the recovery of pore pressure changes three months after the main event which was coincident with the Nias earthquake.

Masterlark and Wang (2002) considered transient coupling between the Landers and Hector Mine earthquakes to be caused by upper-crust poroelastic rebound with a diffusivity of $0.01 \text{ m}^2/\text{s}$ and lower-crust viscous flow with a viscosity of $5 \times 10^{18} \text{ Pa}\cdot\text{s}$. Diffusivity D controls the time scale of pore pressure recovery due to the migration of fluids and varies between 10^{-11} to $10^5 \text{ m}^2/\text{s}$ (Roeloffs, 1996). Hydraulic diffusivity D is expressed as (Rice and Cleary, 1976)

$$D = \frac{2k}{9\eta} B^2 \mu \frac{(1-\nu)(1+\nu_u)^2}{(1-\nu_u)(\nu_u-\nu)} \quad (2.73)$$

where η and k are viscosity and permeability. All other variables are as previously defined.

In Chapter 5, to model postseismic displacement following the Darfield earthquake (a non-subduction event), we neglect any lateral heterogeneity of poroelastic properties.

We assume poroelastic properties only vary vertically within the continental crust. We use the multi-layered poroelastic rebound method appropriate for non-subduction earthquakes implemented by Wang and Kümpel (2003). This method enables us to consider vertical heterogeneity in a multi-layered poroelastic medium. Fully time-dependent solutions can be estimated by this method which is another advantage over conventional methods that address only two end-members (undrained and drained). In contrast with the FEM, which approximates solutions with shape functions (see Section 2.2.6), poroelastic solutions are approximated by cylindrical surface harmonics (Wang and Kümpel, 2003).

Example models of poroelasticity

We present below the results of simulating poroelastic rebound using the Fortran codes implemented by Wang and Kümpel (2003). The ground displacement is computed in response to right-lateral slip of 5 m on a 90° east striking vertical fault, with both length and width of 10 km. We use P- and S-wave velocities from the national New Zealand velocity model in the Canterbury region (Eberhart-Phillips et al., 2010). Two values for hydraulic diffusivity are considered: $1 \text{ m}^2/\text{s}$ above depth of 1.75 km and $0.001 \text{ m}^2/\text{s}$ below depth of 1.75 km. We set Skempton's and Biot coefficients to values of 0.5 and 0.6, respectively. The cumulative east, north and vertical components of coseismic and postseismic displacement within 365 days of the simulated rupture are shown in Figure 2.16a and b, respectively. The flow of fluids induces subsidence and uplift (Figure 2.16b) in previously (during the coseismic period) compressional (+ sign Figure 2.16a) and dilational zones (- sign Figure 2.16a), respectively.

We also simulate poroelastic rebound within a step-over between two vertical dextral strike-slip faults with lengths and widths of 20 km and 10 km, respectively, striking east to west and slipping by 2 m. In both simulations, the vertical components of displacement due to coseismic and postseismic signals are similar in pattern and different in sign. In Figure 2.17c, we present the time series of poroelastic displacement for a point shown within the step-over. This shows that with the assumed poroelastic parameters, the east and vertical components for the point decay rapidly within approximately 60 and 90 days of the rupture, respectively. In Chapter 5, we apply a similar calculation to estimate the poroelastic rebound effect on the postseismic ground displacement following the Darfield earthquake .

Figure 2.16: Simulated east, north and vertical components of a) Coseismic and b) Accumulated postseismic displacement within 365 days of the simulated rupture due right-lateral slip of 5 m on a vertical fault with the length and width of 10 km. Red colours indicate movement away and blue colours indicate movement towards the direction. Dilational and compressional zones are illustrated by minus and positive signs, respectively.

Figure 2.17: Simulated east, north and vertical components of a) Coseismic and b) Accumulated postseismic displacement within 365 days of the simulated rupture on a step over between two right-lateral faults with lengths and widths of 20 and 10 km, respectively, and slip of 2m. Red colours indicate movement away and blue colours indicate movement towards the direction. Dilational and compressional zones are illustrated by minus and positive signs, respectively. c) Time series of east (blue) and vertical (red) components of displacement for every 30 days calculated from simulated poroelastic rebound.

2.2.5 Shallow fault zone dilatancy recovery

Static and dynamic stress changes after earthquakes cause the process of dilatancy during the coseismic period and the consequent post-recovery (Nur, 1971; Scholz, 1974; Vidale and Li, 2003). Dilatancy and its recovery alter the velocity of seismic waves

after an earthquake and during postseismic period when the fault zone contracts (Vidale and Li, 2003; Schaff and Beroza, 2004; Peng and Ben-Zion, 2006; Reyners et al., 2014). Fialko et al. (2005) observed significant shallow slip deficits associated with the 2003 Bam earthquake in Iran. Subsequently, Fielding et al. (2009) explained this phenomenon to be the result of very shallow fault zone dilatancy recovery. Postseismic displacement due to viscoelastic relaxation in the lower crust/upper mantle, afterslip and poroelastic rebound in the upper crust often dominates the displacement field from dilatancy recovery. For the first time, Fielding et al. (2009) geodetically observed the subsidence signal corresponding to the process of dilatancy recovery 3.5 years after the 2003 Bam earthquake in Iran. To model the subsidence signal due to the contracting fault zone, an array of Mogi point sources are assumed on a fault surface (Figure 2.18). This method simultaneously allows the fault surface to slip and contract. Note that, Fielding et al. (2009) believes more continuous contracting volume source is required to explain the physics of dilatancy recovery. For the purpose of efficiency and simplicity, we use the same method in order to understand the process of dilatancy recovery within the top 4km of the Darfield faults.

The ground displacement resulting from a combination of slip and volumetric displacement on a buried fault is linearly related to the slip components (s_s and s_d) and volume change (Δv). This means that we can express the observation equation relating the observed ground displacement (d) and the parameters of interest (s) as follows

$$d = Gs + \epsilon = \begin{bmatrix} G_f & G_v \end{bmatrix} \begin{bmatrix} s_s \\ s_d \\ \Delta v \end{bmatrix} + \epsilon \quad (2.74)$$

where G is the combined design matrix, G_f and G_v are functions of dislocation model and Mogi point source parameters (Okada, 1985; Mogi, 1958), and ϵ is the residual.

If X , Y and D are the east, north coordinates and depth of the point source with respect to the ground location and $R = \sqrt{x^2 + y^2 + d^2}$, then three-dimensional ground displacement induced by a Mogi point source is given by

$$d_v = \frac{\nu - 1}{\pi R^3} \begin{bmatrix} X \\ Y \\ D \end{bmatrix} \Delta v + \epsilon_v = G_v \Delta v + \epsilon_v \quad (2.75)$$

where ν is Poisson's ratio. In order to solve two components of afterslip on fault planes as well as the volume change, we use the linear least squares approach to invert the displacement signal. Inversion of Equation 2.74 is still an ill-posed problem and

requires spatial smoothing. During the inversion, the following function is minimised

$$f = \| d - Gs \|^2 + \gamma^{-2} \| \nabla^2 s \|^2 \quad (2.76)$$

∇^2 is the approximation of the Laplacian operator. γ determines the trade-off between misfits and smoothness of parameters. A value between spatially smooth and rough extremes is chosen (e.g. Beavan et al., 2012a) for hyperparameter γ , which minimises the misfit with observations and the smoothness.

Figure 2.18: An array of regularly spaced contracting Mogi point sources in the shallow fault zone along with afterslip mechanism after the 200 Bam earthquake in Iran (Figure 4; Fielding et al., 2009).

2.2.6 Numerical computation using the Finite Element Method (FEM)

The FEM is applied for modelling a structure by using finite number of interconnected elements (see Strang and Fix, 1973; Brenner and Carstensen, 2004). For complex structures, analytical solutions for a Boundary Value Problem (BVP) often do not exist.

As the name suggests, FEM begins with discretising the continuum into finite elements. For three-dimensional (3D) problems, tetrahedral and hexahedral (or brick) elements are widely used. Solving the BVPs using the FEM consists of: 1) forming the weak statement of the BVP; 2) approximating the weak form with the use of finite element functions (Hughes, 2012). A weak form is an integral function which is simpler to solve than the strong form of Partial Differential Equations (PDEs).

Using the FEM, element equations are converted to their corresponding matrix forms using variational or weighted residual methods (see Reddy, 1989, 1993). One of the

primary sub-groups of the weighted residual methods is the Galerkin method by which weak forms of the element equations are generated (see Nayroles et al., 1992; Belytschko et al., 1994; Lu et al., 1994).

For a function f , the general weak form over the domain Ω is written as follows

$$\int_{\Omega} w f d\Omega = 0 \quad (2.77)$$

where w is a weight function. The integral equations are solved using integration by parts. We approximate dependent variables (e.g. u at position x and time t) by their corresponding shape functions as expressed by

$$u(x, t) = \sum_{i=1}^n \phi_i(x) u^i(t) = \Phi^T U \quad (2.78)$$

where $\phi_i(x)$ are shape functions. The weighted residual formulation is applicable to a wide variety of BVPs. In the Galerkin method, we use the same weights for the dependent variables and the weak forms ($w \equiv \phi$). There are other methods classified under weighted residual formulation such as techniques of collocation, sub-domain and least squares (e.g. Ackroyd, 1986; Reddy, 1989), but we do not discuss such techniques here.

To convert the element functions using the weak forms, we use the form below

$$\begin{bmatrix} f_1 \\ f_2 \\ \vdots \\ f_n \end{bmatrix} = \begin{bmatrix} K_{11} & k_{12} & \cdots & k_{1n} \\ K_{21} & k_{22} & \cdots & k_{2n} \\ \vdots & & & \vdots \\ K_{n1} & k_{n2} & \cdots & k_{nn} \end{bmatrix} * \begin{bmatrix} u_1 \\ u_2 \\ \vdots \\ u_n \end{bmatrix} \quad (2.79)$$

$$\{f\} = \{K\}\{u\}$$

To illustrate this, we discuss the steps required for a simple heat transfer BVP by Equation 2.80

$$\begin{aligned} f : \frac{\partial}{\partial x}(k_x \frac{\partial T}{\partial x}) + \frac{\partial}{\partial y}(k_y \frac{\partial T}{\partial y}) + \frac{\partial}{\partial z}(k_z \frac{\partial T}{\partial z}) - \frac{\partial T}{\partial t} &= 0; G = -\frac{\partial T}{\partial t} \\ T = T_b : S1 \\ (k_x \frac{\partial T}{\partial x})n_1 + (k_y \frac{\partial T}{\partial y})n_2 + (k_z \frac{\partial T}{\partial z})n_3 + q &= 0 : S2 \\ (k_x \frac{\partial T}{\partial x})n_1 + (k_y \frac{\partial T}{\partial y})n_2 + (k_z \frac{\partial T}{\partial z})n_3 + h(T - T_a) &= 0 : S3 \end{aligned} \quad (2.80)$$

Here T is the unknown temperature, q heat flux, h heat transfer coefficient, n surface normal and k the thermal conductivity. $S1$, $S2$ and $S3$ are the boundaries shown in Figure 2.19.

Figure 2.19: Heat transfer through two materials with three interfaces

The weak form of Equation 2.80 is achieved from Equation 2.77

$$\int_{\Omega} w \left\{ \frac{\partial}{\partial x} \left(k_x \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial y} \left(k_y \frac{\partial T}{\partial y} \right) + \frac{\partial}{\partial z} \left(k_z \frac{\partial T}{\partial z} \right) + G \right\} d\Omega = 0 \quad (2.81)$$

Green's lemma derived from the integration by parts for arbitrary functions α_1 and α_2 over a domain Ω and a boundary S is

$$\int_{\Omega} \alpha_1 \frac{\partial \alpha_2}{\partial x_1} d\Omega = - \int_{\Omega} \frac{\partial \alpha_1}{\partial x_1} \alpha_2 d\Omega + \int_S \alpha_1 \alpha_2 n_1 dS \quad (2.82)$$

Using Green's lemma, Equation 2.81 is written in the form of

$$\begin{aligned} \int_S w \left(k_x \frac{\partial T}{\partial x} n_1 + k_y \frac{\partial T}{\partial y} n_2 + k_z \frac{\partial T}{\partial z} n_3 \right) dS \\ - \int_{\Omega} \left(\frac{\partial w}{\partial x} k_x \frac{\partial T}{\partial x} + \frac{\partial w}{\partial y} k_y \frac{\partial T}{\partial y} + \frac{\partial w}{\partial z} k_z \frac{\partial T}{\partial z} \right) d\Omega + \int_{\Omega_e} G w d\Omega \end{aligned} \quad (2.83)$$

Equation 2.83 is approximated using the boundary conditions in Equation 2.80 as well as the approximation of a dependent variable in Equation 2.78. Using the Galerkin method, the weights for both dependent variables and weak forms are the same (ϕ).

$$\begin{aligned} - \int_S \phi_k q dS + \int_S h \phi_k \phi_i T^i dS + \int_S h T_a \phi_k dS \\ - \int_{\Omega} \left(k_x \frac{\partial \phi_k}{\partial x} \frac{\partial \phi_i}{\partial x} + k_y \frac{\partial \phi_k}{\partial y} \frac{\partial \phi_i}{\partial y} + k_z \frac{\partial \phi_k}{\partial z} \frac{\partial \phi_i}{\partial z} \right) T^i d\Omega + \int_{\Omega_e} G \phi_k d\Omega \end{aligned} \quad (2.84)$$

From Equation 2.84, components of matrices f and K in Equation 2.79 are extracted as follows

$$\begin{aligned} K_{ki} = - \int_{\Omega} \left(k_x \frac{\partial \phi_k}{\partial x} \frac{\partial \phi_i}{\partial x} + k_y \frac{\partial \phi_k}{\partial y} \frac{\partial \phi_i}{\partial y} + k_z \frac{\partial \phi_k}{\partial z} \frac{\partial \phi_i}{\partial z} \right) d\Omega + \int_S h \phi_k \phi_i dS \\ f_k = - \int_S \phi_k q dS + \int_S h T_a \phi_k dS + \int_{\Omega_e} G \phi_k d\Omega \end{aligned} \quad (2.85)$$

Figure 2.20: Tetrahedral mesh geometries generated in Trelis a) A geometry of an elastic-viscoelastic medium for a subduction earthquake b) A simple geometry of an elastic-viscoelastic medium

Figure 2.21: Quadratic tetrahedral element as a shape function used in the Finite Element Method (FEM)

Shape functions

In Chapters 3 and 4, we discretise the 3D continuum into finite tetrahedral elements using a commercial software package Trelis² (e.g. Figure 2.20). Here we briefly describe the shape functions associated with the quadratic tetrahedral element in three dimensions. The natural coordinates of a tetrahedral element are $(\xi_1, \xi_2, \xi_3, \xi_4)$. The i^{th} component of the natural coordinates of the tetrahedron for point P is the ratio between the volume encompassed by the point and the opposite face to the corner i , and the total volume (Figure 2.21). Thus, $\xi_1 + \xi_2 + \xi_3 + \xi_4 = 1$ for each point. The shape functions for the quadratic tetrahedral element shown in Figure 2.21 are

²<http://www.csimsoft.com/help/trelishelp.htm>

$$\begin{bmatrix} 1 \\ x \\ y \\ z \\ u \end{bmatrix} = \begin{bmatrix} 1 & 1 & \cdots & 1 \\ x_1 & x_2 & \cdots & x_{10} \\ y_1 & y_2 & \cdots & y_{10} \\ z_1 & z_2 & \cdots & z_{10} \\ u_1 & u_2 & \cdots & u_n \end{bmatrix} * \begin{bmatrix} \phi_1 \\ \phi_2 \\ \vdots \\ \vdots \\ \phi_{10} \end{bmatrix} \quad (2.86)$$

Equation 2.86 shows how to calculate the coordinates of each point within a tetrahedron using the coordinates of the ten corners of the element and shape functions ϕ . Similarly, an arbitrary dependent variable u is approximated using the value of the dependent variable at ten corners and the shape functions. ϕ_k is a function of the natural coordinates of the tetrahedral element as below

$$\begin{aligned} \phi_1 &= \xi_1(2\xi_1 - 1) \\ \phi_2 &= \xi_2(2\xi_2 - 1) \\ \phi_3 &= \xi_3(2\xi_3 - 1) \\ \phi_4 &= \xi_4(2\xi_4 - 1) \\ \phi_5 &= 4\xi_1\xi_2 \\ \phi_6 &= 4\xi_2\xi_3 \\ \phi_7 &= 4\xi_3\xi_1 \\ \phi_8 &= 4\xi_1\xi_4 \\ \phi_9 &= 4\xi_2\xi_4 \\ \phi_{10} &= 4\xi_3\xi_4 \end{aligned} \quad (2.87)$$

Chapter 3

Postseismic displacement following the M_W 7.8 Dusky Sound earthquake, Fiordland, New Zealand, and its effect on the M_W 7.1 Darfield (Canterbury) earthquake 14 months later

Abstract

Continuous and campaign GPS observations reveal a broad postseismic displacement signal in the southern and central South Island of New Zealand in the months following the M_W 7.8 Dusky Sound earthquake of 15 July 2009. The 2010–2012 Canterbury earthquake sequence that commenced with the M_W 7.1 Darfield (Canterbury) earthquake on 4 September 2010 and involved the fatal M_W 6.2 Christchurch earthquake and other damaging aftershocks occurred within the region of post-Dusky Sound displacement. This raises a fundamental question: did stress perturbations in the Canterbury region induced by the Dusky Sound earthquake 14 months earlier trigger or otherwise affect the Darfield earthquake and the protracted aftershock sequence that ensued? To answer this question, we use an iterative approach to quantify time-dependent afterslip following the Dusky Sound earthquake, and the coupled viscoelastic response to this afterslip and the original mainshock. The main patches of afterslip are inferred to have accommodated maximum cumulative slip exceeding 1 meter within 400 days of the mainshock. Our preferred model of afterslip involves oblique slip mainly at depths between 0–80 km in an elastic-viscoelastic medium with a Maxwell rheology and a viscosity higher than 10^{18} Pa·s. After estimating the rheology and time-dependent afterslip, we calculate the Coulomb stress changes resolved on the faults involved in the Darfield event as well as the cumulative strain in the region. We estimate positive Coulomb stress changes of a few kilopascals immediately prior to the Darfield event due to both coseismic and postseismic mechanisms following the Dusky Sound earthquake. However, neither very small positive Coulomb stress changes of a few kilopascals nor

the estimated strain accumulation seems to have advanced the Darfield earthquake.

3.1 Introduction

The M_W 7.8 Dusky Sound earthquake of 15 July 2009 was the third-largest earthquake of that year globally¹, and the largest New Zealand earthquake since the M_W 7.8 Hawke’s Bay earthquake of 1931². Despite its size, the earthquake caused little damage in southern New Zealand as a consequence of its occurrence in a little-populated area of the country and its south west-ward rupture direction (Fry et al., 2010; Beavan et al., 2010a). Moreover, its effects were quickly eclipsed by those of the 4 September 2010 Darfield (Canterbury) earthquake, which originated 37 km west of Christchurch, and the ensuing protracted aftershock sequence (Beavan et al., 2010b; Gledhill et al., 2010; Holden et al., 2011; Gledhill et al., 2011; Quigley et al., 2012). That sequence included the highly damaging Christchurch earthquake on 22 February 2011, which caused 185 fatalities (Holden, 2011; Fry et al., 2011; Bannister et al., 2011; Beavan et al., 2011; Kaiser et al., 2012). The spatial and temporal proximity of the events lead us to investigate whether the Canterbury earthquake sequence of 2010–2012 may have been triggered by the Dusky Sound earthquake.

In order to understand the interaction between earthquakes, and especially near- or far-field triggering of large earthquakes, it is important to consider the effects of postseismic as well as coseismic displacements. In principle, the effects of coseismic stress changes may be reinforced, diminished, or completely counteracted by postseismic processes such as the following:

- Stress transfer from the ductile lower crust and upper mantle to the brittle upper crust (e.g. Freed and Lin, 1998, 2001; Rydelek and Sacks, 2001; Freed, 2005; Pollitz et al., 2006a; Freed et al., 2007; Tabrez et al., 2008; He et al., 2011; Decriem and Árnadóttir, 2012);
- Changes in effective stress associated with pore fluid pressure redistribution (e.g. Cocco and Rice, 2002; Bosl and Nur, 2002; Masterlark and Wang, 2002; Piombo et al., 2005; Hughes et al., 2010, 2011); and
- Changes in stress caused by time-dependent frictional afterslip (e.g. Hearn et al., 2002a; Perfettini and Avouac, 2004; Perfettini and Ampuero, 2008; Murray-Moraleda and Simpson, 2009; Cattania et al., 2015).

Freed and Lin (1998) proposed that the first of these mechanisms caused lateral ex-

¹<http://www.emsc-csem.org/Earthquake> and <http://earthquake.usgs.gov/earthquakes/>

²<http://www.gns.cri.nz>

pansion of the region of positive coseismic Coulomb stresses following the M_W 6.5 1971 San Fernando earthquake. Similarly, Freed and Lin (2001) modelled viscoelastic stress transfer seven years following the M_W 7.3 1992 Landers earthquake and concluded that positive Coulomb stress changes of up to 0.1–0.2 MPa were produced at the hypocentre of the ensuing M_W 7.1 1999 Hector Mine earthquake. In comparison, Pollitz and Sacks (2002) calculated Coulomb stress increases of approximately 0.07 MPa on the Hector Mine fault surface due to postseismic relaxation. Hearn et al. (2002a) computed post-seismic Coulomb stress changes of 0.1 MPa in response to modelled frictional afterslip in the time interval between the M_W 7.6 1999 İzmit and M_W 7.2 Düzce earthquakes at the Düzce hypocentre. In that study, rheological models incorporating features such as a layered elastic structure at the Düzce hypocentre, a viscous shear zone, and viscoelastic lower crust in the Marmara region yielded greater Coulomb stress changes than those estimated from uniformly elastic frictional afterslip.

Coulomb stress changes in a poroelastic medium due to the redistribution of pore fluids after earthquakes were investigated by Piombo et al. (2005). They inferred from their simulation that the effect of pore fluid diffusion was to alter Coulomb stresses from initially negative values to significantly positive values at some locations. In the same way, pore-pressure recovery following the M_W 9.1 2004 Sumatra-Andaman earthquake has been interpreted to have clock-advanced the M_W 8.6 2005 Nias earthquake (Hughes et al., 2010, 2011).

Based on these and other studies, and taking into account the large moment of the Dusky Sound earthquake, we describe and interpret the postseismic displacement that followed, and consider the effects of this displacement on stress and strain in the vicinity of the Darfield earthquake 14 months later. We first estimate afterslip and viscoelastic effects based on spatial and temporal patterns of displacement measured geodetically. Next, the stress changes and strain associated with these two displacement processes are computed at the position of the Darfield earthquake hypocentre, ~ 600 km away, in order to examine the interaction between the two earthquakes. We presume the effects of pore fluid redistribution to be negligible at the distance from the Darfield earthquake (~ 600 km). Distances of about 2.5 fault lengths (here ~ 200 km) are assumed to be affected by Coulomb stress changes induced by fluid distribution (Piombo et al., 2005).

3.1.1 Tectonic setting

We summarise the main features of present New Zealand plate tectonics from northernmost to southernmost as follows: 1) subduction of the Pacific plate along the Hiku-

rangi Margin beneath the North Island (Barnes and Lépinay, 1997; Barnes et al., 1998; Reyners, 1998; Eberhart-Phillips and Reyners, 1999; Wallace et al., 2004; Barker et al., 2009); 2) a southward transition to transpressive slip along the Alpine Fault (Van Dis-sen and Yeats, 1991; Holt and Haines, 1995; Beavan et al., 1999; Norris and Cooper, 2001); and 3) eastward subduction of the Australian plate beneath the Pacific plate along the Puysegur Trench (Norris et al., 1990; Figure 4.6).

