Spatiotemporal analysis of repeating earthquakes near Porangahau, Hikurangi Margin, New Zealand

By

OLIVIA DIANARA PITA SLLIM

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

Victoria University of Wellington

2021

Abstract

The Hikurangi subduction zone beneath the eastern North Island, New Zealand exhibits a variety of fault-slip related phenomena including tsunami earthquakes, non-volcanic tremor, low-frequency earthquakes, episodic slow slip, and repeating earthquakes. The northern Hikurangi margin hosts shallow slow-slip and is weakly coupled to shallow depths. In contrast, the southern Hikurangi margin is strongly coupled, and only deep slow-slip has been observed. The transition in coupling occurs beneath the township of Porangahau, and is an exemplary focus region for studying how this change in locking is accommodated.

To examine slip processes beneath Porangahau, we have constructed and analysed a catalogue of repeating earthquakes that occurred between 2004 and 2018. Repeating earthquakes are thought to re-rupture the same fault patch at different times, and thus have nearly identical waveforms, locations and magnitudes. Because repeating earthquakes represent cyclic loading, they can be used to detect temporal and spatial changes of slip-rate at depth and hence monitor how stress is transferred to seismogenic zones.

To build a catalogue of repeating earthquakes we first clustered the GeoNet earthquake catalogue by distance and correlation to identify potential repeating events. We then used a stronger cross-correlation threshold of at least 0.95 normalised cross-correlation value at three or more stations to identify repeating earthquakes from the initial clusters. This threshold was determined by our group's previous work on the northern Hikurangi margin. We identified 225 families of repeating earthquakes, with each family having two or more earthquakes in the 14-year study period from 2004 to 2018.

We carried out manual phase picking and polarity identification for the most recent event in each family and computed absolute locations, local magnitudes calibrated with moment magnitude, and high-quality focal mechanisms. For the rest of the events in each family, we conducted cross-correlation re-picking to obtain precise relative locations and relative magnitudes. With precise locations and well-constrained focal mech-

anisms, we determined whether the repeating earthquake families originated within the Pacific Plate, Australian Plate or on the subduction interface. Most of the families are located within the Pacific Plate, and the majority of families that originate on the subduction interface are located near the township of Porangahau. At least 220 of the 532 identified repeating earthquakes locate at the transition from strong- to weak-coupling of the subduction interface near the township of Porangahau.

A variety of slow slip events have been detected near Porangahau in the last two decades. Even though some repeating earthquakes correlate spatially and temporally with slow slip events, temporal and spatial correlations between slow slip events and repeating earthquakes are scarce and sparse. The majority of repeating earthquakes are located up-dip or down-dip of modelled slow slip patches, with very few families having spatial correlation with slow slip events. We obtained a moment-recurrence interval relationship for the catalogue of repeating earthquakes near Porangahau and compared it to the relationship obtained by Nadeau and Johnson (1998) at Parkfield, California. Finally, we computed slip-rates using the families located on the subduction interface and obtained an average slip-rate of 13 mm/yr. The insights gained from this study lay the groundwork for future work constraining processes of strain accumulation at the creeping-to-locked transition zone near Porangahau.

Acknowledgements

I would like to thank my supervisors Calum Chamberlain and John Townend for their guidance, support and help throughout this process. I am grateful for your patience and I appreciate the time and energy you have put into this project. I have learned so much in this past year from you, from coding to what to do in windy Wellington. Thank you for the amazing fieldwork opportunity on the West Coast and the encouragement to present my work at GSNZ and AGU. I also thank Laura Wallace for the opportunity to go on a voyage to offshore Porangahau, the stars at night and sunrises were breathtaking. I thank the examiners Rupert Sutherland and Katie Jacobs for their comments and suggestions, which have helped me to improve this thesis.

This project was funded by a Marsden grant (Project ID: 17-VUW-121), the Frank Evison Research Scholarship in Geophysics, and the EQC Programme in Seismology & Tectonic Geodesy.

Thank you to all my amazing friends in New Zealand for all their support and amazing moments shared together. A huge thank you to the uni crew, Carmen, Yaas, Theo, Emily, Brenton, Sam, Josh and Calvin, for all the lunch time company, the fun nights in town and the occasional beers. Thank you to latinos @VUW, Andrea, Guillermo, Javier, Jenny, William and Daniel, for the fun trips, the company at the postgrad association breakfasts and all the great moments spent together around Wellington! And of course, special thanks to Carmen for being such an amazing friend and a huge support throughout this year!!

Thank you to all the Ollin Yolitzlti dance group for giving me the opportunity to go back to dancing, thank you for all the hugs and the unforgettable performances. Can't wait to fully return to rehearsals and enjoy dancing again with all of you.

To my parents, thank you for all the encouragement and support you have always given me. Thank you for supporting my dream of becoming a seismologist, even if that meant moving to the other side of the world. To Alfredo, thank you for always being there for me, even though we are thousands of kilometres apart. Siempre los llevo en lo profundo de mi corazón, pronto nos veremos.

Gracias, gracias, gracias.

Contents

Abstract	iii
Acknowledgments	v
List of Figures	xi
List of Tables	xiii
1 Introduction	1
1.1 Motivation and objectives	2
1.2 Study location: Porangahau	4
1.2.1 The Hikurangi margin	4
1.2.2 Seismic and aseismic phenomena near Porangahau	5
1.3 Thesis structure	9
2 Background	11
2.1 Repeating earthquakes definition	11
2.1.1 Nomenclature	12
2.2 Identification of repeating earthquakes	13
2.2.1 Identification using waveform similarity	14
2.2.2 Identification using other criteria	16
2.3 Mechanics of repeating earthquakes	17
2.3.1 Earthquake cycle models	17
2.3.2 Recurrence intervals of repeating earthquakes	20
2.4 Repeating earthquakes in different settings	21
3 Data and Methodology	25

3.1	Data	26
3.2	Repeating earthquake identification	28
3.3	Clustering into repeating earthquake families	32
3.4	First arrivals and polarity identification	33
3.4.1	Pick corrections	35
3.5	Location of repeating earthquakes	37
3.5.1	Absolute locations of repeating earthquakes	38
3.5.2	Relative locations of repeating earthquakes	38
3.6	Local magnitude calculation	40
3.7	Focal mechanism calculation	43
4	Results	45
4.1	Nomenclature of repeating earthquake families	45
4.2	Repeating earthquake locations	47
4.2.1	Absolute locations	47
4.2.2	Relative locations	49
4.3	Repeating earthquake magnitudes	52
4.4	Repeating earthquake focal mechanisms	55
4.5	Results summary	59
5	Discussion and Conclusions	61
5.1	Repeating earthquake catalogue completeness	61
5.2	Repeating earthquake locations	68
5.2.1	Families in the Australian Plate	68
5.2.2	Families in the Pacific Plate	70
5.2.3	Families on the subduction plate interface	71
5.3	Repeating earthquakes and other phenomena	71
5.4	Seismic moment–recurrence interval relationship	84
5.5	Slip–rate relationships	86
5.6	Conclusions and future work	90

A	Repeating Earthquake Catalogue Information	108
A.1	Catalogue of repeating earthquakes near Porangahau	108
A.2	Stations corrections from magnitude inversion	115
A.3	Focal Mechanism Solutions	117
A.4	Focal Mechanism Plots	120
B	Earthquake Location Input Files	137
B.1	NonLinLoc input file	137
B.2	GrowClust input file	139

List of Figures

1.1	Tectonic setting of New Zealand	6
1.2	Map of seismic phenomena in the North Island	7
1.3	Map of aseismic phenomena in the North Island relevant to this study.	8
2.1	Schematic model of repeating earthquakes on a creeping fault with locked area.	12
2.2	Diagram of recurrence interval and family duration of repeating earthquake families.	13
2.3	Recurrence model of an idealised earthquake cycle.	17
2.4	Recurrence models of time-Predictable and slip-predictable models . . .	18
2.5	Asperity model of repeating earthquakes	19
2.6	Repeating earthquake studies in New Zealand	24
3.1	Flow diagram outlining the steps undertaken in this thesis.	25
3.2	Location of the stations used in this research	27
3.3	Waveform pre-processing and multisegment normalized cross-correlation.	30
3.4	Correlated waveforms of the same family on two different stations. . . .	31
3.5	Dendrogram of an initial family showing the final clusters obtained . .	32
3.6	P and S phase identification	34
3.7	Correction of the P and S phase arrivals using cross-correlation	36
3.8	Map with the locations of earthquakes used to calibrate the local magnitudes	42
3.9	Diagram illustrating different take-off angles due to the uncertainties in P-wave arrival times.	44
4.1	Map of the geographical groups of the repeating earthquakes families. .	46
4.2	Map and cross-sections of the absolute locations obtained for the most recent events of each family using NonLinLoc	48

4.3	Map and cross-sections of the hypocentral probability density function computed with NonLinLoc.	50
4.4	Map and cross-sections illustrating the final location of all the repeating earthquakes, containing both absolute and relative locations.	51
4.5	Calculated local magnitudes of all the events through the 12 years study period.	53
4.6	Local magnitude changes through time of families with five or more events.	54
4.7	Histogram illustrating the difference between local magnitudes calculated and the mean of the family.	54
4.8	Stereonets and P-T contour plots of two events from the focal focal mechanisms catalogue.	57
4.9	Map and cross-sections of the selected focal mechanisms coloured using a ternary diagram classification	58
5.1	Magnitude of completeness of the GeoNet catalogue near Porangahau from 2004 to 2020.	62
5.2	Amplitude spectrograms and signal-to-noise ratios for three earthquakes of different local magnitudes.	66
5.3	Calculated source radii of the repeating earthquakes using constant values of stress drops of 3 and 10 MPa	67
5.4	Map illustrating the locations of the repeating earthquake families in relation to the the interface locking model and the cumulative slip of slow slip events.	69
5.5	Local magnitudes of continual- and burst-type repeating events through time with seismic and aseismic phenomena.	73
5.6	Cumulative moment of continual- and burst-type repeating events through time with seismic and aseismic phenomena.	74
5.7	Time-series and map illustrating the repeating earthquakes that occur close in time to the 2009 Riversdale slow slip event	77
5.8	Time-series and map illustrating the repeating earthquakes that occur close in time to the 2016 triggered East Coast slow slip event	79
5.9	Time-series and map illustrating the repeating earthquakes that occur close in time to the 2010/2011 Manawatu slow slip event	81
5.10	Time-series and map illustrating the repeating earthquakes that occur close in time to the 2008 Kaimanawa slow slip event	82
5.11	Relationship between moment and recurrence interval for repeating earthquake families near Porangahau and other published studies.	85
5.12	Box plot illustrating the distribution of slip-rate values obtain from repeating earthquakes near Porangahau	88

List of Tables

4.1	Summary of errors associated with the absolute locations obtained using NonLinLoc.	49
4.2	Summary of standard errors associated with the relative locations obtained using GrowClust.	52
5.1	Dates and locations of the seismic and aseismic phenomena to which we are spatially and temporarily comparing the repeating earthquake catalogue	75
5.2	Summary of the seismic moment–recurrence interval relationship identified for repeating earthquakes near Porangahau.	86
5.3	Slip-rates calculated from the slip-seismic moment relationship derived by Nadeau and Johnson (1998).	89
A.1	Summary of the repeating earthquake families identified near Porangahau in this study.	108
A.2	Stations corrections from magnitude inversion.	116
A.3	Focal Mechanism Solutions	117

Chapter 1

Introduction

Geodetic methods provide a means of monitoring the inter-seismic strain accumulation across faults (Avouac, 2015), but do not provide direct on-fault evidence of strain accumulation, or the rate of strain accumulation at seismogenic depths. However, repeating earthquakes may provide a way of monitoring slip-rates and the changing strain and stress fields of seismogenic zones (Nadeau and Johnson, 1998; Nadeau and McEvilly, 1999). Repeating earthquakes are thought to re-rupture the same patch of fault (Uchida et al., 2003). The cumulative slip from each rupture represents the slip in the region surrounding the patch. For this reason, repeating earthquakes can be used to quantify temporal and spatial changes of slip rates at depth and monitor the changing strain field of seismogenic zones (Nadeau and McEvilly, 1999).

The Hikurangi margin, located east of the North Island, New Zealand, exhibits a variety of seismic and aseismic phenomena including tsunami earthquakes, tremor, slow-slip events and repeating earthquakes (Wallace et al., 2009). Porangahau is a particularly intriguing area as it is located in the transition between creeping and locked zones beneath the North Island (Wallace et al., 2012b; Hughes et al., 2021). In this thesis, we ¹ construct and analyse a catalogue of repeating earthquakes near Porangahau from 2004 to 2018 to provide a starting point for monitoring strain accumulation at the creeping to locked transition zone.