The fault surface associated with the Dusky Sound earthquake forms the plate bound-ary along the Puysegur subduction interface, dipping eastwards beneath southwestern New Zealand in the Fiordland region (Fry et al., 2010). Understanding the plate boundary, the geometry of the slab, the thickness of the elastic layer, and the heteroge-neous lithospheric structure are of great importance in adapting our models for more complexities, as discussed further below.

Lebrun et al. (2000) located the Pacific-Australian plate boundary between the Alpine Fault and Puysegur Trench using bathymetric data. They also emphasised the dif-ference between the northern subduction transition at the Hikurangi Margin and the southern transition at the Fiordland Margin. The northern transition occurs gradu-ally via strain partitioning, whereas the southern transition involves abrupt changes in the strike direction and dip angle. This abrupt transition is associated with tearing of the subducting plate, which creates a crustal discontinuity at the Resolution Ridge System (Lebrun et al., 2000).

Subducting of the Australian plate beneath the Pacific plate in the Fiordland region accommodates the major component of plate motion. Near Doubtful Sound a sharp direction change in the strike of the subducting slab from 023° to 040° occurs from south to north (Figure 4.6 shown in the circle; Reyners and Webb, 2002). This bending of the subducting slab leads to a positive Bouguer gravity anomaly in the Fiordland region in a non isostatic manner (Malservisi et al., 2003).

Reyners et al. (1991) noted a contrast in dipping seismic zones from less steep in the south-west to near vertical north-east of the 4 June and 19 July 1988 Te Anau earthquakes. Down-dip tearing of the subduction zone was inferred from the orientation of the fault surfaces (Reyners et al., 1991), as previously suggested by Smith and Davey (1984). Smith and Davey (1984) concluded that the intermediate-depth events north-east of the Te Anau earthquake are situated on a rotated section of the Australian plate interface.

Eberhart-Phillips and Reyners (2001) imaged the Fiordland subduction zone using 3-D seismic tomography. They noted the major crustal features as follows: 1) low velocity

Figure 3.1: Tectonic setting of the Fiordland region and focal mechanisms of recent earthquakes, namely the 1988 Te Anau, 1989 Doubtful Sound, 1993 Secretary Island, 2000 Thompson Sound, 2003 Fiordland and 2009 Dusky Sound earthquakes, which were widely felt in the South Island, New Zealand. Subduction zone steepening and change of strike from south to north of Doubtful Sound are shown in the circle (Reyners and Webb, 2002).

within the upper 10 km beneath the Te Anau and Waiau basins, 2) Western Fiordland Orthogneiss with high velocity V_p and low ratio V_p/V_s , 3) high velocity Pacific mantle next to the deepest earthquakes, 4) high V_p/V_s ratio where the subduction slab bends and expresses a distributed fracturing and finally 5) low velocity crustal root in the transition to the Alpine Fault, northern Fiordland.

3.1.2 Recent seismicity in southern New Zealand

We here describe the characteristics of recent large earthquakes in the Fiordland region, namely the 1989 Doubtful Sound, 1993 Secretary Island, 2000 Thompson, 2003 Fiordland and 2009 Dusky Sound earthquakes, each of which was widely felt in the South Island. The focal mechanisms of these events are illustrated in Figure 4.6.

The M_W 6.4 Doubtful Sound earthquake on 31 May 1989 involved oblique slip on a reverse fault (Anderson et al., 1993). The southeast-dipping nodal plane computed from first motion data was inferred to be the fault surface assuming that the mainshock occurred on the plate interface (Reyners and Webb, 2002). Four years later, the shallower (22 km) M_W 6.8 Secretary Island earthquake on 10 August 1993 initiated a rich aftershock sequence (Reyners and Webb, 2002). The rupture zone from this event seems to abut that of the earlier Doubtful Sound earthquake and lies within a zone of positive Coulomb stress changes, beneath the surface trace of the eastern strand of the Alpine Fault (Reyners and Webb, 2002). The results of body-wave inversion suggest that the Secretary Island earthquake involved nearly pure reverse faulting. The larger number of aftershocks associated with this earthquake than the Doubtful Sound event may reflect its shallower depth and depth-dependent variations of frictional properties at the plate interface (Reyners and Webb, 2002).

The Thompson Sound earthquake occurred on 1 November 2000. Robinson et al. (2003) relocated the mainshock and aftershock hypocentres using the double-difference method and demonstrated that the aftershocks define a west-dipping zone extending from 12 km to 24 km.

Teleseismic body-wave inversions indicate that the Thompson Sound earthquake involved reverse slip with a subordinate left-lateral component. The cumulative Coulomb stress changes produced by a M_W 6.7 earthquake near Te Anau in 1988 and the 1989 Doubtful Sound and 1993 Secretary Island earthquakes were positive on the Thompson Sound rupture surface.

The M_W 7.2 Fiordland earthquake on 21 August 2003 was the largest earthquake in southwest New Zealand in the 41 years prior to the Dusky Sound event (Reyners et al., 2003; McGinty and Robinson, 2007). Reyners et al. (2003) analysed the fault geometry of the Fiordland earthquake using campaign GPS measurements acquired shortly after the event between 29 August and 3 September 2003. Their model indicates that the earthquake was a shallow thrust event on the plate interface with a strike angle of 030° , a finding consistent with centroid moment tensor calculations. McGinty and Robinson (2007) noted that the aftershock distribution of the Fiordland earthquake did not define

a planar rupture surface. They interpreted this event to have been triggered by the Secretary Island earthquake a decade earlier.

The largest earthquake to strike South Island of New Zealand in 80 years was the M_W 7.8 Dusky Sound earthquake. Based on analysis of strong-motion records, Fry et al. (2010) inferred that rupture initiated at a depth of 38 km and proceeded upwards and southwestwards at a velocity of approximately 2.6 km/s.

Beavan et al. (2010a) modelled the fault geometry and slip distribution of the Dusky Sound earthquake using campaign and continuous GPS data and ascending and descending tracks of the ALOS/PALSAR radar satellite. Their results suggest a maximum slip of ~ 6 m occurring on a curve-shaped patch of approximately 4000 km² area, equivalent to a scalar moment of $\sim 4.0 \times 10^{20}$ N·m. Li et al. (2011) also computed fault geometry and slip parameters and obtained similar results.

Despite its large size, the Dusky Sound earthquake's effects were eclipsed internationally by the very damaging earthquakes and tsunami that struck Tonga and Samoa in September 2009, and in the New Zealand public's mind by the much closer-to-home Darfield earthquake and then the damaging Christchurch earthquake. Approximately 14 months after the Dusky Sound earthquake, the Darfield earthquake nucleated 37 km west of Christchurch. This earthquake originated on the Charing Cross Thrust Fault (Holden et al., 2011; Gledhill et al., 2010, 2011) and propagated eastwards along the previously unknown Greendale Fault (Gledhill et al., 2011; Kaiser et al., 2012; Quigley et al., 2012). Following the mainshock, a prolonged aftershock sequence ensued, resulting with the 22 February 2011 Christchurch earthquake only 10 km from the central business district (Holden, 2011; Bannister et al., 2011; Beavan et al., 2011). That earthquake resulted in the loss of 185 lives and extensive damage to buildings and other infrastructure.

3.2 Methods

3.2.1 Observations

The observational data used in this chapter are 1) continuous GPS (cGPS) observations made at 16 sites in the southern to central South Island (locations and time series are represented in Figure 4.6 and Figure 3.4, respectively)³; 2) a campaign GPS dataset acquired shortly after the earthquake in August 2009 (Beavan et al., 2010a); and 3)

³Processed by Dr. Euan Smith

Differential Interferometric Synthetic Aperture Radar (DInSAR) observations of post-seismic displacement, acquired by PALSAR/ALOS satellite for the July 15–August 30 ascending path and the July 23–September 7 descending path (Beavan et al., 2010a).

Daily recordings of GPS data from the New Zealand network (GeoNet) were de-trended to determine mean secular velocities, as follows. First the longest period unaffected by regular or slow earthquakes prior to the Dusky Sound earthquake was identified by inspection. A weighted regression displacement against time was fitted. Weights were calculated as reciprocals of error estimates supplied by GeoNet. The residuals from the regression were then stacked to estimate short-term position changes (Figure 3.2; cf. spatial filtering; e.g. Wdowinski et al. 1997). This was removed from the time series and a seasonal effect was calculated using a 30 day moving average, which was also subtracted from the time series. The regression was then re-fitted to obtain velocities and uncertainty estimates. Mean residual variance reduction was 87%.

The data from selected sites for the inter-earthquake period between the Dusky Sound and Darfield earthquakes (IEP; 416 days) was then detrended using the secular velocities. Sites from the eastern and central North Island were not considered because of ongoing slow earthquakes in the Hikurangi subduction zone (Douglas et al., 2005; McCaffrey et al., 2008; Wallace and Beavan, 2010). Co- and post-Dusky displacement was apparent on most South Island sites (Beavan et al. 2010). Three components of displacement were visible on sites within 200 km of the Dusky Sound centroid and used to represent the earthquake as a point source. The largest component was a westward movement, which was 3–5 times the southward component and > 7 times the vertical component. Inspection suggested that the ongoing postseismic displacement was growing logarithmically. A model for the co- and post-Dusky displacement was fitted as follows. First an initial model for the displacement was fitted to an ensemble of sites, of the form

$$d(t) = d_0 + L \log \left(1 + \frac{t}{c} \right) + e \quad (3.1)$$

where d is displacement, t is time (days), c is a timescale factor, d_0 is displacement at $t = 0$, L is the postseismic productivity and e is an error, assumed Gaussian. For a set of observed displacements d_j and times t_j and a specified value of c , the maximum likelihood estimates of d_0 and L can be found by (weighted) least squares regression of d against $\log(1 + t/c)$.

For the first fitting, c was set to 1 day for all sites. Inspection showed that the model fit was not sensitive to the choice of c . However, for the stronger signals the best fit for a range of c values was given by $c = 0.5$ or $c = 1$ day.

Figure 3.2: Stack of residuals pre-Dusky Sound earthquake from 22 sites.

Residuals from all but the nearest three sites were stacked for spatial filtering and a 30 point moving average of the filtered residuals was then calculated and subtracted. This method estimates the actual seasonal signal, including inter-seasonal variability (see Figure 3.3). Outliers amongst the residuals were down-weighted using Jeffreys' weighting, which downweights outliers in proportion to the probability that they arise from a central Gaussian distribution (Jeffreys, 1961). The fitted model represented by Equation 3.1 was then added back to the resulting time series. For the westward component, the best-fitting c value was now 0.5 day at all sites except the nearest one (PYGR) where it was 1 day. For the final model fitting, $c = 0.5$ day was adopted for all components at all sites. The final, overall Root Mean Square Error (RMSE) of the data-model was 1.1 mm (22 sites).

3.2.2 Modelling time-dependent afterslip

At time t , the elastic-viscoelastic displacement $\vec{d}$ in an arbitrary direction is presumed to be expressed by the superposition of responses to unit-step slip ($\vec{s}$) occurring at time $\tau < t$. Accordingly,

$$\vec{d}(t) = \vec{d}^c(t) + \int_0^t G(t - \tau) \vec{v}(\tau) d\tau \quad (3.2)$$

Figure 3.3: Selection of seasonal models, at HAAS, DUND KARA and CNCL sites, derived from 30 point moving averages as described in Section 3.2.1.

where G is the matrix of corresponding elastic-viscoelastic Green's functions and $\vec{v}$ is the afterslip rate. By discretising the integral, we approximate Equation 3.2 as

$$\vec{d}(t_j) \simeq \vec{d}^c(t_j) + \sum_{k=1}^{j-1} G(t_j - t_k) \vec{s}_k \quad (3.3)$$

In Equations 3.2 and 3.3, the superscript c indicates the viscoelastic displacement in response to coseismic stress changes.

At this point we have not discussed the viscoelastic response to time-dependent afterslip and we discuss how to account for this response in Section 3.2.3. We compute the viscoelastic response to the coseismic event and then subtract this from the observed displacement to leave $\vec{d}$ (the second term in Equations 3.2 and 3.3). The remainder is the elastic displacement due to time-dependent afterslip mechanism. Using elastic

Figure 3.4: cGPS displacement observations (blue symbols) at PYGR, MAVL, BLUF, LEXA, HAAS, DUND sites. Locations of the sites are presented in Figure 4.6.

Green's functions (G_e) Equation 3.3 can then be rewritten as follows

$$\begin{aligned}
 \vec{d}_{n \times 1}^1 &= G_e \vec{s}_{m \times 1}^1 \\
 \vec{d}_{n \times 1}^2 &= G_e \vec{s}_{m \times 1}^1 + G_e \vec{s}_{m \times 1}^2 \\
 \vec{d}_{n \times 1}^3 &= G_e \vec{s}_{m \times 1}^1 + G_e \vec{s}_{m \times 1}^2 + G_e \vec{s}_{m \times 1}^3 \\
 &\vdots \\
 \vec{d}_{n \times 1}^T &= G_e \vec{s}_{m \times 1}^1 + G_e \vec{s}_{m \times 1}^2 + \dots + G_e \vec{s}_{m \times 1}^T
 \end{aligned} \tag{3.4}$$

cGPS displacement observations (blue symbols) at KARA, CNCL, QUAR, WAKA, WAIM and MTJO sites. Locations of the sites are presented in Figure 4.6.

where n is the number of displacement components, m the number of slip components on the fault surface and T the number of epochs. G_e represents the elastic Green's functions with dimensions of $n \times m$. The displacement vector at epoch T is written in terms of cumulative afterslip ($\vec{S}^T$) as

$$\vec{d}_{n \times 1}^T = G_e \vec{S}_{m \times 1}^T \quad (3.5)$$

cGPS displacement observations (blue symbols) at HOKI, MQZG, LKTA and KAIK sites. Locations of the sites are presented in Figure 4.6.

Equation 3.5 is expressed in matrix form as

$$D_{n \times T} = G_e \mathbf{S}_{m \times T} \quad (3.6)$$

where D and $\mathbf{S}$ are the time-dependent displacement and cumulative afterslip matrices, respectively.

The strategy underpinning our time-dependent afterslip calculations is based on a Singular Value Decomposition (SVD; see Forsythe and Moler, 1967) of the spatio-temporal displacement signal into a space function U and a time function V (Kositsky and Avouac, 2010). In other words, we represent the matrix D consisting of displacement components measured at T epochs as

$$D = U \Lambda V' \quad (3.7)$$

where U is an $n \times n$ matrix whose columns are the left singular vectors of D , V is a $T \times T$ matrix whose columns are the right singular vectors of D , and Λ is an $n \times T$ diagonal matrix containing the singular values of D .

Given the linearity of the problem, the slip function with m components is simply expressed as

$$\mathbf{S} = G_e^{-1} U \Lambda V' \quad (3.8)$$

where the $n \times m$ matrix G_e is calculated from elastic Green's functions relating the fault slip to the surface displacement (Okada, 1985). G_e^{-1} refers to the regularised inverse of G_e .

The Principal Component Analysis-based Inversion Method (PCAIM) introduced by Kositsky and Avouac (2010), without considering the viscoelastic effects, simplifies the time-dependent inversion process to one in which the inversion step is applied only to the spatial function (U) obtained by Singular Value Decomposition (SVD) and temporal information is retrieved from the time function (V). In this method, temporally dense information such as continuous GPS or InSAR time series are used to decompose the displacement signal into spatial and time functions. Additionally, we can constrain the spatial pattern of fault slip (U_s) using temporally sparse data such as InSAR or campaign GPS data (D_{sparse} ; Kositsky and Avouac, 2010). If $\mathbf{S}$ is written in terms of slip spatial and temporal functions:

$$\mathbf{S} = U_s \Lambda_s V_s' \quad (3.9)$$

using Equation 3.6

$$D = G_e U_s \Lambda_s V_s' \quad (3.10)$$

$$V_s^t = \Lambda_s V_s' \quad (3.11)$$

then

$$D = G_e U_s V_s^t \quad (3.12)$$

where U_s , Λ_s and V_s' are the matrices of spatial, scaling and time functions of fault slip, respectively ($'$ denotes matrix transpose). V_s^t is expressed as the multiplication of the singular values of the slip matrix by the slip time function. If the campaign dataset spans the time interval between t_1 and t_2 , the sparse constraint equation for a set of sparse observations is written as

$$D_{sparse} = G_e [V_s^t(t_2) - V_s^t(t_1)] U_s \quad (3.13)$$

The PCAIM method decomposes the observed displacement signals into components of decreasing singular value and thus of successively lower significance. Correspondingly, instead of using the full SVD representation of the displacements D , we can restrict the analysis to only the first r components, in which case

$$D_r = U_r \Lambda_r V_r' \quad (3.14)$$

Here D_r is a filtered form of the full displacement matrix D , of the same dimensions $n \times T$; U_r is of dimension $n \times r$, representing the first r columns of U ; Λ_r is of dimension $r \times r$, formed from the first r rows and columns of Λ ; and V_r is of dimension $T \times r$ and represents the first r columns of V . The optimal value of r is chosen by minimising the misfit between the displacement signal D and the filtered signal D_r (Kositsky and Avouac, 2010). Filtering by single value truncation is performed assuming the single values are sorted from high to low.

3.2.3 Modelling coupled viscoelastic rebound and afterslip

Postseismic displacement following large earthquakes has generally been modelled in terms of three separate mechanisms: 1) short-term afterslip in the upper/lower crust on the mainshock fault surface (e.g. Savage and Church, 1974; Johnson et al., 2012; Wen et al., 2012a; Reddy et al., 2013); 2) poroelastic rebound (e.g. Booker, 1974; Rice and Cleary, 1976; Barbot and Fialko, 2010b; Lubis et al., 2013); or 3) long-term viscoelastic relaxation in the lower crust/upper mantle (e.g. Elsasser, 1967; Savage and Prescott, 1978; Pollitz et al., 2000; Pollitz, 2003b; Rousset et al., 2012b). Here, we allow for short-term time-dependent afterslip and longer-term viscoelastic rebound to both occur during the postseismic time interval in response to deviatoric stresses induced coseismically (Barbot and Fialko, 2010a).

A simple Maxwell rheology consisting of an elastic layer overlying a viscoelastic half-space of uniform viscosity and hence uniform relaxation time is the most widely adopted representation of viscoelastic relaxation (e.g. Koseluk and Bischke, 1981; Pollitz and Dixon, 1998; Hao et al., 2012). Multiphase relaxation can be treated using either more complex rheologies, such as a bi-viscous Burger's body with two phases of viscoelastic creep (e.g. Pollitz, 2003b; Hetland and Hager, 2005) or a Standard Linear Solid (SLS) (e.g. Pollitz et al., 2000; Decriem and Árnadóttir, 2012), or by using different layerings, such as that consisting of a viscoelastic layer sandwiched between an elastic upper layer and a viscoelastic lower half-space (e.g. Hetland and Hager, 2005).

In several previous studies, other factors such as coupling between displacement and gravity field (e.g. Fernández and Rundle, 2004; Wang, 2005; Wang et al., 2006), lateral heterogeneity (e.g. Pollitz, 2003a), stratification (e.g. Hetland and Hager, 2005) and compressibility (e.g. Tanaka et al., 2006) have been considered in order to provide more realistic models. As a first attempt at examining the postseismic displacement induced by the Dusky Sound earthquake, we consider only simple Maxwell rheologies of various viscosities and a single bi-viscous Burger's body in an elastic-viscoelastic medium. In

each case, we use a homogeneous elastic layer with a thickness of 45 km, appropriate for this region based on the 3-D seismic velocity model of New Zealand (Eberhart-Phillips et al., 2010), overlying a viscoelastic half-space. This simple viscoelastic forward model is used in this chapter due to several limitations: 1) sparsity of the continuous GPS measurements in the central and southern South Island; 2) lack of campaign GPS measurements spanning the early postseismic interval excluding the coseismic signal; 3) lack of sufficient structural data. The fault surface is approximately inferred by Beavan et al. (2010a) from initial geodetic and seismic models (Fry et al., 2010); and 4) solving the problem in an inversion framework rather than forward modelling.

Our modelling strategy is illustrated schematically in Figure 3.5. To couple viscoelastic and afterslip models, we first compute the viscoelastic response to coseismic stress changes using the coseismic slip model of (Beavan et al. 2010a; $D_{\text{co-viscous}}$) and the numerical code PyLith (Aagaard et al., 2009, 2013, 2015). We subtract the viscoelastic response from the observations (D_{pseudo}), to remove the viscoelastic response to the coseismic event. Next, after centering the time series of the displacement about zero, we invert the remaining signal using elastic Green’s functions to extract the time-dependent afterslip using the decomposition technique described in the previous section. In doing so we must account for the fact that the stress changes associated with afterslip can themselves produce a viscoelastic response that has not yet been accounted for. Therefore, we re-estimate displacement time series as the response to the previously inferred time-dependent afterslip in an elastic-viscoelastic medium with the assumed rheology in PyLith. If the viscoelastic response to the previous afterslip model is significant enough, the displacement can vary from the previously calculated time series.

To address this, we perform an iterative calculation in which a correction term equal to the difference between observed and calculated displacements $D_{\text{correction}} = D - D_{\text{PyLith}}$ is added to the previous calculation of ground displacement to incorporate the viscoelastic response to afterslip. The lower the viscosity of the viscoelastic medium, the larger the viscoelastic response to afterslip becomes, meaning that more iterations are required for convergence. For reasons discussed below, the lowest viscosity used here is 5×10^{18} Pa·s and even in this case only one iteration is required to fit the observations.

We note that using elastic Green’s functions throughout the PCAIM inversion method does not allow us to incorporate the viscoelastic effects directly in our inversion. Alternatively, time-dependent numerical techniques such as Aagaard et al. (2009, 2013, 2015) or semi-analytical propagator matrix methods of Fukahata and Matsu’ura (2006) for instance, are used to estimate the viscoelastic Green’s functions and thus to relate

Figure 3.5: Flowchart illustrating the iterative approach used in order to couple time-dependent afterslip and viscoelastic rebound.

slip vectors to ground displacements measured geodetically. Adapting the PCAIM-inversion method in order to accommodate viscoelastic Green’s functions is the subject of further study.

3.3 Results

Figure 3.6 illustrates viscoelastic responses to coseismic slip for Maxwell rheologies with viscosities of $10^{17} - 10^{20}$ Pa·s and for a Burger’s body with a short-term transient viscosity of 10^{17} Pa·s and steady-state viscosity of 10^{20} Pa·s. To illustrate the spatial extent and characteristics of the viscoelastic displacement, we illustrate in Figure 3.7 the near-field horizontal and vertical ground displacement and the far-field horizontal ground displacement for the Burger’s model.

It is evident from Figure 3.6 and 3.7 that the viscoelastic response to coseismic stress changes does not account for several of the features observed in the cGPS time series. In particular, neither the observed southward motion of sites such as PYGR located within the dashed box in Figure 3.7a nor the spatial and temporal patterns of the observed ground displacement are well fit. Lower values of viscosity, such as 10^{17} Pa·s, yield rapid decays in the postseismic displacement for sites near the epicentral area, which is consistent with observations of the east component of displacement, but at far-field sites the decay rate of calculated displacement exceeds that observed.