In this chapter we first introduce the motivations and objectives of this research in Section 1.1. In Section 1.2 we provide an overview of the tectonic setting of the Hikurangi margin (Section 1.2.1) and the seismic and aseismic phenomena observed near

¹This thesis documents original research conducted by me, Olivia Dianara Pita Sllim, under the supervision of Dr. Calum John Chamberlain and Prof. John Townend. To reflect standard scientific practice and my supervisors' input throughout the project and to facilitate publication of these results in a peer-reviewed journal, this thesis is written in the first person plural ("we").

and surrounding Porangahau (Section 1.2.2). The final section of this chapter outlines the structure of the rest of this thesis (Section 1.3).

1.1 Motivation and objectives

Geodetic observations have shown variations in interface interseismic coupling at the Hikurangi subduction margin (Wallace et al., 2004, 2009, 2012b). Interseismic coupling is defined as the ratio of slip deficit during the interseismic period divided by the long term slip that would have happened if the fault had been slipping at the long-term average slip rate (Kaneko et al., 2010). A coupling coefficient of 1 means there is no relative motion between the two sides of a fault and they are “locked” together. In contrast, a coupling coefficient of 0 corresponds to “creeping” at the long-term slip rate. Observations suggest that large interplate earthquakes tend to rupture regions that were locked and accumulating strain during the interseismic period (Wallace et al., 2004; Bürgmann et al., 2005; Loveless and Meade, 2011; Avouac, 2015).

The subduction interface beneath southern North Island has a high interseismic coupling coefficient, meaning that the Pacific and Australian Plate are locked together. The coupling coefficient decreases northward along the margin, where the subduction interface is creeping or freely sliding. The transition in coupling occurs in southern Hawkes Bay, making this area an exemplary region for studying how this along-strike change in locking is accommodated. In addition to this, the Hikurangi margin has not had a subduction interface earthquake larger than M_w 7.2 since historical records began in $\sim$ A.D. 1840 (Doser and Webb, 2003; Wallace et al., 2009; Clark et al., 2019), and the interface may produce earthquakes as large as M_w 9 (Stirling et al., 2012). Understanding how this region is loaded may be useful for understanding earthquake nucleation processes in general and the seismogenic potential of the Hikurangi margin.

Repeating earthquakes are one group of seismic events with the peculiarity of having nearly-identical waveforms, locations and magnitudes (Geller and Mueller, 1980; Vidale et al., 1994; Nadeau et al., 1994). These earthquakes are interpreted as the repeated rupture of the same patch caused by stress accumulation on the locked patch due to aseismic slip on the surrounding area (Nadeau et al., 1995; Beeler et al., 2001; Uchida and Bürgmann, 2019). Under this assumption, the cumulative slip of repeating earthquakes is assumed to be equal to the fault creep and provide a direct *in situ* measurement of on-fault deformation at seismogenic depths (Nadeau and McEvilly, 1999; Igarashi et al., 2003; Uchida, 2019). For example, Uchida and Matsuzawa (2011)

used repeating earthquakes to better constrain the spatial distribution of interplate coupling along the Japan Trench before the M9.0 Tohoku earthquake, and found that the seismic coupling in the area was stronger than previously thought.

Extensive research of repeating earthquakes has been carried out at other plate boundary zones, particularly at the San Andreas Fault system (Nadeau et al., 1994, 1995; Nadeau and Johnson, 1998; Schaff et al., 1998; Chen and Lapusta, 2009), the Japan Trench (Uchida et al., 2003; Igarashi et al., 2003; Kimura et al., 2006; Mavrommatis et al., 2015; Gardonio et al., 2018), Taiwan (Rau et al., 2007; Chen et al., 2008) and China (Schaff and Beroza, 2004; Schaff and Richards, 2011). In New Zealand, repeating earthquakes remain relatively unstudied. The first decadal-scale catalogue of tectonic repeating earthquakes in New Zealand was recently obtained by Hughes et al. (2021), along with a composite criterion for identifying repeating earthquakes. Wallace et al. (2017) constructed a repeating earthquake catalogue around Porangahau using less-strong criteria than Hughes et al. (2021), who demonstrated the need for broad-band widths and waveforms including significant coda.

This research aims to further develop the catalogue of repeating earthquakes in New Zealand and to lay the basis for monitoring fault slip rates at depth in the transition from creeping to locked near Porangahau using repeating earthquakes. This project addresses the following four objectives:

1. Detect repeating earthquakes near Porangahau using cross-correlation techniques using a clustered GeoNet catalogue created by Chamberlain et al. (2019) and a threshold for identification of repeating earthquakes in New Zealand determined by Hughes et al. (2021).
2. Create a catalogue of repeating earthquakes near Porangahau containing precise locations and seismic parameters, including recurrence interval, focal mechanism, magnitude and seismic moment.
3. Analyse the spatial and temporal distribution of the repeating earthquakes in relation to other seismic and aseismic phenomena.
4. Analyse the relationship between seismic moment and recurrence interval and calculate slip-rates for repeating earthquakes located on the subduction interface using the parameters obtained via Objective 2.

1.2 Study location: Porangahau

The study area of this thesis encompasses the region between Cape Kidnappers and Castlepoint, including Porangahau on the east coast of North Island, New Zealand. In this section, we describe the regional tectonic setting and the observed seismic and aseismic phenomena occurring in the region.

1.2.1 The Hikurangi margin

New Zealand is located on the boundary between the Australian and Pacific Plates (Figure 1.1). Beneath the North Island, the Pacific Plate is subducting under the Australian Plate at the Hikurangi margin. This subduction margin transitions into a transpressional regime at northern South Island, forming the Marlborough faults which coalesce into the Alpine Fault: the main plate-boundary fault through South Island. Further south, plate-motion is accommodated by subduction of the Australian Plate beneath the Pacific Plate at the Puysegur margin (Mortimer et al., 2017).

The active tectonics in the North Island are dominated by subduction at the Hikurangi margin, back-arc rifting in the Taupō Volcanic Zone (TVZ), and strike-slip faulting in the North Island dextral fault belt (NIDFB) (Beanland and Haines, 1998; Wallace et al., 2004, 2009). Subduction at the Hikurangi margin began at approximately 24 Ma (Ballance, 1976; Kamp, 1999; Nicol et al., 2007) with the subduction of the Hikurangi Plateau, a Cretaceous basaltic oceanic plateau (Mortimer and Parkinson, 1996).

The forearc of the eastern North Island is currently rotating clockwise at $\sim 4^\circ/\text{Myr}$ relative to the Australian plate (Wallace et al., 2004; Lamb, 2011). Through the inversion of 10–15 years of global navigation satellite system (GNSS) data, Nicol and Wallace (2007) determined that the majority of plate convergence is principally accommodated on the plate interface, while most of the margin-parallel motion is accommodated by strike-slip faulting and tectonic rotation of the Australian Plate. Lamb (2011) reconstructed the plate motion using paleomagnetic measurements and found that since 4 Ma the rotation has increased in the northern part of the Hikurangi Margin, with $20\text{--}25^\circ$ of clockwise rotation near Hawkes Bay and $\sim 10^\circ$ further south in Wairarapa. A variation of the convergence rate along the Hikurangi Trough is observed, from $\sim 60\text{ mm/yr}$ in the north to $\sim 20\text{ mm/yr}$ in the south (Wallace et al., 2004; DeMets et al., 2010).

There are also changes in the characteristics of the subduction margin offshore the east coast of the North Island. At the northern end of the Hikurangi margin, the

Hikurangi Plateau is $\sim 4\text{--}9$ km thick, the sedimentary sequence overlaying the plateau is less than 1 km thick, and the accretionary wedge is less than 20 km wide (Davy and Wood, 1994; Wallace et al., 2009; Barker et al., 2009; Bassett et al., 2010). In comparison, at the central and southern Hikurangi margin, the Hikurangi Plateau, the sedimentary sequence, and the accretionary wedge are thicker: $\sim 12\text{--}15$ km, ~ 5 km, and 30–70 km respectively (Davy and Wood, 1994; Mochizuki et al., 2019; Wallace et al., 2009; Barnes et al., 2010; Barker et al., 2009). Barker et al. (2009) suggested that the northward reduction in thickness combined with the increase in convergence rate, lead to differences in dip and kinking of the subduction interface. At the northern Hikurangi margin, the subduction thrust near the coast has a kink downward at angles of $>8^\circ$ at depths of 10–15 km, and towards the southern margin the dip of the subduction is $<8^\circ$ for the same depths (Barker et al., 2009).

1.2.2 Seismic and aseismic phenomena near Porangahau

The Hikurangi subduction margin has not produced a subduction interface earthquake larger than M_w 7.2 since historical records began in $\sim$ A.D. 1840 (Wallace et al., 2009), although there is stratigraphic evidence of previous subduction earthquakes (Clark et al., 2015, 2019). The January 2014 M_w 6.3 normal-faulting Eketāhuna earthquake (Wallace et al., 2014) is the only earthquake larger than M6.0 to have occurred near Porangahau in the last 21 years (red stars in Figure 1.2), but there were three large earthquakes during the twentieth century in the region (yellow stars in Figure 1.2). Two of these earthquakes were part of the 1990 Weber sequence, one involving oblique-normal faulting within the subducting Pacific Plate (M_w 6.2) and the other thrust-faulting within the Australian Plate (M_w 6.4) (Robinson, 1994; Webb and Anderson, 1998; Doser and Webb, 2003). The third twentieth century earthquake larger than M6.0 is the mainshock of the Hawkes Bay sequence of 1931 of M_w 7.4–7.6 and involving strike-slip faulting in the Australian Plate (Hull, 1990; Doser and Webb, 2003).

Several slow slip events (SSEs) have been recorded on the Hikurangi margin since the installation of GNSS in New Zealand (see Wallace (2020) for a review). The characteristics of SSEs on the Hikurangi margin change along-strike: The northern and central Hikurangi margin presents shallow (<15 km), short (<1 month) and frequent slow-slip, while in the southern margin deep (>30 km), long (>1 year) and less frequent slow slip events have been observed (Wallace, 2020; Ikari et al., 2020, and references therein).

In the region surrounding Porangahau, SSEs with diverse characteristics have been

Figure 1.1: Tectonic setting of New Zealand showing the Australian-Pacific plate boundary zone. Vectors show the motion of the Pacific Plate relative to the Australian Plate (DeMets et al., 2010). The township of Porangahau is marked with a red triangle. Cape Kidnappers, Cape Turnagain and Castlepoint are marked with pink, orange and purple diamonds, respectively. TVZ – Taupo Volcanic Zone, HB – Hawkes Bay, NIDFB – North Island Dextral Fault Belt. Faults from The New Zealand Community Fault Model (Van Dissen et al., 2021).

Figure 1.2: Map illustrating the seismic phenomena in the North Island, New Zealand. The study area is indicated with the red circle. Circles in the background represent the seismicity in 2014 catalogued by GeoNet¹. Red stars represent the earthquakes above magnitude 6.0 that have been recorded by GeoNet since 2000². Yellow stars represent large historical earthquakes near the study of area mentioned in the text. Red contours represent the subduction interface depth model of Williams et al. (2013).

¹ <https://quakesearch.geonet.org.nz/csv?bbox=172,-42,179,-34.5&maxdepth=300&startdate=2014-01-01T0:00:00&enddate=2015-01-01T0:00:00>

² <https://quakesearch.geonet.org.nz/csv?bbox=172,-42,179,-34.5&minmag=6&startdate=2000-01-01T0:00:00&enddate=2021-01-01T0:00:00>

Figure 1.3: Map illustrating the aseismic phenomena in the North Island relevant to this study. Red contours indicate the cumulative slip of slow slip events in mm, modified from Wallace (2020). The blue colours of the interseismic coupling represent a low coupling coefficient and red colours represent high coupling coefficient (Wallace et al., 2012b). Green and yellow shaded areas indicate areas of observed ambient tremor (Romanet and Ide, 2019) and high reflectivity materials (Barker et al., 2009), respectively.

observed (Figure 1.3). Between Cape Turnagain and southern Hawkes Bay, several shallow slow-slip events have been observed, which we refer to here as East Coast slow slip events.

SSEs south of Hawkes Bay occur every 1–2 years and SSEs offshore Cape Turnagain every ~ 5 years. The duration of the events is usually 2–3 weeks with magnitudes equivalent to M_w 6.1–6.8 (Ikari et al., 2020). One shallow SSE that occurred offshore from Porangahau is the 2009 Riversdale SSE (Wallace et al., 2012b). There is little information available regarding this event, but Jacobs et al. (2016) observed an increase in seismicity on the southern edge of the slow-slip patch, outside our study area.

Further inland and west of Porangahau, deep (>25 km) Manawatu SSEs occur every 4–5 years and have a duration of 1–2 years. These deep Manawatu events occur in different stages and represent total moment release equivalent to $M_w \sim 7.1$. Three Manawatu events have been observed, in 2004/2005, 2012/2011 and 2014/2015 (Wallace and Beavan, 2010; Wallace et al., 2018; Wallace, 2020). The Kaimanawa SSEs are also included in this category of deep SSEs. The Kaimanawa SSEs occurred in 2006 and 2008 beneath the Kaimanawa Ranges, down-dip of the shallow East Coast SSEs and lasted between 2–3 months Wallace and Eberhart-Phillips (2013).

Slow slip events have been observed in conjunction, in space and time, with tectonic tremor near or offshore Porangahau. Romanet and Ide (2019) observed tremor occurring offshore Cape Turnagain during three East Coast slow slip events, in 2006, 2011 and 2016. Triggered (Fry et al., 2011) and ambient tremor (Ide, 2012; Romanet and Ide, 2019) have also been observed related to the deep Manawatu SSEs, although the area of observation is north of our area of study.

The diverse slip behaviour that occurs at the transition from locking to creeping of the plate interface beneath Porangahau provides an intriguing environment to study the interplay between seismic and aseismic processes.

1.3 Thesis structure

This thesis consists of five chapters including this Introduction. The structure of the remaining chapters is as follows:

Chapter 2. Background

Chapter 2 contains a review of the prevailing theory regarding how repeating earth-

quakes are generated and includes: the definition and nomenclature of repeating earthquakes, the different criteria for identification, the mechanical processes underlying their occurrence, and the observations of repeating earthquakes in different settings.

Chapter 3. Data and Methodology

Chapter 3 describes the data used and the methodology followed to identify repeating earthquakes. This chapter also covers the methods used to calculate locations, magnitudes and focal mechanisms of repeating earthquakes.

Chapter 4. Results

Chapter 4 presents the main results of this research: a catalogue of repeating earthquakes near Porangahau.

Chapter 5. Discussion and Conclusions

Chapter 5 reviews the catalogue of repeating earthquakes in the context of the literature and existing knowledge of the region near Porangahau. At the end of the chapter, concluding statements are presented.

Appendices

Appendix A.1 contains a table with the families that each geographical group contains.

Appendix A.2 lists the station corrections from the local magnitude inversion.

Appendix A.3 contains the final focal mechanism solutions and their corresponding calculated error.

Appendix A.4 contains the stereonets and PT axis contour plots of the final focal mechanism catalogue.

Appendix B.1 contains the input file for NonLinLoc.

Appendix B.2 contains the parameters used as input for Growclust.

Chapter 2

Background

This chapter reviews the definition and theoretical understandings of repeating earthquakes (Section 2.1) and their nomenclature (Section 2.1.1). Section 2.2 outlines the different criteria used to identify repeating earthquakes. Mechanics and models regarding repeating earthquakes are presented in Section 2.3. Studies of repeating earthquakes in the world and New Zealand are reviewed at the end of this chapter in Section 2.4.

2.1 Repeating earthquakes definition

Seismograms represent the time-varying seismic wavefield recorded at a point by a seismograph. The resulting seismogram can be thought of as the output of a source signal convolved with a number of linear filters. Therefore, a seismogram u can be described as a source signal (x) convolved with effects of earth structure (g) and the response of a seismometer (i) (Stein and Wysession, 2003):

$$u(t) = x(t) * g(t) * i(t) \quad (2.1)$$

Repeating earthquakes are groups of earthquakes with nearly identical waveforms (u), source locations and fault geometries (Aster and Scott, 1993; Nadeau et al., 1994, 1995; Uchida, 2019). Neglecting any changes in instrument characteristics (i.e. $i(t)$) and minimal variations in path, nearly identical locations and fault geometries are interpreted as the repeated rupture of the same fault patch at different times (Geller and Mueller, 1980; Poupinet et al., 1984; Vidale et al., 1994; Nadeau et al., 1994, 1995).

This process is represented in Figure 2.1, where two almost identical waveforms are

Figure 2.1: Schematic model of repeating earthquakes on a creeping fault with locked area (patches). Waveforms of two repeating earthquakes obtained in this research are shown in red and blue. The red seismogram was recorded in 2018 and the blue one in 2010, both were recorded at station PRHZ located in the township of Porangahau. Below the blue waveform, the components of a seismogram are described. Figure modified from Uchida and Bürgmann (2019).

assumed to have nearly identical source processes (x) and then go through the same medium (g) until reaching the seismometer. For this reason, repeating earthquakes are used to examine temporal and spatial variations in source processes (e.g. Uchida et al., 2007), medium (e.g. Poupinet et al., 1984), and fault zone properties (e.g. Nadeau et al., 1994; Igarashi et al., 2003).

2.1.1 Nomenclature

A *family* is one of the names given to a group of events that have met a specific criterion and declared to be repeating earthquakes. The terms *Doublets*, *triplets* and *multiplets* are also used (e.g. Poupinet et al., 1984; Nadeau et al., 1994) and indicate more directly the number of events that form a group. The terms *sequence* or *cluster* are also used in repeating earthquake studies (e.g. Nadeau et al., 1994; Li et al., 2007; Templeton et al., 2008; Igarashi et al., 2003; Uchida and Bürgmann, 2019). Sequence or cluster make no reference to a number of events but they have been used to indicate groups of events that are not repeating earthquakes. For this reason, in this thesis we use the term *family* as it can be used no matter the number of events and it is not commonly used outside the repeating earthquakes context.

Families of repeating earthquakes are formed by two or more events. The *family*

Figure 2.2: Components of a repeating earthquake family. Red circles are earthquakes belonging to the same family. The difference between recurrence interval and family duration is illustrated. Modified from Hughes et al. (2021)

duration is the time difference between the *most recent event*, or *youngest*, and the *first recorded earthquake*, or *oldest* (Figure 2.2). The family duration is influenced by the duration of the catalogue analysed and the magnitude of completeness of the catalogue analysed (Hughes et al., 2021). Family duration is not to be confused with *recurrence interval*, which indicates the average time between events also referred to as inter-event time (Mesimeri and Karakostas, 2018; Hughes et al., 2021). As illustrated in Figure 2.2, for example, events occur every two years (e.g. 2010, 2012, 2014...) so the recurrence interval of the family is ~ 2 years and the duration of the family is 10 years.

Burst-type and **continual-type** repeating earthquake families are categories formed based on the family duration and recurrence interval. Burst-type families are characterised by having short recurrence interval times, lasting from a few hours to a few days (Igarashi et al., 2003; Templeton et al., 2008; Mesimeri and Karakostas, 2018) and as a result also have short family durations that are often less than a year, but which vary among authors: e.g. Yamashita et al. (2012) established a family duration of less than three years and Mesimeri and Karakostas (2018) of one year to define burst-type families. On the other hand, continual-type families have longer recurrence intervals and family durations compared to burst-type (Igarashi et al., 2003; Mesimeri and Karakostas, 2018).

2.2 Identification of repeating earthquakes

Repeating earthquakes have been identified using different criteria including waveform similarity, hypocentre location, magnitude, and time interval between events (e.g. Geller and Mueller, 1980; Poupinet et al., 1984; Rubin et al., 1999; Gibowicz

and Lasocki, 2007). There is not an established single criterion to identify repeating earthquakes as different tectonic settings, networks, instruments, and periods of time are analysed. In this section we review the rationale behind the dominant classification metrics, waveform similarity, and discuss other criteria.

2.2.1 Identification using waveform similarity

One of the most used criteria for identifying repeating earthquakes is waveform similarity (e.g. Nadeau et al., 1995; Igarashi et al., 2003; Schaff and Beroza, 2004; Li et al., 2011; Hughes et al., 2021). The cross-correlation coefficient (CC) is used to measure the degree of similarity between two signals by performing a normalized cross-correlation as shown in equation 2.2:

$$CC_{xy}(l) = \frac{1}{N} \sum_{n=0}^{N-1} f_x(n) f_y(n-l) \quad (2.2)$$

where f_x and f_y represent two signals (waveforms), l is the lag time shift between f_x and f_y , and N is the number of samples (Proakis and Dimitris, 1990; Rao and Swamy, 2018). Equation 2.2 is defined as a “normalized” cross-correlation because it is scaled by a normalization factor, which in this case is $\frac{1}{N}$ (Proakis and Dimitris, 1990). There are other normalization techniques, such as using the the square root of the energy of one of the signals to normalize both of them or to remove the mean from both signals (Martin and Crowley, 1995). In this thesis we use a normalized cross-correlation implemented in EQcorrscan (Chamberlain et al., 2018):

$$c(t) = \frac{\sum_y (a_y - \bar{a})(b_{t+y} - \bar{b}_t)}{\sqrt{\sum_y (a_y - \bar{a})(a_y - \bar{a}) \sum_y (b_{t+y} - \bar{b}_t)(b_{t+y} - \bar{b}_t)}} \quad (2.3)$$

where $c(t)$ is the normalized cross-correlation at time t and the two waveforms being cross-correlated are represented as a and b . This normalized cross-correlation function also removes the mean from b at every time-step, which reduces the effects of jumps in average amplitude (Chamberlain et al., 2018). Chapter 3 contains the details regarding the process followed in this study to identify repeating earthquakes using normalized cross-correlation.

Several studies have used cross-correlation to identify repeating earthquakes (e.g. Geller and Mueller, 1980; Nadeau et al., 1994; Igarashi et al., 2003; Uchida et al., 2003; Schaff and Beroza, 2004; Rau et al., 2007; Zhang et al., 2008; Yu, 2013; Schmittbuhl et al.,

2016; Hughes et al., 2021). The minimum cross-correlation threshold varies from study to study, but it typically ranges from 0.90 to 0.98 (Uchida and Bürgmann, 2019) and can be as low as 0.75 (Rubinstein et al., 2012).

The operational parameters such as window length and filter frequency band have a significant effect on the correlation values obtained (Gao and Kao, 2020; Uchida, 2019). Window lengths range from seconds to minutes, and this choice typically depends on the distance between the earthquake and the recording station, making the lengths unique for each waveform (Gao and Kao, 2020, and references therein). When using longer time windows, more phases are being compared and may contain low signal-to-noise ratio coda waves. In this instance a high cross-correlation threshold may be too strict a criterion and events may be missed (Uchida, 2019). If shorter time windows are used, high cross-correlation values are statistically less robust and large-amplitude phases may bias the correlations and result in the detection of events that are not really repeating earthquakes (Gao and Kao, 2020; Uchida and Bürgmann, 2019). Using a fixed time window for all the waveforms is not ideal, due to the length of seismic wave trains increasing with source-receiver distance (Gao and Kao, 2020).

Selecting an appropriate filter frequency-band is also important, as waveforms are frequently filtered to enhance certain features and suppress others such as noise (Havskov and Ottemoller, 2010). Gao and Kao (2020) examined the effects of using different filter frequency-bands to compute the cross-correlation value and found that the cross-correlation coefficient value drops significantly once the dominant seismic energy (in the 1–2 Hz frequency band) is filtered out. Furthermore, a low frequency pass-band may not yield sufficient spatial resolution to identify neighbouring events (Uchida, 2019; Gao and Kao, 2020). Geller and Mueller (1980) suggested that for events to have nearly-identical waveforms, they must be located within a quarter wavelength.

Some other researchers (e.g. Geller and Mueller, 1980; Poupinet et al., 1984; Lengliné and Marsan, 2009; Uchida et al., 2009; Rubinstein and Ellsworth, 2010; Mavrommatis et al., 2015) have used coherence as a measure of waveform similarity to classify repeating earthquakes. Coherence (coh) is a measure of similarity between two signals in the frequency domain (ω):

$$\text{coh}_{xy}(\omega) = \sqrt{\frac{|S_{xy}(\omega)|^2}{S_{xx}(\omega)S_{yy}(\omega)}} \quad (2.4)$$

where S_{xy} is the cross-spectrum between earthquakes x and y obtained after computing the Fourier transform of the original waveforms f_x and f_y (Uchida, 2019; Uchida and Bürgmann, 2019). For two events to be selected as repeating earthquakes, the coherence

value must be higher than a minimum threshold, which is usually 0.95, across a range of frequencies (Uchida, 2019; Uchida et al., 2006; Uchida and Bürgmann, 2019). In most cases, a mean coherence range is obtained for a frequency range which is commonly 1–8 Hz (Uchida, 2019).

In most previous studies of repeating earthquakes, either the cross-correlation or coherence threshold must be exceeded at several stations for two events to be identified as repeating earthquakes. The required minimum number of stations is usually two, although other studies have used only one (Schaff and Beroza, 2004) and as many as five (Lengliné and Marsan, 2009).

To reduce computation time, cross-correlations or coherence calculations are normally made only for events that already passed a simple preliminary criterion (Aster and Scott, 1993), usually based on either distance or magnitude. For example, Uchida et al. (2003, 2006); Igarashi et al. (2003); Materna et al. (2018) only computed cross-correlation or coherence for pairs of events less than 30 km apart. Similarly Igarashi et al. (2003) required a preliminary magnitude threshold to be exceeded, and only compared events with magnitudes larger than 3.