To improve the fit with geodetic observations, we next consider the effects of time-dependent afterslip coupled with viscoelastic rebound. Appropriate values for viscosity

Figure 3.6: cGPS displacement observations (blue symbols) and predicted viscoelastic displacement induced by the earthquake (coloured lines) calculated for a Maxwell rheology of different viscosity values (10^{17} , 10^{18} , 5×10^{18} , 10^{19} , 5×10^{19} and 10^{20} Pa.s) and a Burger's body with transient and steady-state viscosity values of 10^{17} and 10^{20} Pa.s, at PYGR, MAVL, BLUF, LEXA, HAAS, DUND sites.

are chosen by decomposing the displacement time series after subtracting the viscoelastic response to coseismic slip. Figure 3.8 illustrates the first and second temporal modes of the displacement after removing viscoelastic responses to the earthquake induced in Maxwell materials of two different viscosities, 10^{18} Pa.s and 5×10^{18} Pa.s. The first

cGPS displacement observations (blue symbols) and predicted viscoelastic displacement induced by the earthquake (coloured lines) calculated for a Maxwell rheology of different viscosity values (10^{17} , 10^{18} , 5×10^{18} , 10^{19} , 5×10^{19} and 10^{20} Pa·s) and a Burger's body with transient and steady-state viscosity values of 10^{17} and 10^{20} Pa·s, at KARA, CNCL, QUAR, WAKA, WAIM and MTJO sites.

temporal modes (i.e. the first column of V ; V_1) shown in blue resemble each other for both viscosities, whereas the second temporal modes V_2 shown in red are distinctly different. For a viscosity value of 10^{18} Pa·s, V_2 reaches a maximum value almost 100 days after the mainshock; in contrast, for a viscosity value of 5×10^{18} Pa·s (illustrated)

cGPS displacement observations (blue symbols) and predicted viscoelastic displacement induced by the earthquake (coloured lines) calculated for a Maxwell rheology of different viscosity values (10^{17} , 10^{18} , 5×10^{18} , 10^{19} , 5×10^{19} and 10^{20} Pa·s) and a Burger's body with transient and steady-state viscosity values of 10^{17} and 10^{20} Pa·s, at HOKI, MQZG, LKTA and KAIK sites.

or higher, V2 exhibits only random variations.

As discussed in Section 3.2.2, in an elastic medium a slip distribution in time maps directly to the time function of the decomposed ground displacement. Therefore, a temporal mode with a turning point either implies a change of slip direction on the fault, the existence of a different source of displacement in the opposite direction, or under-estimation of viscosity value. We consider a change in the direction of afterslip to be physically unreasonable. If the viscoelastic response to afterslip occurs in the opposite direction resulting in a temporal mode with a turning point, the method described in flowchart 3.5 cannot be used. The reason is that we are limited to perform the inversion on the observational data using elastic Green's functions. Accordingly, based on the direction change in the second temporal mode associated with viscosities equal to or lower than 10^{18} Pa·s, we adopt viscosities of 5×10^{18} – 10^{20} Pa·s for the coupled model.

Figure 3.7: a) Horizontal, b) Vertical and c) Regional horizontal components of accumulated displacement due to the viscoelastic rebound of a Burger's body with a transient viscosity of 10^{17} and steady-state viscosity 10^{20} Pa.s. Corresponding time series of displacements are shown as magenta in Figure 3.6.

After choosing appropriate values for viscosity, we perform the inversion method. In principle, the iterative calculations must be continued until the modelled ground displacement from the coupled afterslip and viscoelastic displacement match the geodetic observations. We find that only one iteration is required to produce satisfactory fit with GPS and InSAR observations. We can explain that to be the result of insignificant viscoelastic response to time-dependent afterslip with a maximum slip value of ~ 1 m, accumulated within 400 days.

We consider the extension of the coseismic fault surface to be 20 km on each side and to a maximum depth of 200 km. Throughout the iterative approach, four options have been considered for the boundary of the estimated slip on the extended fault surface, Option 1: unconstrained slip boundary (Figure 3.9a); Option 2: no slip-boundary

Figure 3.8: First and second temporal modes (V1 and V2) obtained by decomposing the cGPS data after removing the coseismic viscoelastic rebound effects with a viscosity of a) 10^{18} b) and 5×10^{18} Pa.s. For the first viscosity, we can see that the results of SVD method suggest direction change within 400 days. Therefore we only run the models with viscosities higher than 10^{18} Pa.s.

constraint on the sides, top and bottom of the fault (Figure 3.9b); Option 3: no-slip boundary constraint on the sides of the fault (Figure 3.9c); and Option 4: no-slip boundary constraint on the sides and bottom of the fault (Figure 3.9d). Misfits with cGPS observations are measured in terms of RMSE values. RMSE values of 1.8 mm are computed for Option 1, Option 3 and Option 4 with a viscosity value of 5×10^{18} Pa.s and an elastic thickness of 45 km. A RMSE value of 1.9 mm is obtained from Option 2 with no-slip boundary constraint on the sides, top and bottom of the fault.

The campaign GPS data acquired in August 2009 include the coseismic signal as well as early postseismic effects. consequently, direct measurements of the early postseismic displacement for the campaign GPS sites are not available. We subtract the coseismic signal estimated by Beavan et al. (2010a) from the observed displacement, in order to extract the early postseismic displacement at the campaign GPS sites. Figure 3.10a and b illustrate fits with horizontal and vertical displacements calculated from the postseismic coupled model spanning the time interval reoccupied by the campaign observations (Option 4; observation and modelling results are presented in blue and cyan, respectively). We partially associate the misfit with the residuals of coseismic displacement estimated by Beavan et al. (2010a). The sites with poor accuracy of GPS measurements based on the uncertainty values published by Beavan et al. (2010a) are indicated with grey stars.

We assign uniform weights to the campaign and InSAR measurements. Figure 3.11 illustrates the observed and estimated displacements of the postseismic model for Option

Figure 3.9: Accumulated slip 400 days after the Dusky Sound earthquake. a) Option 1: unconstrained slip boundary. b) Option 2: no slip-boundary constraint on the sides, top and bottom of the fault. c) Option 3: no-slip boundary constraint on the sides of the fault. d) Option 4: no-slip boundary constraint on the sides and the bottom of the fault. Viscosity and the thickness of elastic layer are assumed to be 5×10^{18} Pa·s and 45 km, respectively.

Figure 3.10: a) Horizontal and b) Vertical fits with the early postseismic displacement (August 2009) for the coupled viscoelastic rebound with a viscosity of 5×10^{18} Pa·s and time-dependent afterslip (Option 4). The coseismic displacement estimated by Beavan et al. (2010a) is subtracted from the measurements in a-b.

4 in ascending (a, b) and descending ALOS orbits (c, d).

For each of Option 1-4, we identify two main patches located on either sides of the coseismic slip patch (white contours in Figure 3.9) reaching to the ground surface and

Figure 3.11: a) Observed and b) Estimated postseismic displacements, calculated from Option 4 of the coupled model, projected in the Line Of Sight (LOS) direction of ALOS ascending orbit. c) Observed and d) Estimated postseismic displacements, calculated from Option 4 of the coupled model, projected in the LOS direction of ALOS descending orbit.

depth 80 km. The coupled viscoelastic and afterslip model accounts well for both the spatial and temporal patterns of postseismic displacement with RMSE of 1.8 mm for Option 4.

Figure 3.12 shows the displacement time series corresponding to Option 4 of the coupled model (red) and observations (blue) at 16 continuous GPS sites. The patterns of ground displacement due to the same coupled model, with a viscosity of 5×10^{18} Pa·s and a 45 km thick elastic layer, for a Maxwell material are illustrated in Figure 3.13. The accumulated far-field horizontal ground displacements, 400 days after the earthquake estimated from the coupled time-dependent afterslip and viscoelastic rebound model are presented in Figure 3.13c. The figure shows larger responses but similar directions compared with the ground displacement calculated from the viscoelastic rebound model of the Burger's body (Figure 3.13c compared to Figure 3.7c). In the western portion of the dashed box, the purely viscoelastic model predicts north-directed movement (Figure 3.7a), while the coupled model predicts southward movement (Figure 3.13a). Station "PYGR", which is located within this region, displays movement that is more

Figure 3.12: cGPS displacement observations (blue symbols) and predicted coupled afterslip with viscoelastic rebound (red lines) calculated for a Maxwell rheology of viscosity 5×10^{18} Pa·s, at PYGR, MAVL, BLUF, LEXA, HAAS, DUND sites (Option 4).

consistent with the coupled model. Subsidence patch located in-land from the coast and a partially off-shore uplift patch are the main features of the vertical displacement calculated 400 days after the earthquake from the postseismic model (Figure 3.13b). The vertical component shows an uplift signal at the nearest site “PYGR”.

cGPS displacement observations (blue symbols) and predicted coupled afterslip with viscoelastic rebound (red lines) calculated for a Maxwell rheology of viscosity 5×10^{18} Pa·s, at KARA, CNCL, QUAR, WAKA, WAIM and MTJO sites (Option 4).

cGPS displacement observations (blue symbols) and predicted coupled afterslip with viscoelastic rebound (red lines) calculated for a Maxwell rheology of viscosity 5×10^{18} Pa·s, at HOKI, MQZG, LKTA and KAIK sites (Option 4).

Figure 3.13: a) Horizontal, b) Vertical and c) Regional horizontal components of accumulated displacement due to coupled viscoelastic rebound (viscosity 5×10^{18} Pa·s) and time-dependent afterslip 400 days after Dusky Sound earthquake (Option 4). Corresponding time series of displacements are shown as red in Figure 3.12.

3.4 Discussion

3.4.1 Alternative viscoelastic models

We attempt to elucidate the postseismic signal following the Dusky Sound earthquake by examining viscoelastic responses to the earthquake for different rheologies. The examined values of viscosity, range from 10^{17} to 10^{20} Pa·s, for the lower crust and the mantle in this chapter are consistent with values assumed elsewhere, from very low ($\leq 10^{17}$ Pa·s; e.g. Vladimirova et al., 2011; Pollitz et al., 1998a; Ryder et al., 2010a) to very high viscosity ($\geq 10^{20}$ Pa·s; e.g. Vladimirova et al., 2011; Pollitz et al., 1998a;

Table 3.1: Root Mean Square Errors of viscoelastic models for Maxwell rheologies with viscosity values of 10^{17} , 10^{18} , 5×10^{18} , 10^{19} , 5×10^{19} and 10^{20} and a Burger’s body with transient and steady-state viscosity values of 10^{17} and 10^{20} Pa·s at continuous GPS sites.

	No afterslip						
Viscosity	10^{17}	$10^{17} - 10^{20}$	10^{18}	5×10^{18}	10^{19}	5×10^{19}	10^{20}
RMSE	25.0	14.7	14.8	15.2	15.3	15.4	15.4

Ryder et al., 2010a).

The nearest site to the epicentral area is PYGR, which exhibited westward and southward postseismic motion. The computed viscoelastic response involves westward and northward motion; in other words, we are unable to achieve reasonable fit with the observed north component of the measured displacement. Neither varying the Maxwell viscosity from 10^{17} to 10^{20} Pa·s nor altering the Burger’s transient viscosity of 10^{17} to steady-state viscosity of 10^{20} Pa·s can account for the southward displacement at PYGR (Figure 3.6).

The viscoelastic response for a very low viscosity of 10^{17} Pa·s lags behind the observed east component at the same site, PYGR. Further from the epicentral area the computed viscoelastic response decays more rapidly than the observed east and north components. Accordingly, we conclude that the observed signal cannot be accounted for in terms of viscoelastic rebound only with the range of viscosities from 10^{17} to 10^{20} Pa·s. RMSEs are calculated for all viscoelastic models in Table 3.1. The two timescales obtained by a Burger’s body allow us to represent or approximate transient processes (e.g., fault slip) that are not possible with a single viscosity. Consequently, we can see the bi-viscous Burger’s body provides better fit with observations compared to other viscoelastic models (RMSE=14.7 mm).

To summarise, the key findings of alternative viscoelastic models neglecting afterslip are as follows:

1. None of the viscoelastic models accounts for the southward motion observed in the epicentral area;
2. The response of a bi-viscous (Burger’s) body predicts the observations more accurately than that of a simple Maxwell rheology with different viscosity values (RMSE error 14.7 mm); and
3. The simple Maxwell rheology with the lowest viscosity of 10^{17} Pa·s deviates significantly from the cGPS measurements with the RMSE of ~ 25.0 mm.

3.4.2 Coupled viscoelastic rebound and afterslip

The iterative approach for coupling the viscoelastic rebound and time-dependent inversion after achieving reasonable fit with campaign and continuous GPS has been followed. As we described in Section 3.3, using Options 1-4, different boundary constraints have been imposed within the PCAIM-inversion method. The main patches using Options 1-4 are located on either sides of the coseismic patch. The unconstrained slip boundary in Option 1 estimates spurious slip on the edges of the extended fault surface. The estimated slip on the off-shore southern side of the fault surface is not well constrained with observations due to the lack of GPS and InSAR observations in the region. We also infer non-uniform rake angles, inconsistent with the average coseismic rake by Beavan et al. (2010a) and the expected rake of the plate motion on the plate interface (DeMets et al., 2010; Beavan et al., 2002), on the edges of the fault surface. Option 2 with a RMSE value of 1.9 mm presented a relatively poor estimation of the postseismic displacement model. The poor accuracy of Option 2 indicates that slip reaches the top edge of the fault surface. Although slip is allowed on the bottom edge of the fault surface, Option 3 predicts insignificant slip at the bottom extension of the fault surface. Thus Option 4 is selected as the final preferred postseismic slip model in an elastic-viscoelastic medium with viscosities of 5×10^{18} – 10^{20} Pa.s.

The grey stars in Figure 3.10 represent sites with inaccurate postseismic measurements. We suspect the higher uncertainties at these sites stem from uncertainties in the estimation of secular motion and removal from the total displacement. At these locations, the misfit of the model with campaign measurements, in spite of satisfactory fits to the cGPS data is explained as a combination of

1. Uncertainties in the coseismic model used to remove the coseismic part from the total displacement. Early postseismic measurements are contaminated by coseismic deformation.
2. Uncertainties in the estimation of secular motion at several sites.

3.4.3 Stress and strain changes in Canterbury

We can address the question of whether or not the Canterbury earthquake sequence of 2010–2012 was advanced by the Dusky Sound earthquake of 2009 by better understanding the mechanisms of postseismic displacement. We consider what the cumulative stress changes produced by the Dusky Sound earthquake were at the hypocentre

of the Darfield earthquake, which originated on the reverse Charing Cross Fault, and at the centre of the Greendale Central Fault on which most slip subsequently occurred (Beavan et al., 2010b).

The geometry and the average rake angle of the reverse Charing Cross Fault and the right-lateral Greendale Central Fault were determined by Beavan et al. (2012b). We calculate the cumulative Coulomb stress changes produced by coseismic and postseismic displacement 400 days after the Dusky Sound earthquake on each of these faults at a depth of 4 km. We assume an isotropic poroelastic rheology for the frictional model (Harris, 1998) with Skempton's and frictional coefficients of 0.5 and 0.6, respectively.

On the Charing Cross Fault, striking 035° and dipping 070° south-east, the Coulomb stress changes for both postseismic (Figure 3.14a) and accumulated coseismic and postseismic processes (Figure 3.14b) are negative. In contrast, on the Greendale Central Fault, which strikes 086° and dips 080° southwards, the Coulomb stress changes induced by postseismic mechanisms alone (Figure 3.14c) or by coseismic and postseismic mechanisms together (Figure 3.14d) after 400 days show positive changes of the order of a few kilopascals.

Our analysis indicates that the Coulomb stress changes due to the Dusky Sound earthquake are consistent with further ruptures on the Greendale Fault in the central South Island, in the Canterbury region but of a very small order. However, the Darfield rupture initiated on the Charing Cross Fault (Holden et al., 2011; Gledhill et al., 2010, 2011) on which negative Coulomb stress changes are estimated due to the Dusky Sound earthquake.

Another way of considering the effects of the Dusky Sound earthquake on the Canterbury region is to examine the pattern of cumulative horizontal strain (Figure 3.15). The cumulative postseismic principal contractional strain and azimuth of maximum horizontal contraction can be calculated from cGPS observations and compared with those inferred from the numerical model. Monte Carlo bootstrapping is used to estimate uncertainties in the strain parameters. Figure 3.15 illustrates the strain ellipses for three different areas in the southern South Island. The ellipses are annotated with the 95% confidence intervals for the principal strains and axes. The results shown in blue are computed from data recorded at all 16 cGPS sites in Figure 4.6. The results and corresponding stations illustrated in green and red represent smaller areas.

Using all 16 cGPS sites yields spatially averaged estimates for the cumulative principal strain components and horizontal contraction azimuth, with values of $[8.3 \text{ to } 13] \times 10^{-8}$

Figure 3.14: a) The receiver fault is set to the geometry of the Charing Cross Fault in order to calculate the cumulative Coulomb stress changes induced by the coupled viscoelastic rebound with a viscosity of 5×10^{18} Pa·s and time-dependent afterslip, b) The receiver fault is set to the geometry of the Charing Cross Fault in order to calculate the cumulative Coulomb stress changes induced by the coupled model added to the coseismic stress changes, c) The receiver fault is set to the geometry of the Greendale Fault in order to calculate the cumulative Coulomb stress changes induced by the coupled model and d) The receiver fault is set to the geometry of the Greendale Fault in order to calculate the cumulative Coulomb stress changes induced by the coupled model added to the coseismic stress changes. The isotropic poroelastic rheology for the frictional model is assumed with Skempton's and frictional coefficients of 0.5 and 0.6, respectively. For Greendale Fault, strike 086° and dip 080° southward, for Charing Cross Fault, strike 035° and dip 070° south-east are considered. All calculations are performed within 400 days after the Dusky Sound earthquake at a depth of 4 km.

(mean 9.4×10^{-8}), $[-8.6 \text{ to } -0.5] \times 10^{-8}$ (mean -3.0×10^{-8}) and $[111^\circ \text{ to } 157^\circ]$ (mean 143°), respectively, 400 days after the Dusky Sound earthquake. The corresponding values for the green and red subsets are $[-18.1 \text{ to } 21.9] \times 10^{-8}$ (mean 15.4×10^{-8}), $[-14.2 \text{ to } +11.2] \times 10^{-8}$ (mean -4.4×10^{-8}) and $[110^\circ \text{ to } 167^\circ]$ (mean 144°) (green, southern) and $[-1.2 \text{ to } 1.8] \times 10^{-8}$ (mean 1.2×10^{-8}), $[-1.0 \text{ to } +0.9] \times 10^{-8}$ (mean -0.4×10^{-8}) and $[79^\circ \text{ to } 125^\circ]$ (mean 106° ; red, northern; extension is positive and contraction is negative). Other calculations, not shown, demonstrate that a displacement gradient across the South Island from north-west to south-east contributes to the cGPS strain estimates. The red strain set includes more sites spanning the north-west to south-east direction as shown in Figure 3.15, and the effect of the displacement gradient in this direction is to rotate the red strain ellipse anticlockwise with respect to the green one.

Figure 3.15: Ellipses for strain accumulated 400 days after the Dusky Sound earthquake shown for three strain sets. Blue indicates the strain ellipse for all 16 cGPS observations shown in Figure 4.6. Second and third strain sets and the calculated strain ellipses are illustrated in green and red. The smaller the number of cGPS sites used, the larger the uncertainties that are estimated from Monte Carlo bootstrapping. PHC: Principal Horizontal Contraction azimuth.

Figure 3.16 illustrates the strains calculated using the numerical model of postseismic displacement following the Dusky Sound earthquake, taking into account the coseismic displacement. Maximum extension directions in blue and contraction directions in red are plotted on a background of the second invariant and, for clarity, we only show results in the south-eastern South Island and Canterbury. The upper panel (Figure 3.16a) shows the coseismic strain and the lower panel (Figure 3.16b) shows the sum of coseismic and postseismic strains over a 400 day period.

The postseismic model does not incorporate secular strain in the South Island or any other sources of strain associated, for instance, with the Alpine Fault or other components of the plate boundary. Therefore, the results are not directly comparable with the strains illustrated in Figure 3.15. Nevertheless, some conclusions can be drawn about the observed and modelled distributions of strain. Focusing on the Canterbury region, we note that the modelled direction of maximum contraction is $\sim 151^\circ$ similar to the blue and green cGPS estimates (mean 144°) but approximately 45° clockwise around the red cGPS result for both coseismic and the sum of coseismic and postseismic strains. Principal strains $\sim 2.4 \times 10^{-8}$ and $\sim -0.3 \times 10^{-8}$ are calculated near the Darfield earthquake epicentre from the coupled time-dependent model within 400 days of the Dusky Sound earthquake, taking into account the coseismic strain. Principal strains of -6.9×10^{-8} and 2.7×10^{-8} and principal horizontal contraction azimuth, 109° are obtained from the average regional strain rate tensor (annual rate). We calculate

Figure 3.16: a) Coseismic b) The sum of coseismic and cumulative postseismic principal strain components, blue arrows pointing outwards for extension and red arrows pointing inwards for contraction and the color contours show the second invariant of strain (Option 4).

the average regional strain rate tensor from the secular velocities of four sites in the box shown in Figure 3.16. The sum of coseismic and postseismic principal extensional and contractional strain rates are approximately 80% and 4% of the equivalent secular rates in the Canterbury region, respectively. However, the principal horizontal contraction azimuth of the sum of coseismic and postseismic strain model is oriented $\sim 42^\circ$ clockwise from the secular motion contraction axis azimuth. Therefore, even if other sources of strain associated, for instance, with the Alpine Fault or other components of the plate boundary can rotate the modelled contraction axis favourably towards

S_{Hmax} , of $116^\circ \pm 18$ in the Canterbury region (Holt et al., 2013; Townend et al., 2012), the effect of the Dusky Sound earthquake and the associated postseismic can be only 4% of the secular motion.

3.5 Conclusions

In this study, we examine different postseismic processes following the Dusky Sound earthquake and consider the farfield effects of this event on the subsequent Darfield earthquake. The iterative method described here reveals: 1) afterslip occurring at depths between 0-80 km on both sides of the coseismic slip patch evolving logarithmically in time; and 2) a viscosity higher than 10^{18} Pa·s for the viscoelastic half-space underlying the elastic layer with a thickness of 45 km.

In order to investigate the effect of the Dusky sound earthquake postseismic responses, we compute Coulomb stress and strain models for both postseismic and the sum of postseismic and coseismic changes. Lateral expansion of the region of positive coseismic Coulomb stresses due to the postseismic mechanisms, following the Dusky Sound earthquake on the Greendale Fault, is inferred from Figure 3.14b and c. In addition, contractional and extensional strain arrows (inferred from Figure 3.16a to b) become larger with considering the postseismic strain model. Therefore we infer that the Dusky Sound earthquake postseismic responses enhanced the coseismic Coulomb stress and strain models significantly. However, the earthquake and the associated postseismic mechanisms do not seem to have directly triggered the Darfield earthquake, based on 1) the negative Coulomb stress changes computed for the Charing Cross Fault on which the Darfield earthquake nucleated; 2) positive but very small Coulomb stress changes on the Greendale Fault; and 3) the magnitude of the principal strain rates, 4% of the regional principal contractional strain rate, and contraction azimuth, 42° clockwise around the regional contraction axis.