2.2.2 Identification using other criteria

Although waveform similarity is the most common criterion to identify repeating earthquakes, it is not the only one that has been used. Since repeating earthquakes are thought to rupture the same fault area, similarity of hypocentre locations has been used as a classification criterion (e.g. Vidale et al., 1994; Nadeau and Johnson, 1998; Yu, 2013; Yao et al., 2017; Bohnhoff et al., 2017). However, the confidence of the result is strongly based on how precise the relative locations of the earthquakes are. Accordingly, hypocentral distance has not been used as the only criteria with which to identify repeating earthquakes.

Some studies have also used the time between events (e.g. a minimum of 30 days) to distinguish repeating earthquakes from swarms (Gibowicz and Lasocki, 2007; Bohnhoff et al., 2017; Schaff and Richards, 2011). Others use more than one criterion to select repeating earthquakes. In addition to waveform similarity and hypocentral distance, the similarity of differential S-P times can be used as a threshold (Chen et al., 2008; Rau et al., 2007; Uchida et al., 2006). Hughes et al. (2021) and Uchida (2019) have both provided detailed and complete information of the criteria that have been used to identify repeating earthquakes.

2.3 Mechanics of repeating earthquakes

2.3.1 Earthquake cycle models

Reid (1910) proposed the theory of elastic rebound, which is one of the central ideas regarding earthquake mechanics. The elastic rebound theory states that crustal stresses resulting from distant large-scale shearing motions (e.g. tectonic plate movements) cause elastic strain to accumulate near faults. Abrupt frictional sliding occurs (i.e. an earthquake) when the resulting shear stress on the fault reaches a threshold determined by the rock properties of the fault. During the earthquake, some of the elastic strain is released and after the earthquake elastic strain begins to accumulate again. This behaviour is called stick-slip, as earthquakes are “slip” episodes followed by periods of no slip (“stick”).

Figure 2.3: Recurrence model of an idealised earthquake cycle in which all earthquakes have the same stress drop and coseismic slip. Red vertical lines represent earthquakes and green lines represent the inter-seismic phase with a constant loading rate. Figure modified from Shimazaki and Nakata (1980).

Variants of the elastic rebound theory that also assume a constant loading rate have been proposed. Figure 2.3 is the characteristic earthquake model (Schwartz and Coppersmith, 1984) in which shear stress accumulates to the level of fault strength τ_1 . When the shear stress reaches τ_1 , the fault ruptures and the shear stress then decreases to the same level as the friction on the fault τ_2 . In this model, the shear stress released during each earthquake is the same, as is the slip on the fault and the recur-

rence interval between earthquakes. This model leads to perfect periodicity. Figure 2.4 shows the (a) time-predictable and (b) slip-predictable models (Shimazaki and Nakata, 1980). In the time-predictable model, fault strength is constant, the fault will always rupture when reaching τ_1 but slip on the fault can be different every rupture. In the slip-predictable model, every rupture reduces the shear stress to τ_2 and the shear stress is proportional to the potential fault slip.

Figure 2.4: Recurrence models of (a) Time-Predictable and (b) Slip-predictable models. Red vertical lines represent earthquakes and green lines represent the inter-seismic phase with a constant loading rate. For both models, the time between earthquakes is not necessarily the same. In the time-predictable model, the fault will always rupture when reaching τ_1 . In the slip-predictable model, every rupture reduces the shear stress to τ_2 .

If we define repeating earthquakes as earthquakes of similar magnitude rupturing the same fault patch at different times (e.g. Nadeau et al., 1994; Vidale et al., 1994; Ellsworth, 1995), in theory, they could be explained by the characteristic earthquake model. High waveform similarity supports the fact that they are located close to each other, in the same fault area. The similarity in magnitudes indicates that the shear stress released in every rupture is constant and if the recurrence times between events is also constant, there must be a constant loading rate.

The reality is that not all repeating earthquakes can be explained by the characteristic model, or even the time- and slip-predictable models, and there is evidence that the assumptions made (e.g. long constant loading rate, constant τ_1 , all the slip in the region being liberated by the rupture) do not hold true in field and laboratory observations (Rubinstein et al., 2012, and references therein).

The most questionable assumption from the previous mentioned models is the constant loading rate. Previous models assume that the cause of accumulating shearing stress are large-scale shearing motions, but the observed values of crustal stress accumulation rates (1–50 kPa/year) are small in relation to typical earthquake stress drops (1–10 MPa) (Uchida and Bürgmann, 2019). With these observed values, it should take at

Figure 2.5: Asperity model of repeating earthquakes. Repeating earthquakes rupture locked asperities in an aseismically sliding area (creeping). Figure from Uchida et al. (2003).

least a thousand years for the shear stress to accumulate and reach τ_1 . Therefore, repeating earthquakes with recurrence intervals of a few years or less must be driven by a faster loading.

The most used model of repeating earthquakes occurrence portrays them as the re-rupturing of locked asperities within an aseismically sliding (creeping) area (Figure 2.5) (Ellsworth, 1995; Nadeau and Johnson, 1998; Igarashi et al., 2003; Uchida et al., 2003; Johnson, 2010). The term asperity refers to locked patches of the fault plane that are resistive to breaking (Aki, 1984). The area of the asperity is presumed to be proportional to the size of the earthquake when it ruptures, meaning that large asperities generate large earthquakes. As illustrated in Figure 2.5, large locked asperities that previously rupture are surrounded by a creeping area with small locked asperities. According to this interpretation, repeating earthquakes are the ruptures of these small asperities (Nadeau et al., 1994; Beeler et al., 2001; Ellsworth, 1995; Igarashi et al., 2003; Uchida et al., 2003).

Some studies have proposed different rupture processes for burst- and continual-type repeating earthquakes. Continual-type repeating earthquakes are thought to rupture small asperities quasi-periodically (Ellsworth, 1995; Nadeau and Johnson, 1998; Igarashi et al., 2003; Beeler et al., 2001). When an asperity is locked, it accumulates shear stress as the surrounding area is aseismically slipping (creeping) until a critical

threshold is reached and rupture occurs. That being so, the cumulative slip of repeating earthquakes of an asperity reflects the cumulative slip occurring in the area surrounding that asperity (Nadeau and Johnson, 1998). If the asperity always slips seismically, the amount of aseismic slip in the area surrounding the asperity can be estimated from continual-type repeating earthquakes. (Nadeau and Johnson, 1998; Nadeau and McEvilly, 1999; Igarashi et al., 2003; Uchida et al., 2003).

Burst-type repeating earthquakes can not be explained by the constant loading-rate models due to this short family durations. This may explain why burst-type repeating earthquakes have been associated with swarm activities, aftershocks of large earthquakes (e.g. Igarashi et al., 2003; Kimura et al., 2006; Templeton et al., 2008; Yao et al., 2017) and fluid intrusion (e.g. Mesimeri and Karakostas, 2018; Shaddox and Schwartz, 2019). As an example of this distinction, Mesimeri and Karakostas (2018) found streaks of continuous-type repeating earthquakes at 10 km depth near two mayor faults, indicating creeping along the brittle-ductile transition zone. In contrast, burst-type repeating earthquakes were associated with small fault segments dipping at high angles and triggering due to fluid intrusion.

2.3.2 Recurrence intervals of repeating earthquakes

On the basis that earthquake renewal processes are cyclic, the source characteristics of repeating earthquakes have long been studied in relation to their recurrence intervals times. Nadeau and Johnson (1998) developed a scaling relationship between seismic moment (M_0) and recurrence interval (T_r):

$$T_r \propto M_0^{\frac{1}{6}}. \quad (2.5)$$

According to Equation 2.5, events with larger magnitudes are expected to have longer recurrence intervals. This relationship was developed using repeating earthquakes from, California but appears to hold in other regions including Japan (Chen et al., 2007), Taiwan (Chen et al., 2007), the Tonga-Kermadec-Vanuatu subduction zones (Yu, 2013) and the Guerrero subduction zone in Mexico (Dominguez et al., 2016). The application of this relationship to the repeating earthquakes obtained in this thesis is considered in Chapter 5.

In some cases, the $M_0 - T_r$ relationship does not follow Equation 2.5. Chen et al. (2010) observed that after an M6 in Parkfield, California, some families increased in seismic moment while their recurrence intervals decreased. This was interpreted as

faster loading rates due to the nearby earthquake. This behaviour was also observed in repeating earthquake families after the M9.0 Tohoku-Oki earthquake in Japan (Uchida, 2015). Peng et al. (2005) also observed a negative dependency between M_0 and T_r for deep (5.5 km) repeating earthquakes in the southern San Francisco Bay and attributed this to changes in the medium due to the postseismic period of the M6.2 Morgan Hill earthquake nearby.

The triggering of repeating earthquakes has also been observed and studied. In addition to the studies mentioned in the paragraph above, Hatakeyama et al. (2017) observed new families of repeating earthquakes located where the M9.0 Tohoku-Oki earthquake ruptured. The magnitudes of events within these families decreased over time and some families disappeared within a year. The triggering and interaction of repeating earthquakes was studied by Chen et al. (2013) using data from repeating earthquakes in Parkfield, California. Chen et al. (2013) outlined three mechanisms that could produce short apparent or actual recurrence intervals: (1) cross-triggering of closely spaced repeating asperities (e.g. Nadeau et al., 1995; Nadeau and McEvitty, 2004; Lengliné and Marsan, 2009); (2) high aseismic loading rates due to nearby large events (e.g. Vidale et al., 1994; Schaff et al., 1998); and (3) burst-type behaviour due to large nearby earthquakes (e.g. Kimura et al., 2006; Templeton et al., 2008). The recurrence intervals of repeating earthquake families have been shown to be strongly related to the regional loading rate and distribution of creep on a fault in different repeating earthquake studies around the world (Chen et al., 2013).

2.4 Repeating earthquakes in different settings

The faults in California have been the most studied of any in the world using repeating earthquakes. The availability of long-duration high-quality data has made it possible to record and analyse repeating earthquakes (Ellsworth, 1995). Nadeau et al. (1994, 1995); Rubin et al. (1999); Lengliné and Marsan (2009) among others, detected repeating earthquakes using correlation and coherence in the San Andreas Fault near the creeping segment at Parkfield, California. These catalogues of repeating earthquakes have since been analysed extensively: Nadeau and Johnson (1998) found that moment is being released as a function of time in a regular manner and established a relationship between recurrence interval and seismic moment. To explain this relationship, Beeler et al. (2001) proposed a stick-slip and creep-slip model which includes aseismic slip. Rubinstein and Ellsworth (2010) precisely estimated the relative size of repeating earthquakes using singular value decomposition and Abercrombie (2014)

obtained stress drops of the repeating earthquakes. Chen et al. (2013) studied the triggering effects on repeating earthquakes from nearby seismicity and Zechar and Nadeau (2012) investigated the temporal predictability of the sequences using retrospective forecast experiments. Nomura et al. (2014) proposed a stochastic model to estimate spatiotemoral changes. Several others have calculated slow slip at depth on the San Andreas fault by using repeating earthquakes (e.g. Nadeau et al., 1994, 1995; Turner et al., 2013) and illuminated fault structure (e.g. Waldhauser et al., 2004; Rubin et al., 1999)

At the Calaveras fault, also in California, Vidale et al. (1994); Poupinet et al. (1984); Ellsworth (1995); Templeton et al. (2008) have identified sequences of repeating earthquakes. Although studies of the Calaveras fault have not yet been as extensive as those of the San Andreas fault, repeating earthquakes have been used to measure velocity changes after $>M6$ earthquakes (Poupinet et al., 1984; Schaff and Beroza, 2004), study variations of stress drops (Marone et al., 1995; Vidale et al., 1994), calculate fault creep (Templeton et al., 2008), to illuminate fault structure (Schaff et al., 2002) and even to correct earthquake catalogues (Rubin, 2002).

On the other side of the Pacific Ocean, repeating earthquakes have been studied in the subduction zones surrounding Japan. Repeating earthquakes have been detected, from the Ryukyu trench at the south of the Kuril trench at the north (Igarashi, 2010) using mainly waveform similarity (e.g. Uchida et al., 2003, 2009; Igarashi et al., 2003; Kimura et al., 2006; Yamashita et al., 2012). In Japan, the application of repeating earthquakes is extensive: to study interplate coupling (e.g. Igarashi et al., 2003; Igarashi, 2010; Uchida et al., 2009; Uchida and Matsuzawa, 2011; Nomura et al., 2017), to analyse variation in creep/quasi-static slip (Uchida and Matsuzawa, 2011; Uchida, 2015, 2016) and recurrence intervals (Mavrommatis et al., 2015) before and after big earthquakes such as the Mw 9.0 Tohoku-Oki, and to estimate source parameters (Uchida et al., 2007, 2012; Uchida, 2016).