Chapter 4

Coupling time-dependent afterslip with viscoelastic relaxation using analytical and numerical Green's functions

Abstract

Postseismic displacement following large earthquakes is often described in terms of either time-varying afterslip on the coseismically ruptured surface or its extension, or viscoelastic relaxation. Here, we describe an inversion framework with which to estimate afterslip while taking account of viscoelastic effects. Estimating time-dependent afterslip is an ill-posed problem unless steps are taken to avoid high-frequency components of displacement affecting the calculated slip, and this requires regularisation in both space and time. Conventionally, the problem of computing the slip distribution in space is regularised using Laplacian smoothing of the fault slip. In order to regularise the ill-posed problem in time, we adapt the method of Kositsky and Avouac (2010) to accommodate viscoelastic media. Ground displacement in response to a dislocation source is a linear function of slip for an elastic medium, meaning that Singular Value Decomposition (SVD) of the observed displacement time series yields the temporal evolution of slip. The problem remains linear for a linear viscoelastic medium but in this case the Green's functions are time-dependent. We estimate Green's functions representing the response to both afterslip and viscoelastic relaxation using the finite element code PyLith or the semi-analytical solution of Fukahata and Matsu'ura (2006). With this approach, we decompose the time-dependent Green's functions and the ground displacement into space and time functions and use these to represent the evolution of slip. We apply this method to the postseismic ground displacement observed following the $M_W 7.8$ Dusky Sound (New Zealand) earthquake of 15 July 2009 and to two numerical and analytical synthetic examples.

4.1 Introduction

Afterslip in the upper/lower crust on the mainshock fault surface (e.g. Savage and Church, 1974; Johnson et al., 2012; Wen et al., 2012a; Reddy et al., 2013) and viscoelastic relaxation in the lower crust/upper mantle (e.g. Elsasser, 1967; Savage and Prescott, 1978; Pollitz et al., 2000; Pollitz, 2003b; Rousset et al., 2012b) are inferred to be the two most predominant postseismic mechanisms following the subduction earthquakes.

Afterslip models often explain the transient displacement following large earthquakes without taking viscoelastic effects into consideration (e.g. Hsu et al., 2002; Miyazaki et al., 2004; Hsu et al., 2006). Time-dependent inversion of continuous global positioning system (GPS) observations yields estimates of the transient afterslip associated with postseismic periods. Such inversion techniques often approximate the slip history in an elastic medium with its corresponding spatial and temporal basis functions and use additional hyperparameters to regularise the problem in both space and time (Segall and Matthews, 1997; Fukuda et al., 2004, 2008).

Several studies have demonstrated that viscoelastic responses are required to match GPS time series data in addition to afterslip (e.g. Azúa et al., 2002; Sheu and Shieh, 2004; Chan and Stein, 2009; Johnson et al., 2009; Bruhat et al., 2011; Diao et al., 2011; Gunawan et al., 2014; Rollins et al., 2015; Sun and Wang, 2015). If the viscoelastic effects of mantle flow are not taken into consideration, very long-term afterslip is required to explain persistent postseismic displacements (e.g. Sawai et al., 2004). However, such prolonged afterslip is incompatible with the characteristic afterslip time-scales observed, which are typically of the order of a few months to a few years (Pritchard and Simons, 2006; Wang et al., 2012). Furthermore, short-term rapid decay of postseismic displacement is often explained by very low values of viscosity and correspondingly short relaxation times (e.g. Palano et al., 2009). Very short relaxation times can also be interpreted in terms of significant afterslip. Therefore, investigating the relative contributions of afterslip and viscoelastic flow in the mantle is essential.

A forward frictional afterslip model in its rate-strengthening form (e.g. Ruina, 1983; Dieterich, 1994; Lapusta et al., 2000; Rice et al., 2001; Marone et al., 1991; Johnson et al., 2006; Barbot et al., 2009; Barbot and Fialko, 2010a) and corresponding viscoelastic responses form a unified framework that incorporates coupled processes of afterslip and viscoelastic flow (e.g. Perfettini et al., 2005; Hearn et al., 2009; Barbot et al., 2009; Johnson et al., 2009; Barbot and Fialko, 2010a). The rate-strengthening form of the friction theory restricts afterslip to locations outside the coseismic patch at

which stresses increase coseismically (Barbot et al., 2009; Johnson et al., 2009). The other method of estimating afterslip is to invert geodetic observations in which the necessary slip occurs within or outside the main coseismic rupture patches (e.g. Freed, 2007; Kositsky and Avouac, 2010; Evans and Meade, 2012). Accordingly, we need to develop appropriate inversion methods to estimate time-dependent afterslip while also considering viscoelastic responses to coseismic and postseismic stress changes.

Fukahata et al. (2004) developed a framework for geodetic data inversion using viscoelastic Green's functions and taking advantage of Akaike's Bayesian Information Criterion (ABIC). The framework performs the inversion by expanding the slip with respect to basis functions and regularizing the solution using two hyperparameters.

Kositsky and Avouac (2010) demonstrated how to invert geodetic time series without requiring additional hyperparameters using the inverse of the elastic Green's function (G^{-1}) multiplied by a singular value decomposition (SVD) of the ground observations (D). In this approach, referred to as the Principal Component Analysis-based Inversion Method (PCAIM), information about time-dependent slip is obtained directly by decomposing the temporally dense geodetic observations into spatial (U) and temporal (V) functions:

$$D_r = U_r \Lambda_r V_r' \quad (4.1)$$

$$S = G^{-1} U_r (\Lambda_r V_r') \quad (4.2)$$

Here S is a time-dependent slip matrix; r is the number of singular values retained; D_r ($n \times T$ dimensions) is a filtered form of the full displacement matrix D with n observed components at T epochs; U_r ($n \times r$) represents the first r columns of U ; Λ_r ($r \times r$) is formed from the first r rows and columns of Λ , which contains singular values of D ; and V_r ($T \times r$) represents the first r columns of V . The optimal value of r is selected by minimising the misfit between the displacement signal D and the filtered signal D_r .

In this chapter, we adapt the PCAIM method of Kositsky and Avouac (2010) to invert geodetic time series in the case of an elastic-viscoelastic medium. This approach enables us to accommodate either semi-analytical (Fukahata and Matsu'ura, 2006) or hybrid numerical viscoelastic Green's functions (Aagaard et al., 2009, 2013, 2015).

Table 4.1: Table of nomenclature

Symbol	Explanation
n	Number of displacement components
m	Number of slip components
T	Number of epochs
$\vec{d}$	Displacement vector minus coseismic viscoelastic response
$\vec{\tilde{d}}$	Total postseismic displacement vector
$\vec{d}^c$	Displacement vector of coseismic viscoelastic response
D	Temporal displacement matrix
g	Vector of Green's functions
G	Matrix of Green's functions
$\tilde{G}$	Matrix of temporal Green's functions
$\vec{s}$	unit-step slip vector
$\vec{v}$	Slip rate vector
S	Temporal slip matrix
U_X	Decomposed space function of X
Λ_X	Diagonal singular values of X
V_X	Decomposed time function of X
V_X^t	Scaled decomposed time function of X ($\Lambda_X V_X$)
X'	The transpose of a matrix X
$\tilde{V}_v$	Matrix of decomposed viscoelastic time function
$\tilde{V}_e$	Matrix of decomposed elastic time function

4.2 Methods

4.2.1 Mathematical framework

In order to calculate viscoelastic effects, we must first compute appropriate viscoelastic Green's functions. Fukahata and Matsu'ura (2006) used the semi-analytical propagator matrix method to compute viscoelastic Green's functions for a dislocation source embedded in a multi-layered elastic-viscoelastic half-space, taking into consideration the gravity field. It is also possible to compute the equivalent viscoelastic Green's functions numerically, which we do here using the PyLith code (Aagaard et al., 2009, 2013, 2015). Table 4.1 lists the parameters referred to in the analysis below.

Postseismic displacements can be split into elastic and viscoelastic components. At time t , the elastic-viscoelastic displacement $\vec{\tilde{d}}$ in an arbitrary direction is presumed to be expressed by the superposition of responses to unit-step slip ($\vec{s}$) occurring at time $\tau < t$. Accordingly,

$$\vec{\tilde{d}}(t) = \vec{d}^c(t) + \int_0^t G(t - \tau) \vec{v}(\tau) d\tau \quad (4.3)$$

where G is the matrix of elastic-viscoelastic Green's functions and $\vec{v}$ is the afterslip rate. By discretising the integral, we approximate Equation 4.3 as

$$\vec{d}(t_j) \simeq \vec{d}^c(t_j) + \sum_{k=1}^{j-1} G(t_j - t_k) \vec{s}_k \quad (4.4)$$

In both Equations 4.3 and 4.4, the superscript c indicates the viscoelastic displacement in response to coseismic stress changes.

Our approach here is to compute the viscoelastic response to the coseismic event and then subtract this from the observed displacement to leave $\vec{d}$ (the second term in Equations 4.3 and 4.4). Equation 4.4 can then be rewritten as

$$\begin{aligned} \vec{d}_{n \times 1}^1 &= G_{n \times m}^1 \vec{s}_{m \times 1}^1 \\ \vec{d}_{n \times 1}^2 &= G_{n \times m}^2 \vec{s}_{m \times 1}^1 + G_{n \times m}^1 \vec{s}_{m \times 1}^2 \\ \vec{d}_{n \times 1}^3 &= G_{n \times m}^3 \vec{s}_{m \times 1}^1 + G_{n \times m}^2 \vec{s}_{m \times 1}^2 + G_{n \times m}^1 \vec{s}_{m \times 1}^3 \\ &\vdots \\ \vec{d}_{n \times 1}^T &= G_{n \times m}^T \vec{s}_{m \times 1}^1 + G_{n \times m}^{T-1} \vec{s}_{m \times 1}^2 + \dots + G_{n \times m}^1 \vec{s}_{m \times 1}^T \end{aligned} \quad (4.5)$$

where n is the number of displacement components and equals the number of stations multiplied by the number of dimensions (horizontal and vertical), m the number of slip components on the fault plane, and T is the number of epochs. $G_{n \times m}^1$ represents the viscoelastic Green's functions after one epoch, $G_{n \times m}^2$ after two epochs and similarly $G_{n \times m}^T$ after T epochs. Green's functions incorporate both elastic (G_e) and viscoelastic (G_v) responses to slip as

$$G_{n \times m} = G_{v \ n \times m} + G_{e \ n \times m} \quad (4.6)$$

In contrast with the impulsive elastic response, viscoelastic Green's functions are time-dependent. Using SVD of the viscoelastic Green's functions, the time functions can be decoupled from the space functions. We consider the matrix $\tilde{G}_v$ consisting of the viscoelastic Green's functions from epoch 1 to epoch T .

$$\tilde{G}_{v \ (n \times m)T} = \begin{bmatrix} g_{v \ (n \times m)1}^1 & g_{v \ (n \times m)1}^2 & g_{v \ (n \times m)1}^3 & \cdots & g_{v \ (n \times m)1}^T \end{bmatrix} \quad (4.7)$$

Here g_v is the vector form of the viscoelastic Green's functions (G_v) at a particular time. The vector g_v in time is a function of one or more values chosen for viscosity or relaxation time and the rheological models.

In this study, we use reduced SVD (Kawamura and Yamaoka, 2006, 2009), in which it is sufficient to represent $\tilde{G}_v$ at $r = 1$ levels. Assuming more complex geometries, other rheologies and heterogeneity of elastic and viscoelastic materials may impose more than one spatial and temporal mode. Whether or not a reduced number of one ($r = 1$) is sufficient to precisely decompose the Green's functions is a subject of ongoing research.

With this assumption, we can express the viscoelastic Green's function as

$$\begin{aligned}\tilde{G}_v &= U_{\tilde{G}_v \ (n \times m) \times r} \Lambda_{\tilde{G}_v \ r \times r} V'_{\tilde{G}_v \ r \times T} \\ V_{\tilde{G}_v \ r \times T}^{t'} &= \Lambda_{\tilde{G}_v \ r \times r} V'_{\tilde{G}_v \ r \times T}\end{aligned}\quad (4.8)$$

Here $U_{\tilde{G}_v}$ is the matrix of space functions (spatial mode), $\Lambda_{\tilde{G}_v}$ is the diagonal matrix of singular values associated with $\tilde{G}_v$, and $V'_{\tilde{G}_v}$ is the matrix of time functions (temporal modes). We use symbols U , Λ and V to represent space functions, scale values and time functions throughout this chapter.

We now revise Equation 4.5 using Equation 4.8 to obtain the associated elastic displacement $\vec{d}_e$

$$\begin{aligned}\vec{d}_{e \ n \times 1}^1 &= G_e(\vec{s}_{m \times 1}^1) \\ \vec{d}_{e \ n \times 1}^2 &= G_e(\vec{s}_{m \times 1}^1 + \vec{s}_{m \times 1}^2) \\ \vec{d}_{e \ n \times 1}^3 &= G_e(\vec{s}_{m \times 1}^1 + \vec{s}_{m \times 1}^2 + \vec{s}_{m \times 1}^3) \\ &\vdots \\ \vec{d}_{e \ n \times 1}^T &= G_e(\vec{s}_{m \times 1}^1 + \vec{s}_{m \times 1}^2 + \dots + \vec{s}_{m \times 1}^T)\end{aligned}\quad (4.9)$$

and viscoelastic displacement $\vec{d}_v$

$$\begin{aligned}\vec{d}_{v \ n \times 1}^1 &= \sum_{i=1}^r U_{\tilde{G}_v}^i (v_i^1 \vec{s}_{m \times 1}^1) \\ \vec{d}_{v \ n \times 1}^2 &= \sum_{i=1}^r U_{\tilde{G}_v}^i (v_i^2 \vec{s}_{m \times 1}^1 + v_i^1 \vec{s}_{m \times 1}^2) \\ \vec{d}_{v \ n \times 1}^3 &= \sum_{i=1}^r U_{\tilde{G}_v}^i (v_i^3 \vec{s}_{m \times 1}^1 + v_i^2 \vec{s}_{m \times 1}^2 + v_i^1 \vec{s}_{m \times 1}^3) \\ &\vdots \\ \vec{d}_{v \ n \times 1}^T &= \sum_{i=1}^r U_{\tilde{G}_v}^i (v_i^T \vec{s}_{m \times 1}^1 + v_i^{T-1} \vec{s}_{m \times 1}^2 + \dots + v_i^1 \vec{s}_{m \times 1}^T)\end{aligned}\quad (4.10)$$

where v_i^1 is calculated by multiplication of the i^{th} singular value and the decomposed time function of viscoelastic Green's functions at the first epoch, v_i^2 at the second epoch and v_i^T at the T_{th} epoch.

The time-dependent slip matrix (S), elastic (D_e) and viscoelastic (D_v) matrices of displacement can now be expressed as follows

$$S = \begin{bmatrix} \vec{s}^1 & \vec{s}^2 & \dots & \vec{s}^T \end{bmatrix} \quad (4.11)$$

$$D_e = \begin{bmatrix} \vec{d}_e^1 & \vec{d}_e^2 & \dots & \vec{d}_e^T \end{bmatrix} \quad (4.12)$$

$$D_v = \begin{bmatrix} \vec{d}_v^1 & \vec{d}_v^2 & \dots & \vec{d}_v^T \end{bmatrix} \quad (4.13)$$

Accordingly, from equations 4.9 and 4.10, we can express the elastic-viscoelastic displacement in matrix forms as

$$D_e = G_e S \tilde{V}_e \quad (4.14)$$

and

$$D_v = \sum_{i=1}^r U_{\tilde{G}_v}^i S \tilde{V}_v^i \quad (4.15)$$

or after decomposing the slip matrix S , we have

$$D_e = G_e U_S V_S^t \tilde{V}_e \quad (4.16)$$

and

$$D_v = \sum_{i=1}^r U_{\tilde{G}_v}^i U_S V_S^t \tilde{V}_v^i \quad (4.17)$$

Space and time functions of the slip matrix are computed from the elastic displacement matrix (D_e ; Kositsky and Avouac, 2010), namely

$$U_S = (G_e)^{-1} U_e \quad (4.18)$$

$$V_S^t = V_e^t (\tilde{V}_e)^{-1} \quad (4.19)$$

Consequently, the relationships between elastic and viscoelastic space and time functions are as follows

$$U_v^i = U_{\tilde{G}_v}^i (G_e)^{-1} U_e \quad (4.20)$$

$$V_v^{t;i} = V_e^t (\tilde{V}_e)^{-1} (\tilde{V}_v^i) \quad (4.21)$$

where

$$\tilde{V}_e = \begin{bmatrix} 1 & 1 & 1 & \dots\dots\dots & 1 \\ 0 & 1 & 1 & \dots\dots\dots & 1 \\ \vdots & \vdots & \vdots & \ddots & \vdots \\ 0 & 0 & 0 & \dots\dots\dots & 1 \end{bmatrix} \quad (4.22)$$

and

$$\tilde{V}_v = \begin{bmatrix} v_i^1 & v_i^2 & v_i^3 & \dots\dots\dots & v_i^T \\ 0 & v_i^1 & v_i^2 & \dots\dots\dots & v_i^{T-1} \\ \vdots & \vdots & \vdots & \ddots & \vdots \\ 0 & 0 & 0 & \dots\dots\dots & v_i^1 \end{bmatrix} \quad (4.23)$$

The overall displacement is the sum of both displacement components, elastic and

viscoelastic:

$$D = D_e + D_v \quad (4.24)$$

After rearrangement

$$D_v = \sum_{i=1}^r U_{\tilde{G}_v}^i (G_e)^{-1} D_e (\tilde{V}_e)^{-1} (\tilde{V}_v^i) \quad (4.25)$$

such that

$$D = D_e + \sum_{i=1}^r U_{\tilde{G}_v}^i (G_e)^{-1} D_e (\tilde{V}_e)^{-1} (\tilde{V}_v^i) \quad (4.26)$$

where $()^{-1}$ denotes the regularised inverse of a matrix.

With further rearrangement, Equation 4.26 can be formulated as a Sylvester equation (Sylvester, 1884) of the form $C = AX + XB$, for which a unique solution exists if and only if A and B have disjoint eigenvalues (Bartels and Stewart, 1972). We simplify Equation 4.26 by assuming $r = 1$ for in order to decouple the viscoelastic Green's functions into spatial and temporal modes. Equation 4.27 is expressable in terms of spatial and temporal modes of elastic displacement.

$$D = U_e V_e^t + U_{\tilde{G}_v} (G_e)^{-1} U_e V_e^t (\tilde{V}_e)^{-1} (\tilde{V}_v) \quad (4.27)$$

Keeping variable U_e constant

$$A = U_e; \quad B = U_{\tilde{G}_v} (G_e)^{-1} U_e \equiv U_{\tilde{G}_v} U_S; \quad C = (\tilde{V}_e)^{-1} (\tilde{V}_v) \quad (4.28)$$

thus

$$D = AV_e^t + BV_e^t C \quad (4.29)$$

4.2.2 Iterative Decoupling of Afterslip and Viscoelastic relaxation (IDAV)

We estimate the elastic spatial function (U_e) in Equation 4.27 from decomposition of the observed displacement signal, and treat this parameter as unchanging in our inversion. This assumption reduces the task of solving the Sylvester equation to one of iteratively updating the temporal elastic time function V_e . Methods of solving the Sylvester equation in which U_e and V_e are estimated simultaneously will be left for further research.

In order to regularise Equation 4.29, we take advantage of Laplacian smoothing in space and decomposition into separate space and time functions. Solving Equation 4.29 yields

a solution for the time functions of elastic displacement.

The overall method of solving Equation 4.29 is illustrated in Figure 4.1. The expected viscoelastic response to coseismic stress changes can be calculated using known (or assumed) viscoelastic properties, specifically viscosity and the thickness of the elastic layer in a half-space model, in conjunction with the coseismic slip distribution. We start by subtracting this response and centering the displacement time series about zero. Thereafter, the remainder encompasses afterslip and the associated viscoelastic response. We apply SVD to the remaining signal in order to extract the elastic part of the signal from the total displacement. Regular (unconstrained) decomposition can over- or underestimate the elastic part of the signal and disrupt convergence of the proposed method.

Equation 4.17 implies that the viscoelastic displacement in time is a function of time-dependent afterslip (V_S^t) and time-dependent viscoelastic Green's functions ($\widetilde{V}_v^i$). Time-dependent Green's functions are calculated based on the geometry of the region and the fault surface, viscosity and the rheology of the viscoelastic medium. Green's functions vary by changing the geometry of the problem and/or assuming multiple layers with different elastic and viscoelastic properties. Time-dependent afterslip is yet to be determined, meaning that the time functions of viscoelastic displacement are unknown. However, the domain of the viscoelastic time functions spans all possibilities of viscoelastic response to slip with relaxation times between zero ($\tau = 0$) and infinity ($\tau = \infty$).

The faster the afterslip decays, the closer the viscoelastic displacement time functions become to the end-member $\tau = 0$. The temporal mode of the end-member $\tau = 0$ corresponds with the temporal mode of the viscoelastic Green's functions ($\widetilde{V}_v^i$). Accordingly, such a temporal mode of viscoelastic response can be imposed as one component of the time functions in the decomposition method, if unconstrained decomposition does not apply. We approximate the space functions of elastic displacement (U_e) with the result of constrained decomposition by imposing $\widetilde{V}_v^i$ ($\equiv$ end-member $\tau = 0$) or the first component of the unconstrained decomposition. The next step in the calculation is to remove the remaining viscoelastic response from the estimated elastic component and to add it back to the viscoelastic spatial and temporal functions.

To summarize, we solve Equation 4.29 with an iterative block linear method (Miller, 1988) for time functions of elastic displacement (V_e). We invert the elastic space functions (U_e) using Laplacian smoothing. We keep variable U_e constant and therefore iteratively we update the variable V_e to find a solution for Equation 4.29, which also corresponds to Equation 4.26. The initial elastic time functions are set to V_e obtained

Figure 4.1: Methodology used to obtain unique solutions for the elastic-viscoelastic Sylvester equations 4.26 and 4.29. In the flowchart we show how to decouple elastic from viscoelastic time functions.

from decomposition. Thereafter we solve for an equation in a form of $AV_e^t(k+1) = Z(k)$ where $Z(k) = D - BV_e^t(k)C$ at iteration k . We then update $Z(k+1)$ using the updated $V_e^t(k+1)$. The algorithm terminates once $\text{norm}[V_e^t(k+1) - V_e^t(k)] < \text{tolerance}$ value and the required correction to the elastic time functions are performed (Figure 4.1). We call this framework of solving the elastic-viscoelastic equation “Iterative Decoupling of Afterslip and Viscoelastic relaxation (IDAV)” method.

4.3 Simulating postseismic displacement using analytical Green's functions

To demonstrate the utility of this method, we first simulate deformation in the case of an elastic layer with a thickness of 35 km overlying a viscoelastic half-space with a viscosity value of 10^{18} Pa·s and a constant shear modulus of 30 GPa. In this simulation, exponentially decaying afterslip occurs within two patches with a maximum accumulated slip of 2.5 m down-dip of the coseismic fault plane, striking 023° and dipping between 013° and 050° towards the south-east. Displacements due to afterslip (red; $\vec{d}_e$), the viscoelastic response to afterslip (green; $\vec{d}_v$) and the viscoelastic response to imposed coseismic rupture (blue; $\vec{d}^c$) at each of 40 sites are computed using elastic (Okada, 1985) and viscoelastic Green's functions (Fukahata and Matsu'ura, 2006; Figure 4.2).

We infer from the simulated data for this geometry and the location of sites that the viscoelastic responses to coseismic rupture and to afterslip can occur in different directions. The arrows illustrate the accumulated displacement within 400 days of the simulated rupture. In the time series generated at three representative sites, the total displacement is depicted in black. Note that the associated names do not correspond to real cGPS sites.