Repeating earthquakes have also been identified and studied in Taiwan (Chen et al., 2008; Chen and Lapusta, 2009), China (Schaff and Beroza, 2004; Schaff and Richards, 2011; Li et al., 2007), Turkey (Peng et al., 2005; Uchida et al., 2019), Greece (Mesimeri and Karakostas, 2018), Tonga (Yu and Wen, 2012; Yu, 2013), Mexico (Dominguez et al., 2016) and Canada (Hayward and Bostock, 2017). Repeating earthquakes have also been found in nontectonic settings: in active volcanoes, landslides, mining induced micro-earthquakes, glaciers and even on the moon (Uchida and Bürgmann, 2019).

Studies in New Zealand

In New Zealand, few studies have yet been undertaken in relation to tectonic repeating earthquakes (Figure 2.6). Offshore the Raukumara Peninsula, Shaddox and Schwartz (2019) identified burst-type repeating earthquakes using data from ocean-bottom seismometers. These repeating earthquakes were interpreted as have been triggered by fluid migration during a plate-interface slow slip event due to being co-located within a subducted seamount fracture network. On account of this, repeating earthquakes were postulated to be seismic indicators of shallow slow slip and thus useful for identifying fracture networks. Wallace et al. (2017) identified repeating earthquakes between 2010 and 2017 near Porangahau with a possible association to nearby shallow slow slip events.

Hughes et al. (2021) is the most extensive study to date of repeating earthquakes in New Zealand, and developed a composite criterion for their detection. This criterion was used to identify repeating earthquakes in the Raukumara Peninsula. Hughes et al. (2021) then estimated some of the source properties of the identified repeating earthquakes and attempted to calculate plate-interface slip-rates. The composite criterion that Hughes et al. (2021) developed was also tested in Marlborough and Fiordland to ensure it could be applied successfully in other regions of New Zealand.

We have used the criterion developed by Hughes et al. (2021) to identify repeating earthquakes near Porangahau, Hikurangi Margin to further extend the studies of repeating earthquakes in New Zealand.

Figure 2.6: Locations of repeating earthquake studies in New Zealand. The study of Wallace et al. (2017) overlaps with the study area of this thesis.

Chapter 3

Data and Methodology

This chapter presents the data and processes employed to construct the final catalogue of repeating earthquakes near Porangahau. In Section 3.1 we describe the data used in this study. The methodology used to identify repeating earthquakes is described in detail in Section 3.2. The clustering of events to form repeating earthquake families is outlined in Section 3.3. Section 3.4 describes the process of manual picking and polarity identification. The methodologies we used to obtain locations for all the events are described in Section 3.5, followed by the details of the magnitude calculations in Section 3.6. This chapter ends with a description of the methodology used to obtain focal mechanisms, in Section 3.7. Figure 3.1 provides a general overview of the steps undertaken:

Figure 3.1: Flow diagram outlining the steps undertaken in this thesis.

3.1 Data

The starting point for our analysis is a catalogue of potential repeating families for all New Zealand obtained by Chamberlain et al. (2019). This initial catalogue contains clustered events from GeoNet’s catalogue (Petersen et al., 2011) identified using a composite criterion of a minimum normalized cross-correlation threshold of 0.7 and hypocentral separation of <30 km. This clustering was the starting point for finding repeating earthquakes in the area surrounding Porangahau.

To aid comparison between our final catalogue and the original GeoNet catalogue we maintained the GeoNet event IDs throughout our analysis. For our area of study, we selected earthquakes from the initial catalogue located within 80 km of the township of Porangahau (Figure 3.2) and occurring between 2004 and 2018. After selecting the events, we downloaded the waveforms that GeoNet recorded for these earthquakes using the International Federation of Digital Seismograph Networks (FDSN) service via ObsPy (Beyreuther et al., 2010). Waveforms from broadband and short-period stations located within an epicentral distance of 150 km were downloaded regardless of whether they had been used previously by GeoNet to obtain locations. A total of 73 stations were used (Figure 3.2), available at different periods of time and with different instrumental characteristics.

The initial catalogue details and the waveforms used to identify repeating earthquakes both come from GeoNet. As a result, the completeness of our final repeating earthquake catalogue depends directly on the completeness of GeoNet’s earthquake catalogue. There have been several changes in the GeoNet seismic network within the time span of this study such as the number of stations, the type of recording instruments, the methodologies used to locate events and calculate magnitudes, velocity models, and detection methods. We are aware of these variants in GeoNet’s catalogue and have taken measures to diminish their effects in the identification of repeating earthquakes. The measures taken are briefly mentioned in this chapter and later discussed in Chapter 5.

Figure 3.2: Map of the study area showing the stations in this study. Stations are coloured according to the year of installation. The circle shows the 80 km radius area of interest.

3.2 Repeating earthquake identification

In order to identify repeating earthquakes, we have employed a cross-correlation identification method based on the work of Hughes et al. (2021) and informed by the review of previous repeating earthquake studies outlined in Section 2.2. Here, we first describe the waveform reprocessing undertaken, which includes instrument response removal, filtering, and defining the end of the signal and window length. We then detail the cross-correlation method used to identify repeating earthquakes.

As noted in the previous section, the instruments used to record earthquakes of interest near Porangahau have changed through time. As indicated by Equation 2.1, the instrument response (i.e. $i(t)$) is convolved to form the seismogram, which means that the measured waveform similarity for two truly repeating earthquakes can be affected by a change of instruments. To account for this, we removed the instrument response from the waveforms prior to correlation.

Filtering is an important operational parameter for repeating earthquake detection (Section 2.2.1). We applied a fourth-order bandpass filter between 2–20 Hz to remove the low-frequency noise from the signals while retaining the dominant seismic energy. After applying this filter, the end of the signal was determined using a signal-to-noise (SNR) ratio. We computed the SNR every sample using the root mean square (rms) amplitude of the 15 s signal window divided by the rms amplitude of 15 s of data before the P-pick (noise). The end of the signal was determined when the SNR first fell below 1.1 and remained below this for 0.5 seconds. Once we calculated the end of the signal, we identified the window length to be 75% of the signal and 1 s before the P-pick based on the work from Hughes et al. (2021). We established a window length of minimum 15 s and maximum 100 s for cases where the end of the signal was not found using SNR or for when the end of the signal was >100 s. We refer to this signal window as a *template*.

Prior to computing multisegment normalized cross-correlations, we aligned the templates using pick times and their maximum correlation allowing up to 0.9 s shift to account for uncertainty in pick times. We used the pick times reported by GeoNet to align the templates. For the templates that had no pick time from GeoNet, we used calculated expected pick times based on picks from GeoNet from the same family. We performed a multisegment normalized cross-correlation following the methodology proposed by Gao and Kao (2020). The template was divided into four equal segments and each segment was cross-correlated with the corresponding segment from the other waveform using a module in EQcorrscan (Chamberlain et al., 2018). The final cross-

correlation coefficient was the average of the cross-correlation values obtained for the four successive segments. This procedure is illustrated in Figure 3.3. We used this approach in order to minimise bias arising due to large-amplitude phases when computing the cross-correlation (Gao and Kao, 2020). Following Hughes et al. (2021), we decided to employ a minimum cross-correlation value of 0.95 at three or more stations for earthquakes to be considered repeating earthquakes. Figure 3.4 shows the waveforms of an identified repeating earthquake family in this study.

Figure 3.3: Figure illustrating waveform pre-processing (a) and multisegment normalized cross-correlation (b). a) Top: Raw waveform of event 2018p364597; Middle: Signal filtered with a 2-20 Hz bandpass filter and instrument response removed with a pink background indicating the trimmed waveform; Bottom: Zoom of the filtered and trimmed waveform. b) Filtered, trimmed, normalized and aligned waveforms (templates) of the same event as a) (Top) and event 2763013 (Bottom). Shaded areas indicate the template windows used in the multisegment cross-correlation, each segment with a different colour. All traces shown in this figure are vertical components recorded at station PXZ.

Figure 3.4: Correlated waveforms of the same family on two different stations: a) PRHZ East component and b) DVHZ East component. The red trace indicates the most recent (youngest) event of the family. Events are ordered by time with the correlation coefficients (cc) of each trace with the most recent event (red waveform). Gray backgrounds indicate that no cross-correlation was performed for that event at that station due to lack of data. This is the biggest family obtained, which consists of nine events in the lapse of 11 years (2007–2018).

3.3 Clustering into repeating earthquake families

To cluster events into discrete families of repeating earthquakes, we implemented a hierarchical clustering approach using an UPGMA (Unweighted Pair Group Method with Arithmetic Mean) algorithm by (Müllner, 2011). We implemented this clustering as an extension of the methodology described by Hughes et al. (2021) because of the large number of high correlations from the initial clusters. The hierarchical clustering was implemented using the SciPy clustering package (Virtanen et al., 2020), in which a rooted tree is constructed using a distance matrix. The average correlation values across all stations were converted into a distance matrix, where high correlation values correspond to short distances. The distance matrix was used as the input data for the hierarchical clustering method. In the UPGMA algorithm, each element of the distance matrix starts as being a cluster on its own. Next, the closest two elements are joined together and the distances to the rest of the elements or clusters are recalculated by using an average of the element’s distances inside the cluster. The closest clusters are joined again and the process is repeated until all elements are connected in a single cluster.

Figure 3.5: Dendrogram of an initial family showing the final clusters obtained. The two events on the left with blue branches did not pass the minimum cross-correlation threshold. The four events with green branches on the right side passed the minimum threshold and formed as a family of repeating earthquakes.

Hierarchical clustering does not determine unique clusters, as illustrated by Figure 3.5, in which the branch colours help distinguish the subtrees within a bigger cluster. SciPy includes a leaf-ordering algorithm (Bar-Joseph et al., 2001) to cluster events obtained

from the UPGMA algorithm. These new groups or clusters constitute the repeating earthquake families, which only contain events with a minimum cross-correlation value of 0.95 at three or more stations.

Using this second leaf-ordering clustering, a total of 225 families from 2,170 initial clusters were identified. The families were grouped into geographical groups and named based on this grouping, as discussed in detail in Section 4.1.

3.4 First arrivals and polarity identification

To this point in the analysis, all the information about each earthquake has been provided by GeoNet. Earthquake locations calculated by GeoNet, in particular, have not been consistent though the years as location methods, velocity models and analysis methods have changed. The number of stations in the GeoNet network has been increasing though the years, which means that the most recent events are recorded on more stations and in general have more arrivals and polarities. For our catalogue to be internally consistent, we manually picked the P and S phase arrivals of the most recent event in each family. Alongside the improved consistency in phase-arrival picking, our manual phase analysis also allowed us to pick additional arrivals at stations not picked by GeoNet and to add S-arrivals, which have not been routinely picked by GeoNet in the past.

We picked a total of 4,699 P-phases and 4,680 S-phases at 73 stations using SeisAn (Havskov and Ottemöller, 1999). We also recorded confidence weights for each pick, with weights of 0 marking high-confidence picks, and weights of 4 being zero-confidence. We also manually picked P-wave first-arrival polarities for later use in determining focal mechanisms. Figure 3.6 shows an example of high-confidence picks (weights of 0) on station PRHZ and lower-confidence picks (weights of 1 and 2) at TIWZ due to differences in noise. Polarities were not added to all P arrivals, due to the lack of clear identification of the polarity against the noise in some cases. Filters were used to aid in the identification of arrivals, but all polarities and arrivals were picked on unfiltered traces to avoid possible filter-induced phase shifts.

Figure 3.6: P and S phase identification with the program SeisAn, where the use of weights is illustrated at two different stations. Waveforms shown are unfiltered. These waveforms correspond to earthquake 2018p364597, which is the most recent earthquake of the biggest family. The waveforms have no filter or any other processing in this figure, but the filters already implemented in SeisAn were used to aid in the phase identification.

3.4.1 Pick corrections

Only the most recent event of each family was picked manually, leaving the remaining events in the family with the phase arrivals from GeoNet. In order to have homogeneous picks without picking manually all the events, we corrected the remaining phase arrivals from GeoNet using cross-correlation. For the picks to be corrected using cross-correlation, they have to be from the same family, station and channel as the ones from the most recent event. For traces without GeoNet picks, temporary picks were added using the move-out from the manual pick and the origin time from GeoNet. Move-outs are the time difference between the earthquake origin time and the P-phase arrival at a station.

Using the corresponding temporary and GeoNet picks, we used a function from EQ-corrscan to obtain phase arrivals relative to the most recent event picks. This function computed cross-correlation derived pick corrections following the method of Deichmann and Garcia-Fernandez (1992) to obtain sub-sample precision pick accuracy. We only retained phase arrivals when the cross-correlation value exceeded 0.4. We chose 0.4 as a minimum value after trying different thresholds and plotting the corrected picks. We did not adopt a higher threshold because of noise in the waveforms. Figure 3.7 shows some of the final relative phase arrivals of one event.