We next invert the calculated total displacement ($\vec{d}$) to reproduce the simulated afterslip model. In Figure 4.3a, we decompose the total displacement into two components using the constrained (solid lines) and unconstrained (dashed lines) decomposition methods (see Section 4.1; Equation 4.1). Unconstrained decomposition yields correlated components and consequently underestimates the first component (blue). In order to avoid this, we decompose the signal with a viscoelastic time function corresponding to slip with relaxation time of zero ($\tau = 0$; Figure 4.3b, red). The viscoelastic time function equivalent to constant slip with infinite relaxation time is illustrated in blue in Figure 4.3b as the upper bound of the viscoelastic temporal mode ($\tau = \infty$).

The iterative method presented in Figure 4.1 is followed until convergence, which in this case occurs after three iterations when $\text{norm}[V_e^t(k+1) - V_e^t(k)] < 1$ mm. In Figure 4.3c and d, the initial and final elastic and viscoelastic time functions are presented in blue and red, respectively. The estimated slip after convergence is illustrated in Figure 4.4a in comparison to the originally imposed slip in Figure 4.4b, within 400 days of the simulated rupture. We also illustrate horizontal and vertical fits (blue) to the simulated elastic displacements (red) between one and 400 days of the simulated rupture, in

Figure 4.2: The accumulated displacement within 400 days of a simulated rupture due to afterslip (red), the viscoelastic response to afterslip (green) and the viscoelastic response to coseismic slip (blue). Corners: Time series of different components of simulated displacement illustrated for three sites. Black lines show the total displacement time series.

Figure 4.3: a) Time functions resulting from constrained (solid lines) and unconstrained (dashed lines) decomposition methods. b) Domain of viscoelastic time functions from a response to afterslip with relaxation time of zero (red) and infinity (blue). c) Convergence of elastic time functions from blue to red. d) Convergence of viscoelastic time functions from blue to red.

Figures 4.5a and b.

4.4 Postseismic displacement following the M_W 7.8 Dusky Sound earthquake, Fiordland, New Zealand

Postseismic displacement following the Dusky Sound earthquake and associated Coulomb stress changes and strain accumulation in the time period between the 15 July 2009 Dusky Sound and 4 September 2010 Darfield earthquakes have been calculated in Chapter 3. We illustrate focal mechanisms of the Dusky Sound and Darfield earthquakes, the region of study, the locations of cGPS sites and cGPS displacement times series at the two sites nearest to the epicentral area in Figure 4.6.

In Chapter 3, we explained the postseismic displacement observed with continuous GPS and early campaign GPS data acquired during August 2009 and Differential Interfero-

Figure 4.4: a) Estimated slip accumulated 400 days after coseismic rupture b) Simulated slip used as input in order to produce the displacement time series c) Difference between estimated and simulated slip.

Figure 4.5: a) Horizontal and b) Vertical fits (blue) to the simulated elastic displacement (red) between day one and day 400 following the simulated rupture.

Figure 4.6: a) Focal mechanisms of recent earthquakes, namely the 15 July 2009 Dusky Sound, 4 September 2010 Darfield and 22 February 2011 Christchurch earthquakes, which were widely felt in the South Island, New Zealand. Locations of cGPS sites are shown in red triangles. b) Time series of cGPS displacement following the Dusky Sound earthquake at two nearest cGPS sites to the epicentral area.

Figure 4.7: Viscoelastic Green's functions for a fault patch at an arbitrary point in time, with different values of viscosity. The cumulative displacement over one day for a Maxwell material with a viscosity of 10^{18} Pa·s is twice that of a viscosity of 2×10^{18} Pa·s

metric Synthetic Aperture Radar (DInSAR) observations in terms of coupled afterslip and viscoelastic relaxation. In this model, two patches on the fault surface (Beavan et al., 2010a) at depths of 0–80 km accommodated afterslip with a maximum slip of ~ 1 m decaying logarithmically in the period between the Dusky Sound and Darfield earthquakes. The fault surface is embedded in an elastic layer with a thickness of 45 km overlying a Maxwell viscoelastic half-space with viscosities of 5×10^{18} – 10^{20} Pa·s.

Here, lower values of viscosity than those considered in Chapter 3, between 2×10^{18} and 4×10^{18} Pa·s for the Maxwell half-space, are examined using the IDAV method introduced in Section 4.2. Green's functions are calculated for the fault patches assuming a viscosity of 10^{18} Pa·s and using the semi-analytical propagator matrix methods of Fukahata and Matsu'ura (2006). The cumulative displacement over one day for a Maxwell material with a viscosity of 10^{18} Pa·s is twice that for a material with a viscosity of 2×10^{18} Pa·s (Figure 4.7). Therefore, in order to reduce the computational time for higher-viscosity models, the Green's functions are computed by scaling the functions associated with a viscosity of 10^{18} Pa·s.

In this case, unconstrained decomposition yields two uncorrelated components of displacement. Equation 4.29 converges after a few iterations with a tolerance of 1 mm, for three values of viscosity. Elastic and viscoelastic time functions in the first and last iterations are presented in Figure 4.8a-c, from initial blue and cyan to final red and magenta, respectively. The total estimated afterslip models 400 days after the Dusky

Table 4.2: Root Mean Square Errors, moment magnitudes and maximum slip values of coupled models for a Maxwell rheology with viscosity values of 2×10^{18} , 3×10^{18} and 4×10^{18} Pa.s.

Viscosity (Pa · s)	Coupled models		
	2×10^{18}	3×10^{18}	4×10^{18}
RMSE (mm)	1.9	1.6	1.5
Moment ($\times 10^{20}$ N · m)	2.8	3.4	3.8
Magnitude (M_w)	7.6	7.7	7.72
Max slip (m)	0.86	1	1.1

Sound earthquake are illustrated in Figure 4.8d-f for different viscosity values. Table 4.2 presents Root Mean Square Error (RMSE) values, the cumulative moment magnitude and the maximum values of slip for each coupled model with different viscosities. The higher the viscosity value, the larger the estimated afterslip. We can describe the difference between afterslip models for different values of viscosity with the effects of viscoelastic relaxation to partially explain the observed displacement signal.

The supplementary file contains the displacement time series estimated from the coupled model with a viscosity of 2×10^{18} in red and observed time series in blue at 16 continuous GPS sites.

4.5 Simulating postseismic displacement using numerical Green’s functions

We can also use the IDAV method of describing coupled viscoelastic relaxation and afterslip to simulate more complex faulting scenarios. We numerically simulate post-seismic displacement following a M_w 8.4 subduction earthquake using the Finite Element Method (FEM) implemented in PyLith (Aagaard et al., 2009, 2013, 2015). The geometry for numerical simulation of the Green’s functions consists of an elastic layer with a thickness of 50 km overlying a viscoelastic half-space (Figure 4.9). We also take into account an elastic slab with a thickness of 50 km. We assume uniform elastic properties with a density of 2.5 kg/m^3 , and P- and S-wave velocities of 5000 and 3000 m/s. We assign viscosities of 10^{18} and 5×10^{18} Pa.s to the continental mantle and oceanic mantle, respectively (Figure 4.9). The north-south-striking fault, dipping 24° westward, extends to the depth of 80 km.

We mesh the geometry with a uniform resolution of 25 km using the software package Trelis (<http://www.csimsoft.com/help/trelishelp.htm>). We calculate Green’s

Figure 4.8: Convergence of elastic and viscoelastic time functions in first and last iterations from initial blue and cyan to final red and magenta, respectively a) Viscosity value of 2×10^{18} Pa.s b) Viscosity value of 3×10^{18} Pa.s c) Viscosity value of 4×10^{18} Pa.s. Accumulated slip 400 days after the Dusky Sound earthquake d) Viscosity value of 2×10^{18} Pa.s e) Viscosity value of 3×10^{18} Pa.s and f) Viscosity value of 4×10^{18} Pa.s. The temporal accumulated slip in magnitude differs when we increase the viscosity from 2×10^{18} to 4×10^{18} Pa.s.

Figure 4.9: Viscoelastic half-space considering the slab geometry. Viscosities of 10^{18} and 5×10^{18} Pa·s are assigned to the continental and oceanic mantle, respectively. The north-south striking fault, dipping 24° westward, extends to the depth of 80 km and the elastic slab has a thickness of 50 km.

functions at each vertex, on a simulated fault, associated with 55 ground control points, as responses to unit strike- and dip-slip. Figure 4.10 illustrates unit-step slip at a vertex on the fault surface, which yields the Green's functions shown in the graph associated with a ground point (A1).

We compute the elastic (red) and viscoelastic (blue and green) components of displacement due to simulated time-dependent afterslip (Figure 4.11). Viscoelastic responses to the simulated rupture and afterslip are shown in blue and green, respectively and the total displacement in black. In Figure 4.11, we compare the elastic and viscoelastic components of displacement. We also illustrate time series of elastic component (red), viscoelastic responses to the simulated rupture (blue) and afterslip (green) in Figure 4.11.

We decompose the time-dependent displacement signals into spatial and time functions imposing the temporal mode of the Green's functions. The elastic and viscoelastic time functions converge to the final solutions (red and magenta, respectively) after five iterations with a threshold value of 1 mm (Figure 4.12). Initial elastic and viscoelastic time functions obtained from the constrained decomposition are shown in blue and cyan, respectively.

We show the simulated afterslip model in Figure 4.13a and the estimated afterslip model after inverting the displacement time series in Figures 4.13b. The discrepancy between simulated and estimated afterslip models is approximately 7% of the slip magnitude.

Figure 4.10: Numerically calculated Green's functions as responses to unit-step slip on a simulated fault. The north-south striking fault, dipping 24° westward, extends to the depth of 80 km and the elastic slab has a thickness of 50 km.

Figure 4.11: The accumulated displacement model of the slab geometry within 1095 days of a simulated subduction earthquake due to afterslip (red), viscoelastic response to afterslip (green) and viscoelastic response to coseismic slip (blue). Corners: Time series of different components of simulated displacement illustrated for three sites.

Figure 4.12: Elastic and viscoelastic time functions in different iterations from initial blue and cyan to final red and magenta, respectively.

Figure 4.13: a) Simulated slip used as input in order to produce the displacement time series. b) Estimated slip from inverting the displacement time series induced by the slab geometry.

4.6 Discussion

We have shown how to formulate the total postseismic displacement corresponding to elastic and viscoelastic components in terms of a Sylvester Equation 4.29 and developed a method of solving this equation in order to estimate each component (IDAV method).

The analytical and numerical simulations of postseismic displacement demonstrate that after large earthquakes, such as the 2009 Dusky Sound event, the viscoelastic response to afterslip can be significant. Within 400 days of the Dusky Sound earthquake, consideration of viscoelastic responses means that the moment magnitude of afterslip changes from M_w 7.6 to 7.7, as the assumed viscosity varies from 2×10^{18} to 4×10^{18} Pa·s. The difference between afterslip models for different viscosity values in this time interval are insignificant. In contrast, the effects of viscoelasticity do affect the longer-term displacement signal. Three years of numerically simulated postseismic displacement following a M_w 8.4 subduction earthquake reveals significant viscoelastic displacement compared to the displacement signal due to afterslip (see Figure 4.11).

In order to formulate the postseismic displacement in terms of separate elastic and viscoelastic responses, we have decomposed the Green's functions corresponding to the three examples considered into one set of spatial and time functions. Although sufficiently accurate for demonstrating the method's applicability, higher levels of Green's function decomposition should be assessed as part of further work.

For numerical simulations, the accuracy of the Green's functions is likely affected by two factors: 1) interpolating the Green's functions from the finite elements of the mesh onto the finite rectangular patches of the fault; 2) the residuals of displacement at the boundaries of the model domain. In an inversion framework for a real case study, a much larger model domain would be required to ensure accuracy. Assigning varying elastic properties to different parts of the model domain, and particularly to the slab, may result in different displacement models associated with the assumed geometry.

In this inversion method, we have assumed a priori knowledge of viscosity, rheology and the geometry of the viscoelastic half-space. Therefore, the validity of the solutions is contingent on the accuracy of these parameters, and in an ideal situation complementary data would be used to estimate the viscoelastic parameters.

4.7 Conclusions

Postseismic displacement consisting of elastic and viscoelastic components can be expressed as a Sylvester equation, providing a novel means of estimating the time dependent slip distribution in an elastic-viscoelastic medium without introducing additional hyperparameters. The temporal modes of the time-dependent slip are directly inferred from the temporal modes of the viscoelastic Green's functions and the observed postseismic displacement. The IDAV method of solving the equation is straightforward in that only one temporal mode is used to decouple the viscoelastic Green's functions. This assumption is consistent with the analytical and numerical simulations of viscoelastic Green's functions for various geometries, and the simple assumptions of rheology and viscosity considered in this chapter.

We have examined the utility of the IDAV method using postseismic displacement observations made following the Dusky Sound earthquake and two simulated earthquakes, analytically and numerically. The results of inverting the postseismic displacements after the Dusky Sound earthquake imply viscosity values equal to or higher than 2×10^{18} Pa·s for the viscoelastic half-space. The inversion method is applicable to more complex faulting scenarios using numerical simulation of the Green's functions, enabling consideration of realistic elastic-slab geometries for subduction earthquakes and lateral heterogeneity of elastic or viscoelastic properties.

Chapter 5

Observations and modelling of postseismic displacement during and following the 2010–2012 Canterbury earthquake sequence

Abstract

The 2010–2012 Canterbury earthquake sequence commenced with the 4 September 2010 M_W 7.1 Darfield and produced several destructive aftershocks, of which the most damaging, the 22 February 2011 M_W 6.2 Christchurch earthquake, caused 185 fatalities and extensive damage to infrastructure. The sequence also produced postseismic deformation signals, which we analyse here using Interferometric Synthetic Aperture Radar (InSAR) and campaign Global Positioning System (GPS) data. Measurements made with ascending and descending acquisitions by the COSMO-SkyMed satellite suggest that poroelastic deformation induced by the Darfield earthquake terminated within one month, whereas longer-term dilatancy recovery continued for at least five months. Afterslip does not appear to have caused significant postseismic deformation after the Darfield earthquake. TerraSAR-X and campaign GPS measurements spanning the eastern end of the Greendale East Fault, which exhibited the greatest slip in the Darfield earthquake, reveal a pronounced deformation gradient and relative line-of-sight motion of 4 mm/yr across the fault following the Christchurch earthquake. This ~ 15 km-long feature coincides with a concentration of aftershocks defining a previously recognised lineament, and we interpret the postseismic signal to represent afterslip on the lineament which is a continuation of the Western Christchurch Fault and subsidiary faults.

5.1 Introduction

The 2010–2012 Canterbury earthquake sequence commenced with the $M_W \sim 7.1$ Darfield earthquake on 4 September 2010; the ensuing aftershock sequence included the $M_W \sim 6.2$ Christchurch earthquake on 22 February 2011, which caused 185 fatalities, and several other damaging aftershocks.

The Darfield and Christchurch earthquakes both occurred on previously unrecognised fault planes (Figure 5.1; Gledhill et al., 2011; Kaiser et al., 2012; Quigley et al., 2012), but subsequent seismological and geodetic studies have revealed complex fault structures. The Darfield earthquake produced a ~ 30 km-long surface rupture extending along the newly named Greendale Fault. Based on campaign Global Positioning System (GPS) measurements and airborne Light Detecting And Ranging (LiDAR) data, Quigley et al. (2012) estimated a static stress drop for the Darfield earthquake of ~ 13 MPa. Regional and teleseismic moment tensor analysis conducted immediately after the Darfield earthquake revealed that reverse faulting occurred at the hypocentre but that most of the slip occurred on an east–west-striking vertical fault as dextral strike-slip motion (Gledhill et al., 2010, 2011). The state of stress in the crust encompassing the Canterbury earthquake sequence has been the focus of several studies (Sibson et al., 2011; Townend et al., 2012; Holt et al., 2013; Herman et al., 2014). The study of the contemporary regional stress field in the Canterbury region shows that the maximum horizontal compressive stress axis has a trend of $\sim 115^\circ$, similar to the orientation seen throughout much of the South Island (Holt et al., 2013; Townend et al., 2012).

Beavan et al. (2010b) obtained a preliminary geodetic source model for the Darfield earthquake based on the analysis of ground displacement measurements made with GPS and Interferometric Synthetic Aperture Radar (InSAR) data. Their source model revealed a complex rupture involving slip on several fault segments. The preferred model required a combination of reverse faulting on the Charing Cross fault, strike-slip faulting on the Greendale Fault, and normal faulting on the north-western segment as well as failure of the Hororata thrust Fault (Figure 5.1b). Using regional seismicity data, Gledhill et al. (2011) later refined the complex source structure in terms of aftershock hypocentres.

Elliott et al. (2012) modelled the geometry of the Darfield earthquake with seven fault segments using InSAR measurements, SAR amplitude offsets, a LiDAR DEM, aerial photogrammetry and satellite imagery along with teleseismic body-wave analysis. Their results agreed with the rupture initiated on a reverse fault and continued onto strike-slip fault segments. Syracuse et al. (2012) observed a NNW–SSE-striking

band of seismicity corresponding to the seventh fault-segment identified from geodetic observations by Elliott et al. (2012).

Herman et al. (2014) calculated best-fitting double couple moment tensors and source depths for earthquakes having M_W larger than 3.7 in order to determine the stress evolution throughout the Canterbury earthquake sequence. The focal mechanisms indicated major right-lateral slip on ENE fault planes and reverse faulting.

Shortly after the Christchurch earthquake, Beavan et al. (2011) inverted GPS and InSAR data to obtain models of the fault geometry and slip distribution. They explained most of the observed ground displacement by coseismic slip on two sub-parallel fault planes. In a subsequent study, Beavan et al. (2012a) revised the original source models for the Darfield and Christchurch earthquakes, and two significant events in June and December 2011. The results of their study reinforced earlier findings regarding the locations and geometries of the faults that had ruptured in each event, but added a left-lateral NNW-striking fault extending from the epicentre to the Darfield source model and two further faults (the strike-slip Western ENE fault and the reverse NNE fault) to the Christchurch source model (Figure 5.1b).

The Christchurch earthquake was notable for the high ground accelerations it produced near the epicentre. Fry et al. (2011) documented peak ground accelerations of 2g and reported that asymmetry in the vertical-component seismograms was consistent with non-linear behaviour of near-surface materials. Kinematic source modelling of the Christchurch earthquake indicated maximum slip of ~ 4 m at a depth of ~ 4 km and a rupture velocity of 2.8 km/s (Holden et al., 2011).

Barnhart et al. (2011) computed static stress changes induced by the Darfield earthquake and concluded that the Christchurch earthquake nucleated in a positive region of static Coulomb stress changes, with a peak of 0.31 MPa. In contrast, the Coulomb stress changes calculated by Beavan et al. (2011) in the vicinity of the Christchurch event are very small, although the region is still located in the positive Coulomb stress change area. The difference between the magnitude of the Coulomb stress changes can be related to the different source models that have been used for the Darfield and Christchurch events in the studies by Beavan et al. (2011) and Barnhart et al. (2011).

Bannister and Gledhill (2012) summarised the evolution of the Canterbury earthquake sequence as well as reverse and strike-slip aftershocks. This and other studies have identified a lineament defined by the concentration of the aftershocks extending from further south of the Greendale East Fault to the Western Fault that ruptured during the Christchurch earthquake (Figure 5.1b; Sibson et al., 2011; Syracuse et al., 2012; Herman

et al., 2014) . Sibson et al. (2011) inferred a step-over, a potential dilational zone, throughout the area between the Greendale East Fault and the lineament. Dilational zones are known as kinematic barriers to arrest the rupture and dissipate the terminated strain energy by slip transfer to the nearby fault segments (Sibson, 1986).

Motagh et al. (2014b) undertook the first study of postseismic displacement following the Darfield earthquake using differential interferometry applied to two months' observations. They constructed a displacement time series based on ascending COSMO-SkyMed (CSK) acquisitions spanning the interval between September 2010 and February 2011. The early postseismic displacement field reveals motions towards the ascending CSK satellite along the Greendale East Fault and away from the satellite along the Charing Cross Fault and within the step-over between the Greendale East and Central Faults.

This study examines the postseismic ground displacement following the Darfield and the smaller but more destructive Christchurch earthquakes. Following the Darfield earthquake and prior to the Christchurch earthquake, we focus specifically on the area extending from near the Hororata Fault to the easternmost end of the Greendale East Fault (Figure 5.1b and c). The area encloses the faults related to the Darfield earthquake. Following the Christchurch earthquake, the area of interest spans the region between the easternmost end of the Greendale East Fault and the urban area of Christchurch. This area encompasses the faults that ruptured during the Christchurch earthquake and those that ruptured subsequently during the 13 June 2011 and 23 December 2011 events (Figure 5.1b).

In previous studies of postseismic displacement following large earthquakes, several processes yielding different spatiotemporal patterns of surface deformation have been proposed.

- Viscoelastic relaxation, transferring stress from the ductile lower crust and upper mantle to the brittle upper crust (e.g. Freed and Lin, 1998, 2001; Rydelek and Sacks, 2001; Freed, 2005; Pollitz et al., 2006a; Freed et al., 2007; Tabrez et al., 2008; He et al., 2011; Decriem and Árnadóttir, 2012);
- Poroelastic rebound, in which instantaneous changes in confining stress induced coseismically cause pore pressure changes which equilibrate postseismically (Peltzer et al., 1996, 1998);
- Rate-strengthening afterslip on the faults that slipped coseismically (Lapusta et al., 2000; Rice et al., 2001; Barbot et al., 2009), as formulated in terms of a rate-and-state-dependent friction relationship (Ruina, 1983; Dieterich, 1994);
- Inverse afterslip model, meaning afterslip occurring in both the upper and lower

crust on the mainshock fault surface (e.g. Savage and Church, 1974; Johnson et al., 2012; Wen et al., 2012a; Reddy et al., 2013); and

- Dilatancy recovery, by which the crust recovers its preseismic strength (Fielding et al., 2009) and which has been suggested to have occurred following the Darfield earthquake by Reyners et al. 2014 (see Section 2.2.5).

The main objective of this chapter is to understand which of the postseismic mechanisms above can best account for ground displacements observed following the Darfield and Christchurch earthquakes. For the Darfield earthquake, we use InSAR observations acquired within one month of the earthquake and approximately five months of displacement time series from October 2010 to February 2011 using the Italian CSK satellite constellation. For the Christchurch event, we use campaign GPS velocities acquired annually in early 2012 to 2015 and InSAR time-series analysis derived from TerraSAR-X (TSX) acquisitions.

5.2 Methodology

5.2.1 Observations

GPS velocities

We analyse campaign GPS data acquired annually between early 2012 to 2015 in the Canterbury region ¹. We solve for station coordinates, satellite orbit and Earth rotation parameters, and estimate atmospheric zenith delay every two hours and three atmospheric gradients per day. We use the IGS08 (International GNSS Service; Schmid et al., 2007) azimuth and elevation-dependent absolute phase centre model with an elevation cutoff angle of 10° for the ground-based antennas and apply the FES2004 (Finite Element Solutions (FES) tidal atlases; Lyard et al., 2006) ocean loading model. The positions are transformed into the ITRF08 (International Terrestrial Reference Frame; version 2008) Pacific plate fixed reference frame (Altamimi et al., 2012), via a six-parameter transformation (translations and rotations). We minimise the residual of 30 reference stations in the IGS network (Herring et al., 2008) and subtract the velocity of site MQZG in three successive years (Figure 5.2).

¹Processed by Dr. Sigrún Hreinsdóttir

Figure 5.1: a) Tectonic setting of the South Island, New Zealand and focal mechanisms of recent earthquakes, namely the 2010 Darfield and 2011 Christchurch earthquakes (continued on the next page).

Interferometric Synthetic Aperture Radar (InSAR) analysis

To map and interpret postseismic displacements following the Darfield earthquake, we use interferograms acquired by the CSK satellite (Figure 5.3a and b) and generated by Motagh et al. (2014b). The interferograms enclose the principal fault segments

Figure 5.1: b) Map of the study area showing the epicentres of the Darfield (Canterbury) earthquake and the Christchurch earthquake (green and yellow stars, respectively). Also shown are the surface traces of faults inferred to have ruptured during the Darfield and Christchurch earthquakes and large aftershocks: A) Greendale Central, B) Greendale NW-SE, C) Greendale East, D) Charing Cross, E) near Hororata, F) NNW from the epicentre, G) NE from step-over Faults related to the Darfield earthquake, H) Western, I) NNE, J) Eastern Christchurch Faults related to the Christchurch earthquake, K) SW-NE, L) NW-SE Faults related to the 13 June 2011 event and M) SW-NE Fault related to the 23 December 2011 event (Beavan et al., 2012a). Green dots indicate the aftershocks following the Darfield and prior to the Christchurch earthquake. Yellow dots indicate aftershocks following the Christchurch earthquake. Red triangles indicate cGPS sites. c) Time intervals related to the ascending CSK and descending TSX interferograms in period A (first interval) and B (second interval). Red stars indicate the Darfield, Christchurch, 13 June and 23 December 2011 earthquakes. Aftershocks are reported by <http://quakesearch.geonet.org.nz/>.

Table 5.1: Incidence angle and temporal baselines of interferograms in Figure 5.3 and 5.5.

Data type	Image date	Incidence angle $^{\circ}$	B_{temp} (days)
CSK ascending	20100906-20100922	22	17
CSK descending	20100913-20100929	47	17
TSX descending	20100907-20100918	40	12

that ruptured during the Darfield earthquake. In addition, we use 12 ascending CSK images to generate a displacement time series between October 2010 and February 2011 (Figure 5.4). We present temporal displacement results for the step-over between the Greendale East and Central Faults. The Small BAseline Subset (SBAS) network comprises 38 interferograms in which pixels with coherence above a threshold of 0.2 are selected. After 3D phase unwrapping, the time series of displacement was estimated using StaMPS (Hooper and Zebker, 2007; Hooper, 2008; Hooper et al., 2010).

An additional post-Darfield interferogram is generated from two descending TSX acquisitions that in conjunction with the descending CSK interferogram help to constrain the sense of motion along the easternmost end of the Greendale East Fault (Figure 5.5). We generate the interferogram using standard techniques implemented in the commercial software SARscape. We do not model either the descending CSK interferogram (Figure 5.3b) or the TSX interferogram directly (Figure 5.5), because of the strong long-wavelength orbital and tropospheric errors (Motagh et al., 2014b). Table 5.1 presents incidence angle and temporal baselines of each interferogram in Figure 5.3 and 5.5.

Following the Christchurch earthquake, we use 66 TSX acquisitions in descending orbits with an incidence angle of 40° and heading angle of 196° acquired between March 2011 and May 2014. After reading and cropping the Single Look Complex (SLC) products, satellite orbits for all images are extracted and Digital Elevation Model (DEM)

Figure 5.2: Arrows represent GPS horizontal velocities in the Pacific plate reference frame relative to the continuous GPS site MQZG for three periods: a) 2012-2013 (blue), b) 2012-2014 (red) and c) 2012-2015 (maroon).

Figure 5.3: a) Unwrapped interferogram acquired from ascending CSK orbits between the 6th and 22nd of September 2010. b) Unwrapped interferogram acquired from descending CSK orbits between the 13th and 29th of September 2010. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

offset corrections applied using the LINZ (Land Information New Zealand) 40 m digital elevation model. All pairs for each dataset are coregistered and resampled and 65 master-slave interferograms are produced using the software package DORIS (Kampes and Usai, 1999). The TSX acquisition from 20 July 2012 was selected as the master image for the postseismic time interval between 2 March 2011 and 22 May 2014 as it yields the minimum temporal and perpendicular baselines for the master-slave pairs. Time series analysis with the default StaMPs parameters is only conducted for Slowly Decorrelating Filtered Phase (SDFP) pixels (Hooper and Zebker, 2007; Hooper, 2008; Hooper et al., 2010). The maximum density of the candidate pixels is 2 percent (per km²). The effect of turbulent atmospheric delay has been mitigated by stacking numerous acquisitions. The stratified atmospheric delay masks the displacement signal over the area on the south-eastern corner of the case study. For this reason we discard measurements from the region with elevations above 150 m in the final velocity calculations.

The mean Line Of Sight (LOS) velocity for data acquired between March 2011 and May 2014 is illustrated in Figure 5.6a: this velocity includes coseismic offsets associated with the June and December 2011 earthquakes (see Appendix D), as well as postseismic displacements. Due to the effects of the June and December 2011 events, we also solve

Figure 5.4: Observed time series of postseismic displacement acquired from ascending CSK acquisitions following the Darfield and prior to the Christchurch earthquakes on the Greendale step-over between October 2010 and February 2011 relative to 14 October 2010. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

for the mean LOS velocity using data acquired between January 2012 and May 2014 only (Figure 5.6b; the standard deviations of the velocity results are presented in Appendix D).

5.2.2 Analytical and numerical models

In this study, we simultaneously invert the ground displacement after subtracting poroelastic effects to estimate the dilatancy recovery and afterslip responses. Methods of forward modelling viscoelastic, poroelastic, and afterslip deformation are reasonably

Figure 5.5: Unwrapped interferogram acquired from descending TSX orbits between the 7th and 18th of September 2010 within one month of the Darfield earthquake. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

standard and are not discussed in detail here.

We model viscoelastic relaxation, poroelastic rebound and frictional afterslip with the numerical code PyLith (Aagaard et al., 2009, 2013, 2015), Fortran codes of Wang and Kumpel (2003) and RELAX (Barbot et al., 2009; Barbot and Fialko, 2010a), respectively.

Simultaneous poroelastic rebound, dilatancy recovery and afterslip responses

To combine poroelastic rebound, dilatancy recovery and afterslip, we first subtract a forward estimate of poroelastic rebound from the observations. We then invert the residual observations to estimate the components of slip (afterslip) and volumetric changes (dilatancy recovery) occurring on assumed faults. Appropriate constraints must be imposed in order to ensure that the sense of motion represented by the resulting afterslip model is consistent with that expected for each fault.

One way of accounting for dilatancy recovery is to assume a regular grid of Mogi point sources on the fault planes that ruptured coseismically (Mogi, 1958). This approach is simplistic in that it does not completely describe the physical process, as Fielding et al. (2009) explained. However, the advantage of assuming the same grid for the Mogi point sources as the slip distribution is that the only parameter we estimate is the volume change at each point source with defined coordinates and depth. In this study, therefore, we simultaneously solve for afterslip consisting of strike- and dip-slip

Figure 5.6: a) Observed mean LOS velocity acquired from TSX acquisitions following the Christchurch earthquake between March 2011 and May 2014. b) Observed mean LOS velocity acquired from TSX acquisitions following the Christchurch earthquake between January 2012 and May 2014. Red colours indicate movement away from the satellite and blue colours indicate movement towards the satellite. Dashed line is micro-earthquake lineament (extension of H). Other faults are described in Figure 5.1b.

components (s_s and s_d) and volume change (Δv) at each point of the defined regular grid.

We can express the observation equation relating the observed ground displacement (d) and the parameters of interest (s) as follows

$$d = Gs + \epsilon = \begin{bmatrix} G_f & G_v \end{bmatrix} \begin{bmatrix} s_s \\ s_d \\ \Delta v \end{bmatrix} + \epsilon \quad (5.1)$$

where G is the combined (augmented) design matrix, G_f and G_v are functions of the dislocation model and the Mogi point source parameters (Okada, 1985; Mogi, 1958),

and ϵ is the residual.

If X , Y and D are the east, north coordinates and depth of the point source with respect to the ground location and $R = \sqrt{X^2 + Y^2 + D^2}$, then the three-dimensional ground displacement induced by a Mogi point source is given by

$$d_v = \frac{\nu - 1}{\pi R^3} \begin{bmatrix} X \\ Y \\ D \end{bmatrix} \Delta v + \epsilon_v = G_v \Delta v + \epsilon_v \quad (5.2)$$

where ν is Poisson's ratio. Since observations are made in the LOS direction, we project Equations 5.1 and 5.2 in the appropriate direction using the heading and incidence angles of the sensors.

In order to solve for the two components of afterslip on the fault planes as well as the volume changes, we use the linear least squares approach to invert the displacement observations. Equation 5.1 is still an ill-posed problem and requires spatial smoothing. During the inversion, the following function is minimised

$$f = \|d - Gs\|^2 + \gamma^{-2} \|\nabla^2 s\|^2 \quad (5.3)$$

∇^2 is the approximation of the Laplacian operator and γ determines the trade-off between misfits and smoothness of parameters.

5.3 Post-Darfield displacement

5.3.1 InSAR observations

Our observations of displacement following the Darfield earthquake exhibit several notable features (Figure 5.3 , 5.4, 5.5 and 5.7):

1. Localised movement away from the satellite of ~ 20 mm within the Greendale step-over at the intersection between fault segments (A) and (C), within approximately one month of the Darfield earthquake, and cumulative movement away from the satellite of ~ 25 mm between October 2010 and February 2011;
2. Movement away from the satellite on the Greendale NW-SE (B), Charing Cross (D), near Hororata (E), NNW from the epicentre (F) and NE from step-over (G) Faults;
3. Movement towards the satellite on the southern side of the Greendale East Fault (C) and within the region between the Greendale NW-SE (B) and Charing Cross

Faults (D); and

4. Movement away from the satellite in the corner between the Greendale Central (A) and the Greendale NW-SE Faults (B).

The signals were originally noted by Motagh et al. (2014b) based on an analysis of a short-term record of ascending CSK acquisitions. Due to the incidence angle 22° of the acquisitions, ground displacements projected in the LOS of the satellite are more sensitive to the vertical component of displacement than the horizontal components. The descending interferograms obtainable from descending CSK (Motagh et al., 2014b) and TSX observations are contaminated by long-wavelength tropospheric and orbital artefacts, which preclude an accurate interpretation of the sense of motion. Despite the existence of these long-wavelength artefacts, displacement profiles across the Greendale East Fault (Figure 5.7) from descending acquisitions of the CSK and TSX satellites both reveal motions towards the satellites on the southern side of the Greendale East Fault. Accordingly, we can interpret the observed signal as being mostly vertical motion rather than representing right-lateral displacement.

5.3.2 Modelling results

First-order approximation

We calculate the viscoelastic response of a Maxwell material to the coseismic stress changes due to the Darfield earthquake using the numerical code PyLith (Aagaard et al., 2009, 2013, 2015). An elastic thickness of 35 km is appropriate for the Canterbury region based on the 3-D seismic velocity model of New Zealand (Eberhart-Phillips et al., 2010). We assume a viscosity value of 10^{18} Pa·s for the viscoelastic half-space in order to maximise the effect of viscoelastic relaxation. The computed viscoelastic response in the time interval between the Darfield and Christchurch earthquakes is negligible (< 2 mm throughout the model) and is not discussed further below.

To understand what other physical processes produce the observed postseismic displacement, we also consider afterslip on faults that had ruptured coseismically. We use the public-domain software package RELAX, which implements the method described by Barbot et al. (2009), and assume the fault geometries and coseismic slip distributions computed by Beavan et al. (2012a). This approach allows us to compare the predicted three dimensional displacement field with the observations and examine whether an afterslip model alone can satisfy the observed postseismic displacement. We emphasise that the results of employing rate-strengthening afterslip, using RELAX, represent a

Figure 5.7: Profiles of displacement along the X1Y1 line shown in Figure 5.3b and X2Y2 in Figure 5.5. We can see movements towards the satellite of ~ 8 mm along the X1Y1 line and along the X2Y2 in Figure 5.3b and Figure 5.5, respectively. The signal is contaminated and reduced by a large amount of orbital and tropospheric error in the region.

first-order approximation of the sense of postseismic displacement due to afterslip on the assumed faults.

Assuming values of 50 mm/yr and 5 MPa for the two parameters governing rate-strengthening afterslip (rock and frictional properties: s_0 and $a\sigma$; see Barbot et al., 2009) and constraining the afterslip rake angles using the corresponding average coseismic rake angles on each fault, we compute postseismic displacements after the Darfield earthquake. Note that we do not use a formal method to choose these parameters: we test a range of 10–100 mm/yr and 1–10 MPa for the two model parameters to simulate the postseismic displacement within the observed range.

The modelled east, north and vertical components of displacement and the projected

displacement in the ascending LOS of the CSK satellite within approximately one month of the Darfield earthquake are illustrated in Figure 5.8. The movement away from the satellite observed in the step-over is not predicted by the model, nor is the signal in the corner between the Greendale Central (A) and the Greendale NW-SE Faults (B). The east component of displacement does exhibit movement towards the satellite on the southern side of the Greendale East Fault (C), but the LOS component mainly reflects the vertical displacement and is only weakly influenced by the east component. In contrast, the movement away from the satellite along the blind Charing Cross Fault (D) is represented reasonably well by the afterslip model. Movement towards the satellite is observed in the ascending CSK interferogram (Figure 5.3) between the northern part of the Greendale NW-SE Fault and the western side of the Charing Cross Fault. This motion cannot be accounted for with the assumed rate-strengthening afterslip rheology. Therefore, the afterslip model with the assumed parameters does not fit the overall sense of the observations well, with the exception of the signal associated with the Charing Cross Fault.

The forward model of rate-strengthening afterslip described above requires knowledge of two parameters ($\dot{s}_0$ and $a\sigma$). We have also investigated afterslip models estimated using inversion techniques by Motagh et al. (2014b) to examine whether these provide better matches to the observations (Figure 5.9). We observe the following points suggesting that afterslip does not adequately account for the observations:

- Significant reverse-slip on the Greendale NW-SE Fault, whereas the coseismic slip suggested by Beavan et al. (2012a) is mainly of right-lateral and normal sense (Figure 5.9a);
- Significant normal-slip on the right-lateral Greendale Central (Figure 5.9c);
- Significant reverse-slip on the right-lateral NE of the step-over and on the right-lateral Greendale East Faults (Figure 5.9d and 5.9e).

On this basis, we infer that postseismic mechanisms other than afterslip is responsible for producing the short-term post-Darfield ground displacements presented here and in the studies by Beavan et al. (2012a) and Motagh et al. (2014b). We next consider whether poroelastic rebound can better account for the sense and spatial extent of postseismic displacement.

Combined poroelastic rebound and dilatancy recovery

We estimate the displacement induced by poroelastic rebound following the Darfield earthquake using the method of Wang and Kümpel (2003). We use the national New

Figure 5.8: Surface displacement projected in the LOS of ascending CSK, one month after the Darfield earthquake based on a model of rate-strengthening afterslip on the previously ruptured faults during the Darfield earthquake. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite. Other faults are described in Figure 5.1b.

Zealand velocity model to assign the P- and S-wave velocities (Eberhart-Phillips et al., 2010) of between two and nine layers. The two-layer representation is a simplified version of the nine-layer model that is directly inferred from the national velocity model. Values for hydraulic diffusivity of either 0.1 or 1.0 m^2/s above a specified depth and 0.0001 m^2/s below that depth have been considered on the basis of the assumed shallow extent of poroelastic processes. The interface depths at which hydraulic diffusivity changes from higher to lower values are assigned values of either 1.0 or 1.75 km in our models. We assume values of 0.5 and 0.6 for Skempton's coefficient and 0.6 and 0.8 for Biot coefficient, and consider a range of model parameters in fitting the observations (Table 5.2).

Poroelastic rebound curves associated with various model parametrisations are illustrated in Figure 5.10 in the LOS direction of the ascending CSK orbit. The calculated displacement mainly represents the vertical component of movement due to the poroelastic rebound effects.

In order to best account for the observations illustrated in Figure 5.3, a high hydraulic diffusivity value of 1 m^2/s is required. Cases with higher values of 0.6 and 0.8 for Skempton's and Biot coefficients, respectively, overestimate the observed range increase and range decrease patches assuming hydraulic diffusivity of 1 m^2/s (Figure 5.10e). In contrast, a lower hydraulic diffusivity value of 0.1 m^2/s with the same

Figure 5.9: Modeling results for postseismic afterslip (version I, January 2012; Motagh et al., 2014a): Other faults are described in Figure 5.1b.

poroelastic coefficients as in Model 12 poorly accounts for the observed postseismic features (Figure 5.10f).

Our preferred poroelastic model is the two-layer Model 5, which has a Root Mean Square Error (RMSE) of 9 mm, Skempton's and Biot coefficients 0.5 and 0.6, respectively, and diffusivity values of 1 and 0.0001 m²/s above and below a depth of 1 km (Figure 5.10c). Despite its simplicity, Model 5 accounts for both the observed movements towards the satellite on the southern side of the Greendale East Fault (C) and in the region between the western Charing Cross (D) and northern Greendale NW-SE (B) faults, and away from the satellite within the step-over and in the corner of the Greendale Central (A) and Greendale NW-SE Faults (B). Nevertheless, Model 5 cannot account for the observed signal entirely: in particular, movements away from the satellite measured on the Greendale NW-SE (B), Charing Cross (D), near Hororata, NNW from the epicentre and NE from step-over faults are not fully explained.

Table 5.2: Model parameters of poroelastic rebound mechanism with different numbers of layers, depths of interface, values of diffusivity above and below the interface and Skempton's and Biot coefficients.

Number of Layers	Top layer	Model	Diffusivity m^2/s	Skempton's	Biot
9	$\geq 1.75\text{km}$	Model 1	1	0.5	0.6
		Model 6	0.1		
9	$> 1.75\text{km}$	Model 2	1	0.5	0.6
		Model 7	0.1		
9	$> 1.0\text{km}$	Model 3	1	0.5	0.6
		Model 8	0.1		
2	$> 1.75\text{km}$	Model 4	1	0.5	0.6
		Model 9	0.1		
2	$> 1.0\text{km}$	Model 5	1	0.5	0.6
		Model 10	0.1		
9	$> 1.75\text{km}$	Model 11	1	0.6	0.8
		Model 12	0.1		

Figure 5.11 presents the displacement time series extracted from InSAR analysis (shown in Figure 5.4) at point ‘A’ in Figure 5.1b and the corresponding poroelastic responses calculated with different model parametrisations. Although the effects of poroelastic rebound had reduced significantly by the time of the first InSAR acquisition, the signal persisted from 56 to 168 days after the earthquake. We infer that poroelastic rebound occurred during the early postseismic time interval, and that postseismic deformation continued even after poroelastic rebound terminated.

As previously explained, neither afterslip nor poroelastic rebound can account for the entirety of the observed postseismic displacement. To account for the residual deformation, we combine the poroelastic rebound mechanism (Model 5) with the process of dilatancy recovery at depths shallower than 4 km. In this case, as described in Section 5.2.2, the effects of afterslip on the faults that had ruptured coseismically is presumed to accompany dilatancy recovery. The amount of afterslip required on each of the faults other than the Charing Cross Fault to fit the observations is negligible after taking into account poroelastic displacements. Maximum dilatancy recovery is estimated to have occurred on the Greendale NW-SE and Charing Cross Faults, with an approximate value of $7 \times 10^4 \text{ m}^3$. Minor recovery is modelled at the western end of the Greendale East Fault, the NE Fault from step-over and on the left-lateral NNW Fault from the epicentre (Figure 5.12). The dilatancy recovery can partially explain the signal observed along several faults and the step-over in the region. The signal along the Charing Cross Fault can also be induced by afterslip. The observed, modelled and residual displacements of the proposed model and also the amount of displacement due

Figure 5.10: Estimated displacement induced by the poroelastic rebound mechanism following the Darfield earthquake in the LOS of ascending CSK with different values of poroelastic properties. a) Model 1: 9 layers with Skempton's 0.5, Biot 0.6 and diffusivity $1 \text{ m}^2/\text{s}$. b) Model 6: 9 layers with Skempton's 0.5, Biot 0.6 and diffusivity $0.1 \text{ m}^2/\text{s}$. c) Model 5: Shallow 2 layers with Skempton's 0.5, Biot 0.6 and diffusivity $1 \text{ m}^2/\text{s}$. d) Model 10: Shallow 2 layers with Skempton's 0.5, Biot 0.6 and diffusivity $0.1 \text{ m}^2/\text{s}$. e) Model 11: 9 layers with Skempton's 0.6, Biot 0.8 and diffusivity $1 \text{ m}^2/\text{s}$. f) Model 12: 9 layers with Skempton's 0.6, Biot 0.8 and diffusivity $0.1 \text{ m}^2/\text{s}$. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

to the recovery process are shown in Figure 5.13. This combined model reduces the RMSE from 9 mm, for poroelastic rebound only, to 5 mm.

5.4 Post-Christchurch displacement

5.4.1 GPS and InSAR observations

In Figure 5.14, we present horizontal campaign GPS velocities acquired annually between early 2012 and 2015. We show mean LOS velocities computed from TSX data acquired between 26 January 2012 and 22 May 2014, an interval chosen to exclude coseismic effects on the calculated ground displacements of the 13 June 2011 and the 23 December 2011 earthquakes.

Figure 5.11: Observed (from ascending CSK time series) and estimated time series at point A" (Figure 5.1b) within the Greendale Fault step-over induced by different multi-layered poroelastic rebound with two diffusivity values of 1 and 0.1 m^2/s , and the values of 0.5 and 0.6 for Skempton's and Biot coefficients. Model 5 is the preferred spatial model (see Figure 5.10).

Figure 5.12: Regular grid of Mogi point sources on the ruptured fault planes during the Darfield earthquake. Red colour indicates the volume decrease of each Mogi point source. Other faults are described in Figure 5.1b.

Three key features are evident in the velocity results extracted from GPS velocities and InSAR time series:

- Movement away from the satellite at a velocity of ~ 3 mm/yr in the Christchurch urban region and along the coast;
- A ground displacement anomaly across the WSW continuation of the Western Christchurch Fault (H), which coincides with the lineation of aftershocks recognised following the Darfield and Christchurch earthquakes (Sibson et al., 2011; Syracuse et al., 2012; Herman et al., 2014). Beavan et al. (2012a) discussed the possibility of seismic or aseismic slip inducing westward and eastward motions at

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Figure 5.13: a) Observed ground displacement measured by ascending CSK as in Figure 5.3a. b) Estimated displacement induced by the combined dilatancy recovery and poroelastic rebound projected in the LOS of ascending CSK. c) Residual displacement induced by the combined dilatancy recovery and poroelastic rebound projected in the LOS of ascending CSK. d) Estimated displacement induced by the process of dilatancy recovery only.

Figure 5.14: Observed mean LOS velocity acquired from TSX acquisitions following the Christchurch earthquake. Arrows represent GPS horizontal velocities in the Pacific plate reference frame relative to the continuous GPS site MQZG for three periods: early 2012-2013 (blue), 2012-2014 (red) and 2012-2015 (maroon). Red colours indicate movement away from the satellite and blue colours indicate movement towards the satellite. Dashed line is microearthquake lineament (inferred to be the continuation of fault Segment H). Other faults are described in Figure 5.1b.

continuous GPS sites MQZG and SWAN, respectively (location of cGPS sites in Figure 5.1b). The relative motion between MQZG and SWAN is in agreement with the right-lateral velocity observed with InSAR and GPS.

- Movement away from the satellite at the easternmost end of the Greendale East Fault.