Figure 3.7: Correction of the P and S phase arrivals of event 2763013. The manual picks of event 2018p364597 were used as templates (red waveforms). P-pick correction was performed on the vertical component and S-pick correction on the horizontal components. This figure shows 8 of the 22 corrected traces for illustrative purposes. Waveforms are shifted and aligned on pick-time

3.5 Location of repeating earthquakes

Each earthquake is parameterized by its hypocentre and origin time (T). The hypocentre is the physical location: distance east (X), distance north (Y) and depth (Z). According to Lomax et al. (2000), a location is said to be *absolute* when it is specified within a fixed, geographic coordinate system and a fixed time base. When the location is determined with respect to another object (e.g. another earthquake), it is said to be a *relative* location. Locating an earthquake and finding the origin time can be stated as an inverse problem using the first arrival times of the seismic phases recorded at several stations. The observed arrival time (t) at station i can be written as:

$$t_i = t_{calc}(X, Y, Z, x_i, y_i, z_i) + T \quad (3.1)$$

where t_{calc} is the calculated travel time as a function of the station location (x_i, y_i, z_i) and hypocentre location (X, Y, Z) (Havskov and Ottemoller, 2010).

From Equation 3.1, we know each station location and can assume a velocity model, which leaves four unknown variables: X, Y, Z and T . The hypocentre and origin time could be determined with only four arrival time observations, but in reality there are usually more than four stations providing n equations from n observations. When the number of equations surpasses the number of unknown variables, the system is over-determined and other approach is needed to calculate an hypocentre and origin time. Although we can't obtain an exact solution, a location and time that "best" solves the over-determined, inconsistent equation can be calculated (Stein and Wysession, 2003). The "best" solution is said to be found when the sum of the square residuals (r) between the observed (t) and calculated (t^{calc}) arrival times at each of the stations i is minimized (Havskov and Ottemoller, 2010):

$$RSS = \sum_i r_i^2 = \sum_i (t_i - t_i^{calc})^2 \quad (3.2)$$

Although minimizing the sum of squared residuals seems straightforward, the travel time function is a nonlinear function, which cannot, in general, be solved analytically (Havskov and Ottemoller, 2010). ΔX , ΔY , ΔZ , and ΔT are the adjustments to the origin time and location for the residuals to be minimized. When the adjustments are relatively small, the residual can be calculated by approximating the travel time function with the expansion of the first-order Taylor series (Equation 3.3).

$$r_i = \frac{\partial t_i}{\partial X} \Delta X + \frac{\partial t_i}{\partial Y} \Delta Y + \frac{\partial t_i}{\partial Z} \Delta Z + \Delta T \quad (3.3)$$

There are several methods to solve the inverse problem of locating an earthquake, including linearizing the equations and non-linear methods such as grid search or iterative methods (Havskov and Ottemoller, 2010). To obtain the absolute locations we used a probabilistic non-linear method.

3.5.1 Absolute locations of repeating earthquakes

For this thesis we used NonLinLoc (Lomax et al., 2000) to obtain absolute locations of the most recent earthquake of each family. NonLinLoc uses a systematic grid search algorithm to obtain a misfit function, “optimal” hypocentres, and an estimate of the posterior probability density function (PDF) for the spatial hypocentre location (Lomax et al., 2000). NonLinLoc allows the use of 3-D velocity models, which made it possible to use the NZ3D2.1 velocity model created by Eberhart-Phillips et al. (2017) without any modifications. In addition, NonLinLoc has the option of transforming the location results into geographical co-ordinates. When using a 3-D velocity model in NonLinLoc, the travel times between each station and all the nodes of the velocity grid are calculated and stored. The gradients of travel times are then used to estimate the ray take-off angles at each node. With NonLinLoc, the PDF or misfit value is determined using a grid-search algorithm, providing a non-linear spatial solution to the earthquake location problem (Lomax et al., 2000).

To efficiently search the model-space we used the NonLinLoc Oct-Tree search algorithm which recursively subdivides and samples the grid in cells in three dimensions, computing the PDF within each cell. The cell with the largest probability is subdivided into a number of smaller cells and for each cell a PDF is computed. This process is repeated until a minimum cell size or a maximum iteration number is reached. As location method we selected an equal differential time likelihood function with weights, which reduces the PDF values at points with inconsistent Oct-Tree estimates and leads to more compact PDF locations (Lomax et al., 2000). With these parameters in NonLinLoc, we obtained absolute locations for all of our 225 most recent events. The input file containing the parameters used to compute the absolute locations is in Appendix B.1.

3.5.2 Relative locations of repeating earthquakes

We obtained relative locations for two main purposes: 1) to obtain more accurate relative locations for closely spaced repeating earthquake families; and 2) to verify that events within the same family actually had overlapping hypocentres. Relative

locations were computed using GrowClust (Trugman and Shearer, 2017), which is a double-difference algorithm that uses differential travel times and cross-correlation in a hybrid, hierarchical clustering algorithm.

With the double difference algorithm, the relative location of an earthquake is found by minimising the double difference residual (dr_k^{ij}) between earthquakes i and j for each observation at station k (Equation 3.4; Waldhauser and Ellsworth (2000)):

$$dr_k^{ij} = (t_k^i - t_k^j)^{calc} - (t_k^i - t_k^j)^{obs} \quad (3.4)$$

Equation 3.4 strongly depends on the assumption that the travel paths for both events are similar and does not work when events are not close enough. The change in hypocentral distance between events i and j can be obtained with the following equation:

$$dr_k^{ij} = \frac{\delta t_k^i}{\delta x} \Delta x^i + \frac{\delta t_k^i}{\delta y} \Delta y^i + \frac{\delta t_k^i}{\delta z} \Delta z^i + \delta \tau^i - \frac{\delta t_k^j}{\delta x} \Delta x^j - \frac{\delta t_k^j}{\delta y} \Delta y^j - \frac{\delta t_k^j}{\delta z} \Delta z^j - \delta \tau^j \quad (3.5)$$

where t is the travel times vector of events i and j with locations x , y and z and origin times τ (Waldhauser and Ellsworth, 2000). δx , δy , δz and $\delta \tau$ are the changes needed in the hypocentral parameters for the model to better fit the data. These changes are calculated using linear techniques for all stations and hypocentral pairs (Waldhauser and Ellsworth, 2000).

GrowClust uses the L1 norm for residual minimisation which is less sensitive to outliers in comparison to the L2 norms used in other algorithms. However, one disadvantage we encountered is that GrowClust is currently only able to use 1-D velocity models. The differential travel times are obtained from the cross-correlation of pairs of events registered at the same stations. The calculations to obtain the differential travel times from cross-correlation, were obtained using EQcorrscan functions from the package *catalog_to_dd*. These functions allowed us to compute cross-correlations between events and to obtain their respective differential-times required as input for GrowClust. For GrowClust we set a minimum cross-correlation coefficient of 0.5, a maximum station distance of 100 km and a maximum root-mean-square error of 0.5 s. The arrival times are weighted by the correlation values. The input file containing the parameters used to compute the relative locations is in Appendix B.2

The input to GrowClust was a catalogue of all the repeating events, including the most recent ones which had absolute locations. For the events without absolute location, we set their initial location to be the absolute location of the most recent event in the

same family. GrowClust relocated the initial hypocentres using differential times from waveform cross-correlation, providing relative locations for 304 out of the 307 events that did not have an absolute location.

3.6 Local magnitude calculation

The magnitude of an earthquake is used to quantify its size (Stein and Wysession, 2003; Lay and Wallace, 1995). Amplitudes of seismic phases have been used to obtain magnitudes (e.g. Richter, 1935): when using phase amplitudes, the bigger the amplitude, the bigger the size of the earthquake. However, this is only true when the waveforms have been corrected for attenuation and geometric spreading (Stein and Wysession, 2003) and until the values reach a saturation magnitude. One way to construct a magnitude scale is the following (Stein and Wysession, 2003):

$$M = \log(A/T)_{max} + F(h, \Delta) + C \quad (3.6)$$

where A is the amplitude of the signal, T is the period of the signal, F is a correction for earthquake depth (h) and epicentral distance (Δ) and C is a regional scale factor (Stein and Wysession, 2003; Lay and Wallace, 1995).

Richter (1935) was the first to introduce the concept of magnitude scale with the local magnitude M_L or the “Richter scale”. This magnitude scale was designed for earthquakes with hypocentres in southern California that were recorded on a Wood-Anderson seismograph. By using equation 3.7, the relative size of an event is calculated by comparison to a reference event:

$$M_L = \log A(\Delta) - \log A_0(\Delta) + S \quad (3.7)$$

where A is the maximum amplitude recorded at a Wood-Anderson seismometer, A_0 is the maximum amplitude of the reference event, Δ is the epicentral distance and S is a correction factor for the station. The reference event has $M_L = 0$ and was chosen by Richter so that A_0 (maximum amplitude) was 1×10^{-3} m at an epicentral distance of 100 km (Richter, 1935; Gutenberg and Richter, 1942; Lay and Wallace, 1995).

Although we are not working with earthquakes from southern California recorded on Wood-Anderson seismometers, we can obtain a local scaled magnitude (Gutenberg and

Richter, 1942) for our earthquakes using equation 3.8 (Havskov and Ottemoller, 2010):

$$M_L = \log A(\Delta) + a \log(\Delta) + b\Delta + S \quad (3.8)$$

where A is the observed amplitude and a , b , and S are constants representing geometric dispersion (here assumed to be 1), attenuation and station correction term, respectively.

We followed a procedure similar to that described by Boese et al. (2012) to obtain local magnitudes calibrated with moment magnitude M_w . To find appropriate values for a, b , and S for the Porangahau study region, we compared the amplitudes of our most recent events against moment magnitudes obtained using moment tensor inversion from GeoNet (Ristau, 2013). As none of the repeating earthquakes in our catalogue had moment tensor results, we selected 41 earthquakes located in our study area (Figure 3.8) with magnitudes similar to those in our our catalogue. The minimum magnitude from our catalogue was M_L 1.2 and the maximum M_L 3.8 using GeoNet’s reported local magnitudes. The range we selected for magnitudes with moment tensors was M_L 3.4 to M_L 5.2. As moment tensors are not calculated for all the earthquakes GeoNet reports, we were not able to find earthquakes with lower magnitudes than 3.4 in our area of study. We manually picked the earthquakes with moment tensors and obtained amplitudes for all events. The amplitudes were picked automatically on the horizontal component using the manual S phase pick as a reference.

After obtaining the amplitudes, we inverted Equation 3.8 by solving the general inverse problem using LU-matrix factorization:

$$Y = Xm + e \quad (3.9)$$

Y includes the amplitudes of all events and the 41 GeoNet moment magnitudes (M_w). X is a matrix representing which component of each station has an amplitude reading and which events have moment magnitudes. m are the parameters of interest: M_L , attenuation term b and the station corrections terms S . The last term e is the vector of magnitude residuals (Boese et al., 2012).

Appendix A.2 contains a table with the station correction terms from the inversion. Next, we obtained relative magnitudes for the rest of the events within a family using the calibrated local magnitudes. This process was carried out using EQcorrscan function *relative_magnitude* with a minimum cross-correlation threshold of 0.5 following the method of Schaff and Richards (2014).

Since magnitudes are empirical estimates of earthquake size and they do not have a

Figure 3.8: Map with the locations of the 41 earthquakes used to calibrate the local magnitudes as red diamonds. Absolute locations of the most recent events are shown as orange circles.

direct connection to the physics of an earthquake, the size and energy release of an earthquake is best represented by the scalar seismic moment (M_0) (Stein and Wyssession, 2003; Lay and Wallace, 1995). Seismic moment is directly related to the size of the earthquake and can be expressed in terms of the rigidity (μ), average slip ($\bar{D}$) and fault area (S):

$$M_0 = \mu \bar{D} S \quad (3.10)$$

M_0 has been used previously to obtain seismic moment-recurrence intervals for repeating earthquakes as seen in Nadeau and Johnson (1998). In order to use this relationship to analyse our repeating earthquakes, we estimated seismic moments using equation 3.11, after Hanks and Kanamori (1979), which indicates the relationship between seismic moment (M_0) in dyne-cm and moment magnitude (M_w):

$$\log(M_0) = 16.1 + 1.5M_w \quad (3.11)$$

Since we calibrated our local magnitudes with moment magnitude, using Equation 3.11 to obtain seismic moment is a valid approach. We discuss the seismic moment-recurrence intervals of our repeating earthquakes in Chapter 5.

3.7 Focal mechanism calculation

A focal mechanism describes the geometry of the fault based on the radiation pattern of the earthquake that it produced (Havskov and Ottemoller, 2010; Stein and Wyssession, 2003). For simplicity, earthquakes are assumed to occur on a planar surface whose geometry can be described with three angular parameters: strike (ϕ), dip (δ) and rake (λ). When recording an earthquake on a network of seismometers, the polarities of the P-wave first arrivals contain information on the geometry of faulting and can be used to determine the focal mechanism (Stein and Wyssession, 2003). If the side of the fault moves towards a particular station, the resulting P-wave first motion will be compressional (C) and if it moves away from the station, it will be dilatational (D).