The second and third features are likely to be associated with the long-term post-Darfield displacement. Too few acquisitions spanning the eastern half of the Greendale East Fault were made following the Darfield and prior to the Christchurch earthquakes to analyse the post-Darfield displacement at different timescales.

5.4.2 Modelling results

First-order approximation

The viscoelastic responses to the Darfield and Christchurch earthquakes are calculated and projected in the LOS of the TSX satellite for the period between 26 January 2012 and 22 May 2014.

We cannot account for the observed velocity, measured by TSX acquisitions, in terms of the insignificant velocity associated with plausible viscoelastic responses (< 2 mm/yr).

In order to interpret the observed displacement features, we instead consider rate-strengthening afterslip (Barbot et al., 2009) on the lineament and the coseismic faults that had ruptured during the Christchurch earthquake. The lineament is assumed to be a continuation of the Western Christchurch Fault (H in Figure 5.1b) and to have a length of ~ 15 km, a dip of 070° SSE and a strike of 067° . The extent of the microseismicity defines the length of the lineament (extension of H).

Assuming values of 20 mm/yr and 5 MPa for s_0 and $a\sigma$, respectively, we first solve for the shear stress changes and afterslip on the lineament at successive times. The values are optimised to obtain a similar magnitude of discontinuity across the lineament.

Figures 5.15a and b represent the strike and dip components of afterslip on the lineament at four epochs between 137 days and 1095 days after the Christchurch event. Afterslip on the lineament is mostly concentrated on its easternmost ~ 4 km edge, adjacent to the coseismic patch. The computed afterslip model involves predominantly right-lateral motion with a maximum value of ~ 50 mm at the final epoch (three years after the earthquake). This is consistent with the faulting geometry of the Green-dale Fault and other recently recognised eastwest-striking faults in the Canterbury region (Gledhill et al., 2011; Kaiser et al., 2012; Quigley et al., 2012; Beavan et al., 2010b, 2011; Bannister et al., 2011; Syracuse et al., 2012; Sibson et al., 2011; Bannister and Gledhill, 2012; Beavan et al., 2012a). Modelled surface displacement along the lineament reaches a maximum of ~ 10 mm within 1095 days of the Christchurch earthquake (Figure 5.15).

We apply the same method using the previous constitutive parameters to estimate the rate-strengthening afterslip on the combination of the nearby faults and the lineament. Figure 5.16 illustrates the cumulative ground displacement at the final epoch three years after the Christchurch earthquake. The net ground displacement modelled by afterslip exhibits eastward displacement on the northern side of the lineament and the opposite direction of motion in the southern side (Figure 5.16a2). The horizontal motion is explained by continued right-lateral strike-slip faulting. Northward motion is computed by the model on the northern side of the Eastern Christchurch Fault (Figure 5.16a3). An uplift signal is estimated on the eastern side and a subsidence signal on the western side of the NNE Fault (I) due to the reverse afterslip (Figure 5.16a4). The afterslip model appears to reproduce the sense and magnitude of displacement associated with the WSW extension of fault segment (H) delineated by aftershock activity (Figure 5.16c). However, the afterslip model cannot explain the subsidence signal observed in the urban area of Christchurch or along the coast. The computed LOS displacement after three years induced by afterslip on fault segments (I) and (J) show

net uplift in that area (Figure 5.16d). The profiles of estimated velocity in the corresponding interval due to the rate-strengthening afterslip (red) on the lineament and nearby faults along X1Y1 and X2Y2 (Figure 5.14) are shown with the corresponding observations (grey) in Figure 5.16c and d, respectively.

Based on the results of the afterslip modelling, we conclude the following:

- Right-lateral frictional afterslip on the extension of fault segment (H) is localised near the westernmost end of the Western Christchurch Fault. The observed discontinuity across the lineament seems to extend further towards the easternmost end of the Greendale East Fault (C) demonstrated by velocity of the descending TSX time series and GPS velocities (Figure 5.14); and
- Frictional afterslip on the faults that ruptured coseismically during the Christchurch earthquake does not account for the observed subsidence in the Christchurch urban region or along the coast.

Inversion of GPS and InSAR velocities

As previously discussed, a broad eastward displacement is observed in the GPS data, in three successive years, and in descending TSX InSAR timeseries. A forward model of afterslip on the Christchurch faults and the lineament accounts for neither the subsidence in the urban region of Christchurch, nor the uniform discontinuity across the lineament. Therefore, in order to understand the observed discontinuity, we invert the ground displacement measurements in terms of afterslip on the Christchurch faults (Figure 5.14 H, J and M), the lineament (extension of H), the eastern end of the Greendale East (Figure 5.14 C) Fault, and NE Fault from step-over (Figure 5.14 G). In the afterslip models, ground velocities are calculated relative to the continuous GPS site MQZG (location shown in Figure 5.1).

The broad eastward movement continues on the northern side of the faults towards the eastern end of the fault segments associated with the Darfield earthquake. This signal may imply that afterslip is activated not only by the Christchurch earthquake, but also by the preceding Darfield earthquake. The unavailability of long-term observations following the Darfield earthquake, spanning the entire region including the fault segments associated with both earthquakes, means that this hypothesis cannot be verified.

We consider four afterslip models: 1) single fault model on the lineament; 2) six fault model on the lineament as well as NE Fault from step-over (G), Greendale East Fault (C), SW-NE Fault (M), and Western and Eastern Christchurch Faults (H and J, respectively); 3) three fault model on the lineament, NE Fault from step-over and Greendale

Figure 5.15: Estimated a) Strike-slip and b) Dip-slip magnitudes on the subsurface lineament at successive times following the Christchurch earthquake assuming rate-strengthening rheology on the fault with the values of 20 mm/yr and 5 MPa for $\dot{s}_0$ and $a\sigma$, respectively for constitutive parameters. c) Surface displacement projected in the LOS at successive times following the Christchurch earthquake based on a model of rate-strengthening afterslip on the subsurface lineament. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

East Fault; and 4) five fault model on the NE Fault from step-over (G), Greendale East Fault (C), SW-NE Fault (M) and Western and Eastern Christchurch Faults. The magnitudes and rakes of afterslip for each model (rows) and each epoch (columns) are illustrated in Figure 5.17. The second epoch between early 2012 and 2014 is constrained by the InSAR data between January 2012 and May 2014. The velocity of observations (blue) and models (red) are shown in Figure 5.18, 5.19 and 5.20 for each model and three epochs, respectively.

Major right-lateral slip is inferred to reproduce the eastward movement on the northern side of the faults. The amounts of reverse-slip on the SW-NE (M) and Eastern Christchurch Faults are consistent with the oblique coseismic slip. We compare the RMSE of the slip models in Table 5.3. The single fault afterslip model does not re-

Figure 5.16: a) Surface displacement produced by accumulated afterslip on the previously ruptured faults during the Christchurch earthquake and on the subsurface lineament in the three years following the Christchurch earthquake, a1) LOS, a2) East, a3) North and a4) Vertical components of the estimated ground displacement. b) Observed LOS velocity from TSX following the Christchurch earthquake as in Figure 5.14. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite. c) Profile of estimated velocity, in the corresponding interval, and observation (grey) along the X1Y1 line presented in Figure 5.14. d) Profile of estimated velocity and observation along the X2Y2 line presented in Figure 5.14.

produce the observations sufficiently. Model 2 predicts the observations with slightly lower RMSE values in different epochs compared to other models.

Figure 5.17: Magnitudes and rake directions of four afterslip models (rows) for each epoch (columns). Models: 1) on the lineament; 2) on the lineament as well as NE Fault from step-over (G), Greendale East Fault (C), SW-NE Fault (M) and Western and Eastern Christchurch Faults (H and J, respectively); 3) on the lineament, NE Fault from step-over and Greendale East Fault; 4) on the NE Fault from step-over, Greendale East Fault, SW-NE Fault and Western and Eastern Christchurch Faults. The second epoch between early 2012 and early 2014 is constrained by the InSAR timeseries velocity between January 2012 and May 2014.

Figure 5.18: Velocity of observations and models are shown in blue and red, respectively for each model (epoch 2012-2013). Models: 1) on the lineament; 2) on the lineament as well as NE Fault from step-over (G), Greendale East Fault (C), SW-NE Fault (M) and Western and Eastern Christchurch Faults (H and J, respectively); 3) on the lineament, NE Fault from step-over and Greendale East Fault; 4) on the NE Fault from step-over, Greendale East Fault, SW-NE Fault and Western and Eastern Christchurch Faults.

Figure 5.19: Velocity of observations and models are shown in blue and red, respectively for each model (epoch 2012-2014). Models: 1) on the lineament; 2) on the lineament as well as NE Fault from step-over (G), Greendale East Fault (C), SW-NE Fault (M) and Western and Eastern Christchurch Faults (H and J, respectively); 3) on the lineament, NE Fault from step-over and Greendale East Fault; 4) on the NE Fault from step-over, Greendale East Fault, SW-NE Fault and Western and Eastern Christchurch Faults. Models are constrained by the InSAR timeseries velocity between January 2012 and May 2014.

Subsidence signal in the urban region of Christchurch

We consider the observed subsidence signal in the urban region of Christchurch and along the coast to be the result of either 1) poroelastic rebound; and/or 2) dilatancy recovery as observed following the Darfield earthquake; and/or 3) other sources such as liquefaction (e.g. Cubrinovski et al., 2011; Cubrinovski and Taylor, 2011; Mucciarrelli, 2011; Cubrinovski et al., 2012; Quigley et al., 2013; Van Ballegooy et al., 2014). One method of studying areas affected by soil liquefaction involves the use of InSAR coherence maps, as Atzori et al. (2012) did after the February and June Christchurch earthquakes of 2011. The subsidence signal detected in Christchurch is located within the liquefaction affected areas. Displacement velocities were measured within a time interval of more than two years after the last event in December 2011. Therefore, the impact of the soil liquefaction had likely diminished by this time and poroelastic

Figure 5.20: Velocity of observations and models are shown in blue and red, respectively for each model (epoch 2012-2015). Models: 1) on the lineament; 2) on the lineament as well as NE Fault from step-over (G), Greendale East Fault (C), SW-NE Fault (M) and Western and Eastern Christchurch Faults (H and J, respectively); 3) on the lineament, NE Fault from step-over and Greendale East; 4) on the NE Fault from step-over, Greendale East, SW-NE Fault and Western and Eastern Christchurch Faults.

rebound or dilatancy recovery are likely the principal causes of urban subsidence.

Two values for hydraulic diffusivity of $0.1 \text{ m}^2/\text{s}$ above the depth of 1.0 km and a small value of $0.0001 \text{ m}^2/\text{s}$ below the depth of 1.0 km are considered, assuming any poroelastic rebound to be a shallow process. We assume values of 0.6 and 0.89 for Skempton's and Biot coefficients in order to allow the poroelastic rebound effects to last until the final acquisition of the InSAR timeseries in May 2014. Figure 5.21 illustrates the ground displacement produced by poroelastic rebound at four epochs. A zone of subsidence is computed, but the corresponding velocity in the observed time interval is insignificant in the area bounded by the Eastern (J) and NNE (I) fault planes ($<1 \text{ mm/yr}$).

We next examine the effects of possible dilatancy recovery in Christchurch, in order to interpret the subsidence signal using the ground velocity measured by InSAR time series. A regular grid of Mogi point sources is modelled at depths shallower than 4 km .

Table 5.3: Root Mean Square Residuals in three epochs of four slip models: 1) on the lineament; 2) on the lineament as well as NE Fault from step-over, Greendale East Fault, SW-NE Fault and Western and Eastern Christchurch Faults; 3) on the lineament, NE Fault from step-over and Greendale East Fault; and 4) on the NE Fault from step-over, Greendale East Fault, SW-NE Fault and Western and Eastern Christchurch Faults.

	2012-2013	2012-2014	2012-2015
Model 1 (mm)	2.7	1.9	1.9
Model 2 (mm)	2.3	1.6	1.6
Model 3 (mm)	2.5	1.7	1.7
Model 4 (mm)	2.4	1.8	1.7

Figure 5.21: Modelled ground displacement associated with poroelastic rebound with a diffusivity value of $0.1 \text{ m}^2/\text{s}$ and the values of 0.6 and 0.89 for Skempton's and Biot coefficients, projected in the LOS direction. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

The Mogi point sources are located on the fault planes related to the February, June and December 2011 earthquakes (H, I, J, K, L and M). A maximum volume decrease of $\sim 1.8 \times 10^4 \text{ m}^3$ is inferred on the December SW-NE Fault. Figure 5.22 summarises the results.

We combine the ground velocity predicted by Model 2 (best RMSE: Table 5.3) between early 2012 and 2014 as the result of inversion on both GPS and InSAR velocities and the estimated velocity resulting from dilatancy recovery. We present the observed, estimated, and residual displacements of combined afterslip with dilatancy recovery in Figure 5.23.a, b and c, respectively.

Figure 5.22: Volume decreases of a grid of regularly spaced contracting Mogi point sources on the fault planes that ruptured by the 22 February, 13 June and 23 December 2011 Christchurch earthquakes.

5.5 Discussion

Significant orbital and atmospheric artefacts in descending acquisitions of the CSK and TSX satellites make it difficult to determine the sense of postseismic ground motion immediately following the Darfield earthquake (Figure 5.3b and 5.5). However, we can infer from the profiles in Figure 5.7a and b that the movement towards the satellite, observed by ascending and descending interferograms, in the southern side of the Greendale East Fault is mainly a vertical motion.

We have considered possible rate-strengthening afterslip on the faults that ruptured during the mainshock. The results (Figure 5.8) imply that the afterslip model cannot thoroughly account for the spatial variability of short-term observations.

Pore pressure changes induced coseismically induce fluid flow. The increase of the groundwater levels observed in unconfined and semi-confined aquifers in the Canterbury region within a day of the Darfield earthquake provide evidence of poroelastic rebound (Cox et al., 2012). We compare the temporal ground displacement estimated by poroelastic rebound with values of 0.5, 0.6 and 1 m²/s for Skempton's and Biot coefficients and hydraulic diffusivity, respectively, with the time series of measured displacement over the fault step-over in Figure 5.11 and the unwrapped interferogram in Figure 5.3. These comparisons show that the mechanism of poroelastic rebound does not thoroughly estimate the time-dependent surface displacement within the five months between the Darfield and Christchurch earthquakes.

Figure 5.23: a) Observed ground velocity measured by descending TSX as in Figure 5.6b. b) Modelled LOS velocity for TSX produced by combined afterslip (Model 2) and dilatancy recovery. c) Residuals of the model produced by combined afterslip (Model 2) and dilatancy recovery following the Christchurch earthquake. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

We interpret the main observational features after the Darfield earthquake to be induced by dilatancy recovery due to weakening of the crust during the mainshock (Reyners et al., 2014) and poroelastic rebound.

Following the Christchurch earthquake, afterslip yields a broad discontinuity, across the lineament, observed by GPS and InSAR velocities. The ongoing movement between early 2012 and 2015 implies that afterslip persisted after the Christchurch earthquake and associated large aftershocks. Two questions remain unanswered by our analysis:

1) the western extent of afterslip and the corresponding displacement towards the fault segments associated with the Darfield earthquake; and 2) the time of the onset of afterslip following the Canterbury earthquake sequence of 2010–2012.

We have considered afterslip on five fault segments associated with the Darfield and Christchurch earthquakes as well as the proposed subsurface lineament, to explain the observed discontinuity. Model 2 in Table 5.3 provides the best fit with observations. This model involves slip on the lineament in addition to five fault segments.

The strike of the observed aftershock lineament of 067° forms a non-Andersonian angle of $\sim 48^\circ$ with the orientation of the axis of maximum horizontal compressive stress (Townend et al., 2012; Holt et al., 2013). This geometry may either mean that there is a possible dip component of slip on the proposed lineament, so the subsurface lineament is not purely right-lateral, or that the lineament is frictionally weak. Sibson et al. (2011) recognised this lineament by the diffuse aftershock concentration south of the Darfield earthquake surface rupture (see Bannister et al., 2011; Gledhill et al., 2011; Bannister and Gledhill, 2012).

The subsidence signal observed within the urban region of Christchurch is located in the area of shallow confined aquifers along the coast (Cox et al., 2012). The hydrological responses in aquifers and/or liquefaction are likely to have affected the velocity of postseismic ground movement. In addition to hydrological effects of the Christchurch earthquake and aftershocks, dilatancy recovery and/or poroelastic rebound may account for the observed movement away from the descending TSX in the region.

5.6 Conclusions

We have examined several alternative mechanisms of postseismic displacement in order to explain geodetic observations made following the Darfield and Christchurch earthquakes. We infer that processes of poroelastic rebound and dilatancy recovery occurred immediately following the Darfield earthquake. Afterslip may also have occurred on the Charing Cross Fault (D), but afterslip does not account for spatial variability of observations. The subsequent Christchurch earthquake was followed by two large aftershocks on 13 June and 23 December 2011 as well as ongoing afterslip (on the lineament and fault segments G, C, M, H and I) observed in three successive years between early 2012 and 2015. Afterslip is inferred to have occurred on the lineament forming an extension of the Western Christchurch Fault and defined by a band of seismicity. The SW-NE December, Eastern and Western Christchurch Faults along with Greendale East and

NE Fault from step-over are likely to have also undergone postseismic afterslip. Finally, we conclude the observed subsidence signal in Christchurch to be the result of hydrological effects following the earthquake, such as liquefaction or mechanisms of poroelastic rebound and dilatancy recovery.

Chapter 6

Summary and recommendations

6.1 Post-Dusky Sound displacement

In Chapter 3, we use GPS measurements of 16 continuous sites taken within 400 days of the Dusky Sound earthquake, supplemental campaign GPS data acquired during August 2009 and Differential Interferometric Synthetic Aperture Radar (DInSAR) observations to address the postseismic mechanisms following the mainshock.

We demonstrate that viscoelastic rebound models do not account for either observed short-term or southward displacement in the near-epicentral area (Section 3.3). Therefore, significant afterslip is required to entirely explain the postseismic displacement.

We resolve time-dependent afterslip models using PCAIM-inversion (Principal Component Analysis-based Inversion Method) framework (Section 3.2.3). To regularise the time-dependent inversion, we apply Laplacian smoothing and decompose the displacement signals into spatial and temporal functions. In contrast with other time-dependent inversion techniques, this method does not require the introduction of additional temporal hyperparameters.

We account for the viscoelasticity of the lower crust and upper mantle by subtracting viscoelastic responses to coseismic stress changes from observed displacements. A Maxwell rheology with viscosities of 5×10^{18} – 10^{20} Pa·s and an elastic thickness of 45 km are assumed. We consider the viscoelastic responses to postseismic stress changes induced by afterslip throughout an iterative method (Figure 3.5). We demonstrate that afterslip models vary depending on the viscosity of the viscoelastic half-space. We therefore conclude that postseismic mechanisms and associated stress and strain models cannot be unique while the viscosity of the half-space is unknown.

Four options for slip-boundary constraints have been examined. In our final model, with no-slip boundary constraints on the sides and bottom of the fault, we estimate

two main patches located on either sides of the coseismic slip patch (white contours in Figure 3.9) reaching to the ground surface and depth 80 km.

Resolving a unique solution for the afterslip spatial modes depends on the density of ground measurements; The sparsity of the continuous GPS measurements in the central and southern South Island may cause ambiguity in our final postseismic model. There is also a lack of campaign GPS measurements spanning the early postseismic interval excluding the coseismic signal. Consequently, the misfit of the postseismic model with the campaign observations is mostly due to the uncertainty of the slip model used to remove the coseismic effects. The campaign measurements do not span the long-term postseismic interval between the Dusky Sound and the Darfield earthquakes. As a result, long-term spatial modes of the afterslip model are poorly constrained.

Strain and stress changes induced by the earthquake and corresponding postseismic mechanisms are calculated for the southern and central South Island (Section 3.4.3). Taking account of estimated postseismic mechanisms yields enlargement of the contractional and extensional strain vectors and expansion of the zones of positive coseismic Coulomb stress changes. In the Canterbury region, we estimate negative Coulomb stress changes on the Charing Cross Fault and a few kilopascals of positive Coulomb stress changes on the Greendale Fault. Principal contractional axis is calculated as 42° clockwise around the secular contractional axis and 35° clockwise around the previously determined S_{Hmax} . Principal contractional and extensional strains are 4% and 80% of the equivalent secular principal strains, respectively. Accordingly, we infer that Coulomb stress changes and strain rates induced by the earthquake and postseismic mechanisms are insignificant in the Canterbury region. Therefore, we cannot conclude that either the Dusky sound earthquake triggered or advanced the Canterbury earthquake sequence of 2010–2012.

Further research

The cGPS measurements are corrected for seasonal effects and mean secular velocities (Section 3.2.1). This dataset spans the time interval between the Dusky Sound and Darfield earthquakes for a period of approximately 400 days. There is a long-term campaign dataset available taken during February 2014. We did not constrain the long-term effect of postseismic mechanisms with this dataset. Extending the time interval for the cGPS data and constraining the postseismic models with this campaign measurement is recommended for further study. The uniqueness of the postseismic models can be improved considering the long-term effects of viscoelastic rebound.

6.2 Coupling afterslip and viscoelastic rebound

In Chapter 4, we propose the Iterative Decoupling of Afterslip and Viscoelastic relaxation (IDAV) method to decouple the time-dependent postseismic displacement into elastic and viscoelastic space and time functions. We examine our framework with analytical and numerical sample data produced by the semi-analytical viscoelastic Green's functions and the Finite Element Method (FEM), respectively. We also apply this method to the observed postseismic displacement following the Dusky Sound earthquake assuming viscosities of 2×10^{18} – 4×10^{18} Pa.s. We demonstrate ambiguity in modelling the postseismic mechanisms in order to explain the Post-Dusky Sound displacement. Variation of viscosity values equal to or greater than 2×10^{18} Pa.s yields different afterslip models.

Contributions

The main contribution of this chapter is adapting the PCAIM-inversion method to accommodate the viscoelastic Green's functions. The general form of the Sylvester equation (Equation 4.26; Section 4.2) is simplified with the assumptions that:

- One temporal mode suffices to decouple the viscoelastic Green's functions; and
- The spatial mode of the elastic displacement is obtained from unconstrained or constrained decomposition with imposing the temporal mode of the viscoelastic Green's functions.

Further research

We decompose the observed postseismic signals into two components of elastic and viscoelastic displacements. We do not consider temporally independent sources of displacement, which induce independent components in the observed signal. Adapting our inversion method to accommodate other techniques such as the Independent Component Analysis Inversion Method (ICAIM) is a topic of further study (Gualandi et al., 2015). Accordingly, ICAIM is able to separate sources that are temporally independent in observed signals.

We assess the IDAV method for use as an inversion framework for understanding post-seismic transient afterslip. Potentially this method is applicable to the inversion of geodetic data to estimate the spatial and temporal evolution of Slow Slip Events (SSEs)

in an elastic/viscoelastic medium.

The effect of fluid flow due to the coseismic stress changes induces a short-term displacement in the total observed signal (Peltzer et al., 1996, 1998). Using the IDAV method, the total displacement assumed to be the sum of viscoelastic response to coseismic changes and elastic and viscoelastic responses to afterslip. Therefore, corresponding displacement produced by poroelastic effects of earthquakes should be considered in future studies.

The IDAV method converges to a unique solution as a result of spatial Laplacian smoothing and decomposition. However, this method does not guarantee that the solution is physically significant. For instance, a viscosity of 10^{18} Pa·s as used in analysing the postseismic displacement following the Dusky Sound earthquake, yields a turning point in the estimated elastic time function. Other regularisation techniques of the Sylvester equation may ensure the physical significance of the solutions.