For this study, we used the fault-motion polarities obtained in Section 3.4 to calculate focal mechanisms for the most recent events using the probabilistic Bayesian method of Walsh et al. (2009). This method accommodates uncertainties arising from errors in location and take-off angles, as illustrated in Figure 3.9. We use the absolute locations and take-off angles calculated in Section 4.2.1 to account for those errors. With this methodology, fault plane solutions were calculated for all the most recent events. In

Section 4.4 we elaborate on our criteria for selecting the well-constrained focal mechanisms to constitute our final catalogue.

Figure 3.9: Diagram illustrating different take-off angles due to the uncertainties in P-wave arrival times. Different arrival times lead to different possible earthquake locations, each with its own take-off angle. Figure from Walsh et al. (2009).

Chapter 4

Results

This chapter presents the main results of this research. Section 4.1 gives a brief summary of the repeating earthquakes detected and the geographic groupings to which they were assigned. Next, Section 4.2 documents the absolute and relative locations obtained. Section 4.3 addresses the calculated magnitudes, and Section 4.4 outlines the repeating earthquakes' focal mechanisms. The chapter ends with a summary of the results, in Section 4.5.

4.1 Nomenclature of repeating earthquake families

We identified 225 repeating earthquakes families from 2,170 initial clusters occurring between 2004 and 2018, using the methodologies described in Sections 3.2 and 3.3. The total number of events contained in the families is 532, with most families containing only two events. Families were divided into 16 geographical groups based on the most recent event's location (latitude, longitude and depth) as shown in Figure 4.1. The geographical groups were created primarily to easily refer to each family. The grouping of families was subjective, according to how close families were to each other and their depth. The names of each group correspond to nearby places, except for the offshore groups that are named after sea animals¹ in Te Reo Māori. Each family is referred to using the geographical group and a chronological number based on the most recent event's origin time. For example, family CT01 belongs to the Cape Turnagain group and the most recent event in the family occurred in 2018, later than all other events in the Cape Turnagain group. Appendix A.1 contains key information about each repeating earthquake family including its geographical group, number of events, mean

¹“Aihe” means dolphin, “petipeti” is jellyfish, and “toroa” is albatross in Te Reo Māori

location, mean magnitude, focal mechanism and location from Section 5.2. Further details of each event is provided in the QUAKEML file in the electronic appendix.

Figure 4.1: Map showing the location of the most recent event of each family and the geographical groups formed based on these locations. Black lines delineate the assigned boundaries for each group. Red stars represent some of the places used to name the groups and the red circle indicates the area of study. The box on the right contains the 16 group names and the two-letter codes used for the family nomenclature.

4.2 Repeating earthquake locations

4.2.1 Absolute locations

We calculated absolute locations using NonLinLoc (Lomax et al., 2000) for the most recent event in each family. Figure 4.2 shows the locations obtained in a map of the study area and five cross-sections perpendicular to the subduction interface. Some differences amongst the groups can be observed: for example, repeating earthquakes are most densely concentrated near the township of Porangahau, with the Taikorai Group (TK) containing 62 families, the Akitio Group (AK) containing 20, and the Mangaorapa Group (MN) 18. The Kairakau (KK) and Oueroa (OU) groups also contain a large number of families, with 33 and 10 families respectively. Other groups located further inland, such as Mataikona (MT), Dannevirke (DV) and Waipukurau (WP), have events significantly more scattered in comparison. Families belonging to the Petipeti Group (PP) stand out for having depths between 40 and 60 km, making them both the deepest events and the farthest from the subduction interface (see cross-section B-B' (Figure 4.2)). Families TT01, TT02 and DV02 are the deepest families located onshore, with mean depths between 38 and 43 km. In comparison, the two families that form the Ruahine Group (RH) have depths of 38 and 30 km and are located further inland.

Most of the family groups can be recognised as distinct in the cross-sections shown in Figure 4.2, especially in cross-sections A-A' and E-E' which contain the fewest families. Cross-sections B-B', C-C' and D-D' cross through where the majority of the families are located. These three cross-sections reveal that families belonging to the Kairakau (KK), Oueroa (OU), Taikorai (TK), Mangaroapa (MN), Akitio (AK), Toroa (TR) and Cape Turnagain (CT) groups, have similar depths to the subduction interface model obtained by Williams et al. (2013).

The uncertainties in the absolute locations are summarised in Table 4.1. These errors use the travel time residuals for the standard error and the confidence ellipsoid for the horizontal uncertainties and depth error (Lomax et al., 2000). The confidence ellipsoid is an ellipsoidal Gaussian approximation to the hypocentre Probability Density Function (PDF) at the 68% confidence level. This means that there is a 68% probability that the hypocentre is inside the ellipsoid (Lomax et al., 2000). The largest error obtained was 9.12 km and corresponds to the depth uncertainty of the most recent event of family PP02. Family PP02 is the furthest offshore and consequently the family furthest from the stations. In comparison, the most recent event of family

Figure 4.2: Map (left) and cross-sections (right) illustrating the absolute locations of the most recent event in each repeating earthquake family. The repeating earthquakes in the map are coloured by depth. In both the map and the cross-sections, the symbol sizes are scaled by each earthquake's local magnitude. Blue contours represent the subduction interface depth model of Williams et al. (2013), light blue triangles mark the stations used, black lines represent the subduction faults from (Langridge et al., 2016), and red lines mark the position of the five cross-sections. Red contours in the cross-sections indicate family groups. See Figure 4.1 for explanation of the two-letter geographical codes

Table 4.1: Summary of errors associated with the absolute locations obtained using NonLinLoc. Horizontal uncertainties and depth error are in kilometers.

	Minimum	Maximum	Average
Standard error (s)	0.18	0.77	0.43
Maximum horizontal uncertainty (km)	1.77	8.61	3.68
Minimum horizontal uncertainty (km)	1.23	5.47	2.17
Depth error (km)	1.24	9.12	2.31

DV05, located inland and well within the network, has both the minimum depth error and minimum horizontal uncertainty of 1.24 and 1.23 km, respectively.

Figure 4.3 shows the distributions of the hypocentre probability density function scatter samples and represents the uncertainty of the spatial solution arising from picking errors, travel-time calculation errors, the geometry of the observing network, and incompatibility of the picks (Lomax et al., 2000). The optimal hypocentres are shown as black crosses and correspond to the location with the minimum misfit. The scatter samples in the map and cross-sections show how locations are better-constrained for events closer to the stations. That is, the volume of the hypocentral PDF increases the further offshore the event is located.

4.2.2 Relative locations

We obtained relative locations using GrowClust (Trugman and Shearer, 2017) for the 307 events that were not picked manually. Three events were not well correlated with events outside of their family and could not be relocated; their assumed final locations are taken directly from the GeoNet catalogue. Figure 4.4 shows the absolute and relative locations in a map and five cross-sections perpendicular to the subduction interface. From the map and cross-section, we can see that the relative locations mostly overlay the absolute ones. A summary of the errors associated with the relative locations is presented in Table 4.2. The standard error refers to the root-mean-square residual differential travel time residual and the horizontal and depth errors correspond to median absolute deviations of the bootstrap distribution (Trugman and Shearer, 2017).

Subduction interface-perpendicular cross sections

Figure 4.3: Map and cross-sections of the hypocentral probability density function computed with NonLinLoc. Black crosses mark the optimal hypocenters used as the absolute locations. Blue contours represent the subduction interface depth model of Williams et al. (2013), light blue triangles mark the stations used, black lines represent faults from (Langridge et al., 2016), and red lines mark the position of the five cross-sections.

Figure 4.4: Map (left) and cross-sections (right) illustrating all the calculated locations, both absolute and relative. Earthquakes in the map are coloured by depth, circles represent absolute locations and four-point stars represent relative locations. Red five-point stars are earthquakes that could not be relocated. For both the map and cross-sections, symbol sizes are scaled by each earthquake's local magnitude. Blue contours represent the subduction interface depth model of Williams et al. (2013), light blue triangles mark the stations used, black lines represent subduction faults from (Langridge et al., 2016), and red lines mark the position of the five cross-sections.

Table 4.2: Summary of standard errors associated with the relative locations obtained using GrowClust. Horizontal and depth errors are in kilometers.

	Minimum	Maximum	Average
Standard error (s)	0.20	2.27	0.81
Horizontal error (km)	0.001	1.06	0.15
Depth error (km)	0.001	0.63	0.075

4.3 Repeating earthquake magnitudes

Section 3.6 describes the procedure we followed to calculate calibrated local magnitudes for the most recent events and relative magnitudes for the rest. The magnitudes of all the events range between M_L 1.2 and 3.9. Figure 4.5 shows the distribution of magnitudes through time and their marginal distribution in a histogram. The histogram has a left-skewed shape with the peak close to M_L 2.0, which indicates that the mean magnitude is bigger than the median and the mode. The time plot shows how the magnitudes in the repeating earthquake catalogue have varied through time, as an example, it is interesting to note that we have very few earthquakes with magnitudes smaller than M_L 2.0 before 2008. What stands out most from the time plot is the number of repeating earthquakes between late 2008 and mid 2010 and the low magnitudes of these events. This is further discussed in the next chapter.

Of the 225 families, eight contain five or more events whose magnitude variability is shown in Figure 4.6. Family HW02 (coloured purple in Figure 4.6) has a mean magnitude of M_L 1.8 with the event magnitudes having little variation through the period analysed. In comparison, families HW04 (yellow) and TK17 (olive) have magnitudes that vary significantly through time. The histogram in Figure 4.7 shows the difference between magnitudes and the family mean, with ranges between -0.6 and 0.6 . The histogram tends to be bell-shaped with a peak close to zero. This indicates that the magnitude difference's mean, median and mode are similar.

Figure 4.5: Left: Graph illustrating the changes in the local magnitude of repeating earthquakes through the 12 year period analysed in this study. Local magnitudes of all 532 events are presented. Right: histogram of earthquake magnitudes in our catalogue.

Figure 4.6: Local magnitudes of families with five or more events through time. Each family is shown with a different colour, with each dot representing an event.

Figure 4.7: Histogram illustrating the difference between the local magnitudes calculated and the mean of the family they belong to. Differences were obtained by subtracting each local magnitude from the corresponding family mean.

4.4 Repeating earthquake focal mechanisms

We obtained focal mechanisms for the most recent event of each family following the procedure described in Section 3.7 and then selected 75 of the highest-quality focal mechanisms for further analysis. The procedure for selecting the final focal mechanisms involved inspecting each stereonet of first motions and the corresponding P-T axis contour plot and identifying the ones that appeared well-constrained. The pressure (P) and tension (T) axes are geometric properties of each earthquake's focal mechanism and are directly related to the normal and slip vectors of the fault plane (McKenzie, 1969; Arnold and Townend, 2007). Due to the direct relationship between the P- and T- axes and the vectors of the fault plane, the inspection of the P-T axis probability contours was decisive for identifying well-constrained focal mechanisms.

Figure 4.8 shows two representative focal mechanisms, one well-constrained belonging to family TK02 (a) and another less constrained from family TK46 (b). A total of 10 picks, six compressional and four dilatational were used for TK02. The P-T contour plot next to the stereonet shows that the P and T axis are well constrained, as the contours are tightly-spaced and barely cross the nodal planes. For the less-constrained event belonging to family TK46, a total of eight first motions were used to calculate the focal mechanism, of which only two were compressional. The P-T contour plot of the less-constrained event from TK46, shows the P and T contours crossing several times through the nodal planes and encompassing almost all the surface of the focal sphere, indicating that the nodal planes are poorly constrained compared to those of TK02.

As an additional criterion for selecting the highest-quality focal mechanisms, a maximum scalar error of 32° was used to select the final ones. The error of all the focal mechanisms ranges from 20° to 37° with an average of 31° . In comparison, the error for the selected final focal mechanisms has a range from 20° to 32° with an average of 28° . Appendix A.4 contains the stereonets and P-T axis contour plots of the 75 final focal mechanisms.

The resulting focal mechanism catalogue is shown in Figure 4.9. The colour of each focal mechanism corresponds to the assigned position in the ternary diagram classification (Frohlich, 1992). The position of the focal mechanisms in the ternary diagram is computed using the plunge angles of the P, T and B axes. The vertices of the triangle represent pure strike-slip motion on a vertical fault, pure normal motion on a 45° -dipping fault plane and pure thrust motion on a 45° -dipping fault plane. Some families have a predominant focal mechanism. For example, Mangaroapa Group (MN) contains

almost solely normal-faulting focal mechanisms coloured in red tones, and families belonging to the Taikorai Group (TK) exhibit mostly thrust motion and some strike-slip. Other groups such as Hastings East (HE), Dannevirke (DV) and Cape Turnagain (CT) do not show a clear predominance. Toroa Group (TR) was the only group for which a high-quality focal mechanism could not be calculated.