A unified inversion method considers the effects of viscoelastic and poroelastic rebound, the heterogeneity of lithospheric structure and other complexities such as an elastic-slab geometry. This framework provides a more realistic understanding of the spatial and temporal evolution of seismic and aseismic slips and SSEs.

Our attempt in Chapter 4 requires more steps to be taken towards a unified framework for inversion of temporally dense geodetic observations. The ultimate aim of time-dependent inversion techniques is achieving an insight into controlling factors and interactions between various phenomena in a seismic cycle.

6.3 Post-Canterbury displacement

We experience difficulties in interpreting the sense of postseismic movement along and across the faults that ruptured during the Darfield earthquake. The TerraSAR-X (TSX) and COSMO-SkyMed (CSK) acquisitions on descending tracks are contaminated by major orbital and atmospheric artefacts. There is also a lack of campaign GPS measurements spanning the area encompassing the faults that ruptured during the Darfield earthquake. As a result, we cannot evaluate the sense of postseismic displacement using both InSAR and GPS data.

The frictional afterslip is computed as a first order approximation. Displacement profiles across the Greendale East Fault are obtained from the descending TSX and ascending CSK interferograms. Such additional analyses enable us to distinguish between

the postseismic displacement processes in spite of the uncertainties described above. We conclude that poroelastic rebound and dilatancy recovery are the major postseismic mechanisms operating within approximately one month following the Darfield earthquake.

A dataset including ascending CSK acquisitions spans eastern end of the Greendale East Fault towards the western end of the Western Christchurch Fault in the time interval between the Darfield and Christchurch earthquakes. In this time interval, 12 acquisitions do not suffice to mitigate the long-wavelength artefacts completely in the region. Therefore, we cannot interpret the longer-term Darfield earthquake postseismic mechanisms and the corresponding stress transfer to the faults that ruptured during the Christchurch earthquake.

The dataset used following the Christchurch earthquake includes 66 descending TSX acquisitions. The estimated temporal displacement is affected by random variations of tropospheric perturbations. Stacking the interferograms mitigates the effects of random tropospheric artefacts in the estimated velocity. We observe and model a displacement gradient across the subsurface lineament extending from the Greendale East towards the Western Christchurch Faults. To explain this observation, we suggest afterslip on combinations of several faults: the lineament, Greendale East, NE Fault from epicentre and Western and Eastern Christchurch Faults. We cannot infer with certainty that the observed gradient is due to the aseismic slip on the lineament, although the lineament has been associated with a band of seismicity.

A subsidence signal is measured by this dataset in Christchurch and along the coast. We associate this signal to either dilatancy recovery on the faults that ruptured during the Christchurch earthquake and aftershocks in June and December and/or poroelastic rebound and/or hydrological effects of the earthquake such as liquefaction.

Further research

Fault damage and its recovery may justify the lack of afterslip following several strike-slip earthquakes (Fialko et al., 2005; Fielding et al., 2009) around the world such as the Darfield earthquake.

In Section 5.3.2, a regular grid of Mogi point sources is used to model the postseismic subsidence signal on the faults that ruptured during the Darfield earthquake. This model does not address coseismic and postseismic changes of seismic P- to S-wave velocity ratios due to the weakening of the crust. A physical method with which to

estimate the displacement signal resulting from fault-damage is of great use.

6.4 Unanswered questions from Section 1.3

We provide answers to questions posed in Section 1.3 as the closing section of the thesis.

1. What can geodetic data reveal about postseismic displacement in response to multiple earthquakes?

- (a) What circumstances or conditions are favourable in enhancing the effectiveness of InSAR for measuring low-strain rate tectonic deformations?

Measuring low-strain rate tectonic deformations using conventional InSAR techniques encounters several major limitations. Spatial and temporal decorrelation due to changes in scattering properties of the ground surface (Zebker and Villasenor, 1992), spatial and temporal variations of atmospheric effects and errors in satellite orbits and surface elevation often overwhelm the determination of crustal deformation (Zebker et al., 1997). Ferretti et al. (2000, 2001) tackle the problem of coherence loss due to decorrelation by identifying permanent scatterers in urban regions. In urban regions more permanent scatterers can be identified by measuring their amplitude scintillations. Hooper et al. (2004) used phase analysis in order to identify significant number of permanent scatterers in non-urban regions to measure the episodic volcanic deformation. This method is used in our displacement measurement in Section 5.4.1 to estimate the post-Christchurch displacement. Time and space variations of atmospheric water vapour lead to up to 10 cm errors in measuring ground displacement in wet regions such as Hawaii and potentially New Zealand. This effect is a major issue we encounter in our displacement measurement in the Canterbury region using TerraSAR-X acquisitions. The only way to mitigate this error is using multi passes of the same region (Zebker et al., 1997). Time series techniques by averaging independent interferograms and generating velocity maps significantly reduce the effect of atmospheric delays. Tropospheric variations of temperature and pressure are of smaller magnitudes and elevation-correlated (Zebker et al., 1997).

2. What mathematical framework can we use to provide unique solutions of displacement and stress evolution for postseismic time intervals?

- (a) How do we optimally choose rheology parameters in order to model the displacement and stress evolution?
- (b) How can the uncertainty of input parameters for models such as fault geometry and slip distribution change the inferred stress changes?
- (c) Is the inferred model of stress evolution unique?

The mathematical framework introduced in Chapter 4 is one step forward towards providing unique solutions of displacement and stress evolutions. This framework allows us to explain postseismic displacement signals by the coupled mechanism of afterslip and viscoelastic relaxation. Taking into account the effect of poroelasticity is left for further study. We have discussed the non-uniqueness of postseismic solutions in Section 2.2.1. We explained how to improve the uniqueness of postseismic solutions in terms of using other observations from sources such as tomography, thermo-kinematics models, seismicity, gravity signals observed by the GRACE mission and the local hydro-isostatic adjustment associated with the mantle flow. Assumptions made in Chapter 3 regarding the viscosity of the lower crust and upper mantle, the thickness of elastic layer and the geometry of subduction zone require more complementary (structural) information to provide unique solutions of postseismic mechanisms and stress models.

- 3. Can we assess the triggering effect of each earthquake on the later events by modelling the stress and strain evolution? Particularly, can we calculate the strain and stress evolution following the Dusky Sound earthquake and examine its effects on the Canterbury earthquake sequence of 2010–2012?

We have assessed the triggering effect of the Dusky Sound earthquake on the Canterbury earthquake sequence by means of accumulated strain and stress post-seismic added to coseismic models (see Section 3.4.3). This method accounts for the static stress changes due to coseismic and postseismic mechanisms, which is limited to relatively smaller distances of a few 100 kms.

Appendices

Appendix A

Coulomb stress changes

The transformation matrix equation for a stress tensor from the geographic coordinate system (S_{xyz}) to the fault-plane coordinate system ($S_{x'y'z'}$) is

$$S_{x'y'z'} = RS_{xyz}R' \quad (\text{A.1})$$

where

$$R = \begin{bmatrix} 1 & 0 & 0 \\ 0 & -1 & 0 \\ 0 & 0 & 1 \end{bmatrix} \begin{bmatrix} 1 & 0 & 0 \\ 0 & \cos(D) & \sin(D) \\ 0 & -\sin(D) & \cos(D) \end{bmatrix} \begin{bmatrix} \cos(S) & -\sin(S) & 0 \\ \sin(S) & \cos(S) & 0 \\ 0 & 0 & 1 \end{bmatrix} \quad (\text{A.2})$$

and

$$S = s - 90 \quad (\text{A.3})$$

where s and D are the strike and dip angles of the fault (Figure A.1).

Figure A.1: Transformation from the geographic coordinate system to the fault-plane coordinate system

Normal (σ_n) and shear stress (τ) components on the fault plane are obtained from the

transformed stress tensor in the fault-plane coordinate system

$$\sigma_n = S_{z'z'} \quad (\text{A.4})$$

$$\tau = \cos(r)S_{x'z'} - \sin(r)S_{y'z'} \quad (\text{A.5})$$

where r is rake angle and the transformed stress tensor is of the form

$$S_{x'y'z'} = \begin{bmatrix} S_{x'x'} & S_{x'y'} & S_{x'z'} \\ S_{x'y'} & S_{y'y'} & S_{y'z'} \\ S_{x'z'} & S_{y'z'} & S_{z'z'} \end{bmatrix} \quad (\text{A.6})$$

Pore pressure (P) is obtained from average of the normal stress components in the new coordinate system

$$P = (S_{x'x'} + S_{y'y'} + S_{z'z'})/3 \quad (\text{A.7})$$

therefore, Coulomb stress change is expressed as

$$\Delta Cff = \tau + f(\sigma_n - BP) \quad (\text{A.8})$$

where f and B are fault friction and Skempton's coefficients, respectively. All the variables are summarised in Table A.1.

Table A.1: Table of nomenclature

D	Dip
s	Strike
S	$str - 90$
S_{xyz}	Cartesian stress tensor
$S_{x'y'z'}$	Transformed stress tensor
R	Rotation matrix
r	Rake
σ_n	Normal stress
τ	Shear stress
f	Fault friction coefficient
B	Skempton's coefficient
P	Pressure
ΔCff	Coulomb stress changes

Appendix B

Global compilation of viscosity, rheology and depth of the top viscoelastic layer

For each row of Table B.1, we summarise the metadata of the earthquakes. In Table B.2, for each row we mention rheology, vertical or lateral heterogeneities and viscosity value for each layer. Regarding the rheology field, "*M*" stands for Maxwell, "*B*" Burger's body, "*P*" Power-law, "*SLS*" Standard Linear Solid and "*NN*" is the general term Non-Newtonian fluid. Regarding the viscosity field, layers are separated with " – " symbol, multiple viscosities for each layer with "," and the range of viscosity in a parenthesis "()". Table B.3 is the summary of the thickness and depth of each layer. "*SL*" stands for Slab and "*UP*" Upper Plate thickness in case of subduction earthquakes. "*LC*", "*UM*" and "*HS*" stand for depths associated with Lower Crust, Upper Mantle and Half-Space for non-subduction zone earthquakes.

Num	Year	Region	Magnitude M_w
1	2011	Tohoku-oki	9
2			
3			
4	1993	Hokkaido Nanseioki	7.8
5	1891	Nobi	8
6	1896	Riku-u	7.2
7	1951, 1952	Tibetan Plateau	7.7, 7.6
8	2004	Sumatra-Andaman	9.1
9			
10			
11			
12			
13	2004, 2005	Sumatra, Nias	9.1, 8.7
14			
15			
16	2007	Southern Sumatra	8.5
17			
18			
19	2010	Salton trough	7.2
20			
21			
22	1992	Landers	7.3
23			
24			
25	1999	Hector Mine	7.1
26	1992, 1999	Landers, Hector Mine	7.3, 7.1
27	1999	Hector Mine	7.1
28	2004	Parkfield	6.0
29	1812 onwards	Southern California	> 6.5
30	1857	Fort Tejon	7.9
31	1906	San Francisco	7.8
32	1999	North Anatolian	7.6
33			
34			
35	2006, 2007	Kuril	> 8
36			

37			
38	1952-1965	Kurile-Kamchatka	
39			
40	1999	Chi-Chi	7.6
41			
42			
43	1959	Hebgen lake	7.3
44			
45			
46	2008	Damxung	6.3
47	1990	Gonghe	6.4
48	2001	Kokoxili	7.8
49	2001	Kunlun	7.8
50	2008	Nima-Gaize	6.4
51			
52	2000	SISZ	6.5
53			
54		Northeast Iceland	
55	2008	Iwate-MiyagiNairiku	6.9
56	2001	NW India	7.6
57			
58			
59	2002	Denali	7.9
60			
61			
62	1964	Alaska	9.2
63	1957		9.1
64			
65			
66	1960	Valdivia, Chile	9.5
67			
68	1979	Ecuador-Colombia	
69		North Chile and South Peru	
70			
71	1905	Western Mongolia	6.8-8.4
72			
73		Central Nevada	

74	1997	Umbria-Marche, Italy	6.1
75			
76	1980	Irpinia	6.89
77			
78	1974	Northern Caribbean plate	6.1
79		Southeastern caribbean	
80	1989	Loma Prieta	6.9
81			
82	1991	Racha Georgia	7.15
83	1981	Southern Iran	> 6.4
84	1923	Kanto	7.9
85			
86	2001	Southern Peru	8.4
87	1995	Suez-Sinai, Jordan	7.2
88	1811-1812	New Madrid	

Table B.1: Global rheology database

Num	Rheology	Layering		Viscosity ¹ ($10^{18} Pa s$)
		Vertical	Lateral	
1	B	2	No	0.5, 10
2	M	2	No	9
3	P		Yes	$10 - 10^4 - 10^2 - 10^3$
4	M	3	No	4
5	M	3	No	20
6	M	2	No	9.3
7	M	2	No	(60 – 100)
8	M	2	No	8
9	B-M	4 ²	No	$1, 10 - 10^2 - 10^3$
10	P-M	4 ³	No	$((3 - 5) - 10) - 10^2 - 10^3$
11	B-M		No	$3 \times 10^3 - 10$
12	B	2 ⁴	Yes ⁵	0.5, (10 – 50)
13	M	2	No	0.5
14	B-M	4 ⁶	No	$0.4, 8 - 8 - 8 \times 10^2$
15	B-M	4 ⁷	No	$0.5, 10 - 10^2 - 10^3$
16	M	2 ⁸	No	35
17	M	2	No	2.5
18	M–M	3	No	3– ~ 1
	M–P–M			$3 - Pn = 3.5 - \sim 1$
19	M-B	3	Yes ⁹	$32 - 0.12, 3.2$
20	M	4	No	$\sim (10 - 10^2) - \sim (10^2 - 10^4) - \sim (1 - 10)$
21	SLS-M-M	4	No	$16 - 8 - 2.7$
22	Biviscous ¹⁰			$((3 - 4) \times 10^{-3} \text{ }^{11})$
23	M	2	Yes	$((0.1 - 2.9) \times 10^2)$
24	M-B	3	No	$32 - 0.16, 4.6$
25	M-B	3	Yes	$SW : 2 \times 10^2 - 2, 10$
				$NE : 2 \times 10^2 - 4, 40$
26	P		Yes	(50 – 10 ³)

¹Values for multi-viscous rheology are separated by “,” , for stratified layers by “-” and a single viscosity range by “()”

²High viscosity for the depth greater than 210 km

³High viscosity for the depth greater than 210 km

⁴We skip two other layers UM2 and LM

⁵The heterogeneity induced by the slab geometry

⁶Combined with afterslip

⁷Combined with afterslip

⁸We skip two other layers UM2 and LM

⁹Viscous lateral heterogeneity consistent with seismic structure

¹⁰Composite lower crust

¹¹transient creep within 4-34 days after Landers earthquake

27	M		No	1
28	M		No	36 – 12
29	M	2	No	~ 4
30	M	3	No	5 – 95
31	M	2	Yes	$(\sim 1 - 10^2)$
32	SLS-M	2	No	$\sim 2 - 70$
33	B	2	No	$(20 - 50) - 2 \times 10^2$
34	M	3	No	$> 5 \times 10^2$
35	P ¹²	2	Yes ¹³	0.2, 1
36	M	2	Yes ¹⁴	$(0.5 - 1)^{15}$
37	B	2	No	0.3
38	M	2	No	0.5
39	M	3 ¹⁶	Yes ¹⁷	0.5 – (5 – 10)
40	NN	3	No	$0.5 - 5 \times 10^5$
41	M	3	Yes	¹⁸ $\sim 20 - \sim 5$ ¹⁹ $\sim 3 \times 10^3 - \sim 20$
42	M	2	No	4
43	M	2	No	4
44	M	2	No	~ 10
45	M	2	No	2×10^3
46	M	2	No	1
47	M	2	No	90
48	M	2	No	20
49	M	2	No	1
50	M	2	No	0.3
51	SLS	2	No	(1 – 3)
52	M–B	3	No	$(5 - 10) - \sim 3$
53	M	3	No	$\sim 10 - \sim 1 - (3 - 4)$
54	M	3	No	30 – 3
55	M	2	No	(2.4 – 4.8)
56	P ²⁰	2	No	3 – 20

¹²The Maxwell rheology has been used with varying viscosity from 0.2 to 1 in four years which could be assumed equivalent to power-law viscosity

¹³Low viscosity mantle above the slab

¹⁴Low viscosity mantle above the slab

¹⁵Smaller than several subduction zones

¹⁶mid-crust and lower-crust viscoelastic layers

¹⁷Low viscosity region below the Central Range due to high temperature and rapid advection coincident with low seismic velocity and low V_p/V_s ratio.

¹⁸Southeast of the Hebgen Lake fault

¹⁹Central and north-western Hebgen Lake fault

²⁰Systematic increase of viscosity within six months

57	M	3	No	10 – 1
58	M	2	Yes	(10 – (3 – 4))
59	M-B	3	No	20 – 0.1, 2.8
60	M	2	No	(6 – 15)
61	M	2	No	(23 – 39)
62	M	2	No	(12 – 22)
63	M	2	No	(12 – 22)
64	M	2	No	10 ²
65	M	2	No	30
66	M	2	No	(80 – 10 ²)
67	M	2	No	25
68	M	3	No	10 – 10 ³
69	M	2	No	10
70	M	3	No	$3 \times 10^{-2} - 0.2 - (1 - 4)$
71	M	2	No	3
72	M	3	No	(5 – 50) – <i>NA</i>
73	M	3	No	10 ^{2.5} – 10 ²
74	M	4	No	1 – (0.1 – 1) – 10 ³
75	M	4	No	10 – 1 – 10 ³
76	M	2	No	10
77	M	6	No	6
78	M	3	No	30 – 10
79	M	2	No	10 ²
80	M	2	No	0.3
81	M	3	No	14 – ∞
82	M	2	No	0.9
83	M	3	No	1 – 10 ² (10 ² – 10)
84	M	2	Yes	14, 20
85	M	3	No	5
86	P	2	Yes	2.3×10^{-2} , 0.11
87	M	3	Yes	1
88		4	No	$(0.25 - 1.9) \times 10^2 - \infty - (2.5 - 19)$

Table B.2: Continuation of global rheology database: Viscoelastic models and viscosity values

Num	Layers (km)				HS ²⁵	Others
	SL ²¹	UP ²²	LC ²³	UM ²⁴		
1	30	30				Sun and Wang (2015)
2					50	Yamagiwa et al. (2015)
3						Muto et al. (2013)
4			40		90	Ueda et al. (2003)
5			33		163	Pollitz and Sacks (1994)
6					30	Suito and Hirahara (1999)
7					15	Ryder et al. (2014)
8					65	Gunawan et al. (2014)
9					40	Hoechner et al. (2011)
10					60	Reddy et al. (2010)
11	30	50				Pollitz et al. (2008)
12					62	Han et al. (2008)
13					70	Paul et al. (2007)
14					60	Panet et al. (2010)
15					65	Pollitz et al. (2006b)
16					62	Reddy et al. (2012)
17					65	Lubis et al. (2012)
18			10	22.5	~ 40	Rollins et al. (2015)
19			15	30	60	Johnson et al. (2007)
20			15		30	Pollitz et al. (2012)
21			16	30	50	Pollitz et al. (2000)
22						Ivins (1996)
23					~ (44 – 100)	Johnson and Segall (2004)
24			16		30	Pollitz (2003b)
25					43	Pollitz (2015)
26					50	Freed et al. (2012)
27					20	Bruhat et al. (2011)
28						Freed et al. (2007)
29					16	Rydelek and Sacks (2001)
30			11		25	Kenner and Segall (1999)

²¹Slab thickness

²²Upper plate thickness

²³Depth of Lower crust

²⁴Depth of upper mantle

²⁵Depth of half-space

31			12	Yamasaki et al. (2014)
32		20	35	Wang et al. (2009)
33		22	32	Hearn et al. (2009)
34		25	30	Hearn et al. (2002b)
35	62	62		Kogan et al. (2013)
36			62	Kogan et al. (2011)
37				Vladimirova et al. (2011)
38			62	Pollitz et al. (1998a)
39			30	Rousset et al. (2012a)
40		22	30.25	Sheu and Shieh (2004)
41		17	26	Chang et al. (2013)
		18	31	
42			38	Nishimura and Thatcher (2003)
43			38	Holdahl and Dzurisin (1991)
44			$\sim 30 - 40$	Reilinger (1986)
45			20	Koseluk and Bischke (1981)
46			15	Bie et al. (2014)
47			20.5	Hao et al. (2012)
48			15	Wen et al. (2012b)
49			32	Diao et al. (2011)
50			14	Ryder et al. (2010a)
51			15	Decriem and Árnadóttir (2012)
52		10	20	Árnadóttir et al. (2005)
53		10	$\sim (25 - 30)$	Jónsson (2008)
54		12	20	Pollitz and Sacks (1996)
55			$(19 - 23.5)$	Ohzono et al. (2012)
56			34	Chandrasekhar et al. (2009)
57		12	30N, 45S	Biggs et al. (2009)
58			50	Freed et al. (2006)
59		~ 15	38	Pollitz (2005)
60			45-85	Johnson et al. (2009)
61	90	30		Suito and Freymueller (2009)
62		20	80	Wahr and Wyss (1980)
63		50	200	Wahr and Wyss (1980)
64			46	Lorenzo-Martín et al. (2006)
65	30	40		Khazaradze et al. (2002)

66			18	Piersanti (1999)	
67	30	40		Hu et al. (2004)	
68		40	150	White et al. (2003)	
69			80	Casarotti and Piersanti (2003)	
70		35	50	Vergnolle et al. (2003)	
71			30	Calais et al. (2002)	
72		15	30	Hetland and Hager (2003)	
73				Hammond et al. (2009)	
74		20	35	Aoudia et al. (2003)	
75		10	20	35	Riva et al. (2000)
76		20			Dalla Via et al. (2005)
77			18.5		Dalla Via et al. (2003)
78		16	33		Pollitz and Dixon (1998)
79			35		Kozuch et al. (1996)
80					Linker and Rice (1997)
81		16	30		Pollitz et al. (1998b)
82		25	45		Podgorski et al. (2007)
83		~ 15	~ 30		Podgorski et al. (2007)
84			50		Pollitz et al. (2006c)
85			30		Matsu'ura and Iwasaki (1983)
86		45-65			Hergert and Heidbach (2006)
87		15	30		Piersanti et al. (2001)
88		16	33	50	Rydelek and Pollitz (1994)

Table B.3: Continuation of global rheology database: Range of layers in depths

Appendix C

Post-Dusky Sound supplementary files

Here we present cGPS displacement observations (blue lines) versus calculated afterslip coupled with viscoelastic rebound using the IDAV method (red lines) for a Maxwell rheology of viscosity 2×10^{18} Pa·s (See Section 4.4).

Figure C.1: cGPS displacement observations (blue lines) and estimated afterslip coupled with viscoelastic rebound (red lines) for a Maxwell rheology of viscosity 2×10^{18} Pa.s.

Appendix D

Post-Canterbury supplementary files

Figure D.1: Coseismic displacement due to the 13 June 2011 earthquake in Christchurch acquired from two descending TSX acquisitions between the 9th and 20th of June 2011. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

Figure D.2: Coseismic displacement due to the 23 December 2011 earthquake in Christchurch acquired from two descending TSX acquisitions between the 9th and 20th of June June 2011. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

Figure D.3: Standard deviation of mean line of sight velocity acquired from TSX acquisitions between 26 January 2012 and 22 May 2014. Larger values of standard deviation is due to the stratified atmospheric error over the area on the south-eastern corner of the case study with high gradient of elevation. We discard this area in our calculations. Red colours indicate movement away from satellite and blue colours indicate movement towards the satellite.

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