The final focal mechanisms are also shown in five different cross-sections to illustrate their locations in relation to the subducting plate interface. The Taikorai Group (TK) has focal mechanisms with predominantly thrust motions at depths that correlate well with the subduction interface model of Williams et al. (2013). This indicates that the majority of events in this group could be located on the subduction interface or related structures. In contrast, the focal mechanisms belonging to Mangaorapa Group (MN) are predominantly normal motions and are located below the subduction interface, suggesting that they are occurring within the subducting Pacific Plate. With the exception of these two groups, there are few clear examples of a separation between events occurring on the interface, in the overlaying Australian Plate or within the subducting Pacific Plate. Further interpretations regarding the focal mechanisms are discussed in the next chapter.

Figure 4.8: a) Stereonet (left) of one well-constrained event 2018p479843 from family TK02 and scalar error of 27.2° . Solid black line indicates the focal mechanisms solution [strike = 110.1° , dip = 37.3° , rake = 31.3°]. Blue points are compressions and red points are dilatations. P- and T- probability contour plot (right) for the same event. Orange denotes P-axis and green the T-axis. b) Same as a) but for a less constrained event 2012p158837 from family TK46 with a focal mechanisms solution of strike = 143.2° , dip = 45.6° and rake = 36.6° , and scalar error of 31.4° .

Figure 4.9: Map and cross-sections of the selected focal mechanisms coloured using a ternary diagram classification (Frohlich and Apperson, 1992). Black contours represent the boundaries of the geographical groups, blue contours represent subduction interface depth from Williams et al. (2013), light blue triangles are the stations used, black lines represent faults (Langridge et al. (2016)), and red lines correspond to the five cross-sections next to the map. In map-view focal mechanisms are plotted as lower hemisphere projections and NNE-hemisphere projections in cross-sections.

4.5 Results summary

The results presented in this chapter describe repeating earthquakes near Porangahau, where the Hikurangi subduction zone transitions from creeping to a locked subduction interface. Our key results are:

1. We formed a repeating earthquakes catalogue for the area near Porangahau, from 2004 to 2018. This catalogue is composed of 532 earthquakes clustered into 225 families and then into 16 geographical groups using the locations obtained for the most recent event of each family.
2. We obtained absolute locations for the most recent event of each family and relative locations for the rest. Because the stations used are all located on land, the location uncertainties increase for the events offshore compared to those onshore. To aid discussion, we assigned geographic groupings to families. We observe that groups near the coast have more events and depths similar to the subduction interface model of Williams et al. (2013), while groups further inland had fewer events and more variable depths.
3. We calculated local magnitudes for the most recent event in each family and relative magnitudes for the rest of events. According to the histogram of all the magnitudes, most of the magnitudes have values close to M_L 2 (mode of M_L 1.9). We note that there is some variability in magnitude and recurrence of events, and discuss this further in the following chapter.
4. We selected 75 of the 225 focal mechanism obtained from first motion observations based on how well constrained the P-T axes were and the calculated scalar error. Taikorai Group (TK) families were found to have depths and thrust motions consistent with plate interface faulting, while Mangaorapa Group (MN) families indicate normal earthquakes within the subducting Pacific Plate.

Chapter 5

Discussion and Conclusions

In this chapter, we first discuss the factors that influence the completeness of our repeating earthquake catalogue (Section 5.1). In Sections 5.2 and 5.3, we review the repeating earthquake catalogue in the context of other seismic and aseismic phenomena observed on the Hikurangi margin. In Sections 5.4 and 5.5, we review our results in a global context via seismic moment and slip-rate relationships. This chapter concludes with a summary of key findings and suggestions for further research in Section 5.6.

5.1 Repeating earthquake catalogue completeness

The completeness of our final repeating earthquake catalogue is strongly influenced by the completeness of the GeoNet earthquake catalogue on which it is based, as discussed in Section 3.1. The GeoNet earthquake catalogue is constructed and maintained by GNS Science via the GeoNet project (Petersen et al., 2011). As of May 2010, the seismic monitoring network consisted of 111 regional stations and 49 stations from the New Zealand National Seismograph Networks (NZNSN; Petersen et al., 2011). Since then, there have been numerous changes in the number of stations, the sensors used to record earthquakes, the system for earthquake location, magnitude calculation, the velocity model used for location, and procedures used to pick phase arrivals. As a consequence, the GeoNet catalogue’s magnitude of completeness is known to have varied during the period spanned by this project (Figure 5.1). Duration and recurrence interval of the repeating earthquake families are strongly dependent on catalogue completeness and this is likely to have a large impact on families of small magnitudes whose events are near or below the GeoNet catalogue magnitude of completeness M_c (4.3). We observe a sudden increase in the number of repeating earthquakes identified between

Figure 5.1: Graph of the calculated magnitude of completeness (M_c) of the GeoNet catalogue near Porangahau, from 2004 to 2020. The decrease of the magnitude of completeness in 2012 is due to the change from CUSP to SeisComP for earthquake processing in GeoNet. We calculated the completeness using the maximum curvature method from Wiemer and Wyss (2000).

late 2008 and mid 2010 (Figure 4.5). The increase in events is likely to be related to the installation of stations belonging to regional networks (Figure 3.2). Stations PRHZ, ANWZ, DVHZ, PRWZ, POWZ and PNHZ were all installed at the beginning of 2008 and are located well within our study area, which would increase the number of earthquakes detected near Porangahau. However, the reason behind the decrease in identified events after late 2010 does not appear to be associated with network changes. The overview of the GeoNet project written by Petersen et al. (2011) contains no explicit mention of the detection methods used and neither are these mentioned on the GeoNet website¹, so we are not able to identify a cause related to the GeoNet catalogue. In Section 5.3 below we discuss the spatial and temporal correlation of the increase and decrease of events with other phenomena such as slow slip events (SSE).

The duration of our study period also influences the completeness of the final repeating earthquake catalogue. We chose 2004 as the start of our analysis due to the availability of continuous data from broadband and short-period stations near Porangahau, although stations from the regional networks were not installed until after 2008 (Figure 3.2). Limiting our study period from 2004 to 2018 means that we may be missing families with long-recurrence intervals. This likely affects the upper-limit of the magnitudes computed, as large-magnitude events often have long recurrence intervals (Uchida and Bürgmann, 2019).

As of March 2021, a matched-filter analysis of seismicity near Porangahau is in progress to improve the completeness of our catalogue (Calum Chamberlain, pers. comm., March 2021). Each of the repeating earthquakes identified in this thesis is being used as a template for the matched-filter. Due to time constraints and the large number of templates, we are not able to report the results from the matched-filter in this thesis. It is expected, however, that the matched-filter analysis will identify events that are not in the GeoNet catalogue and/or outside our study period.

The methodology used by GeoNet for earthquake analysis has also changed during our study period. From 1986 to 2012, GeoNet used the California Institute of Technology (Caltech)–U.S. Geological Survey (USGS) seismic processor (CUSP) system for earthquake location and magnitude estimation. Since 2012, GeoNet has used the SeisComP (SC) system (Rhoades et al., 2020a). This change in earthquake processing can be observed in Figure 5.1 with the sudden decrease in magnitudes after 2012. The identification of phase arrivals has changed accordingly: for instance, since 2016, additional manual phase analysis has been undertaken by ‘GeoHazard Analysts’, resulting in more S-picks and more manual P-picks than before 2016 (Sara O’Hagan pers.

¹<https://www.geonet.org.nz/>, last accessed March 2021.

comm.). Changes in earthquake processing have resulted in the GeoNet earthquake catalogue having heterogeneous quality locations and inconsistent magnitudes (Ristau et al., 2016) that are now being recalculated under the auspices of the National Seismic Hazard Model (Rhoades et al., 2020a,b).

The procedures described in Chapter 3 were used in order to produce a repeating earthquake catalogue with homogeneous locations and magnitudes in spite of inconsistencies in the starting GeoNet catalogue. For the most recent event in each family, we obtained accurate absolute locations and local magnitudes were calibrated with moment magnitudes. For the rest of the events, we computed relative locations and relative magnitudes. Using this approach, we obtained a repeating earthquake catalogue with homogeneous quality location and consistent magnitudes spanning the entire study period. We do not observe magnitudes below M_L 2.0 before late 2007 (Figure 4.5). The range of magnitudes obtained in our catalogue is also dependent on the smallest magnitude the GeoNet network can detect.

In terms of instrumentation, the regional networks are generally equipped with short-period seismometers, while the NZNSN stations are equipped with broadband seismometers and strong motion sensors. Additionally, sensors at some locations have been upgraded from short-period to broadband. As mentioned in Section 3.2, we corrected for instrument response to allow comparison across these changes in instrumentation.

Section 2.2.1 describes the effects of filter frequency-bands on cross-correlation coefficients. The upper limit of a suitable filter is dependent on the earthquake’s corner frequency and a suitable lower limit should consider the quarter-wavelength rule (Geller and Mueller, 1980; Uchida, 2019). We applied a 2–20 Hz band-pass filter to the cross-correlated traces: we discuss here the implications this may have for detecting repeating earthquakes using waveform similarity.

Seismic energy from earthquakes is mostly concentrated in the higher-frequency part of the earthquake source spectrum near the corner frequency (Bormann et al., 2009) and, in general, the corner frequency increases when the moment or size of an earthquake decreases (Stein and Wyssession, 2003; Allmann and Shearer, 2009). To determine the frequencies at which the seismic energy is concentrated, we computed the amplitude spectra and frequency-domain signal-to-noise ratio of three earthquakes with different calibrated local magnitudes from the repeating earthquake catalogue (Figure 5.2). The three earthquakes are part of the Mangaorapa Group (MN) and are located near the township of Dannevirke (Figure 4.1). The amplitude spectra of these earthquakes were calculated using the vertical component of station PXZ and the corresponding noise

spectra were obtained using a window of the same length as the signal before the P-pick.

For the M_L 1.6 earthquake (Figure 5.2 a), there is no clear corner frequency up to the Nyquist frequency of the data (50 Hz) and the noise and earthquake spectra have similar amplitudes. The signal-to-noise ratio is fairly constant and close to one, which means that the signal amplitudes of this low-magnitude earthquake are close to the noise amplitudes. The second earthquake is of M_L 2.3 (Figure 5.2 b) and has an amplitude spectrum clearly above the noise. The differences in amplitudes between the earthquake and noise signals are even clearer for the M_L 3.3 earthquake (Figure 5.2 c).

The spectra illustrated in Figure 5.2 contain two lines that mark the minimum and maximum frequencies of the filter used. We observe for the M_L 3.3 earthquake that the corner frequency is ~ 10 Hz and within the passband of the filter. For the M_L 2.3 earthquake, the corner frequency is approximately ~ 20 Hz which is also the frequency at which the signal-to-noise ratio starts decreasing. This means that we are still capturing most of the seismic energy. As noted above, the corner frequency cannot be clearly distinguished for the M_L 1.6 earthquake, but the theoretical corner frequencies for M_W 1.0 range from ~ 15 to ~ 100 Hz depending on the stress drop value (Allmann and Shearer, 2009). Although we are filtering below some of the theoretical corner frequencies for the smaller events, the seismic source complexity at higher frequencies is likely lost due to attenuation.

According to Geller and Mueller (1980), if the difference in the length of the propagation paths of two earthquakes is less than one-quarter of a wavelength $\lambda/4$ (the first Fresnel Zone), the waveform similarity of these events will be high even if they are not overlapping. In next paragraphs we consider how this relationship is applied to our repeating earthquakes.

We first calculated the fault radii r of the repeating earthquakes assuming a circular source geometry and a Brune earthquake source model (Brune, 1970) using Equation 5.1 (Stein and Wyssession, 2003) and Equation 3.11. M_0 is the seismic moment in dyne-cm calculated using Equation 3.11 and $\Delta\sigma$ is the stress drop in MPa. To obtain the fault radii values in meters, we converted MPa to dyne/cm² (1 MPa = 10^7 dyne/cm²) and then divided by 100 to convert centimetres to meters.

$$r = \left(\frac{7}{16} \frac{M_0}{\Delta\sigma} \right)^{\frac{1}{3}} \quad (5.1)$$

We obtained a range of source radii between 14 and 495 m using 3 and 10 MPa as rep-

Figure 5.2: Amplitude spectrograms and signal-to-noise ratios for three earthquakes of different calibrated local magnitudes. a) Earthquake 302055 with magnitude M_L 1.6. b) Earthquake 3002767 with M_L 2.3. c) Earthquake 3145133 with M_L 3.3. The input signals used to calculate the spectra were not filtered.

Figure 5.3: Calculated source radii of the repeating earthquakes using constant values of stress drops. The histogram in purple illustrates the values of source radii obtained using 3 MPa and the histogram in red illustrates the distribution of source radii using 10 MPa. The resolution limit of 48.5 m is indicated with a blue dashed line. Event 3279388 from family PP02 has the largest source radius calculated (495 m).

representative values of stress drop (Figure 5.3). Using an S-wave velocity v_s of 3.88 km/s for the region near Porangahau (Eberhart-Phillips and Bannister, 2015) and our 20 Hz high-cut frequency, we obtain a wavelength of 194 m. A quarter of this wavelength is 48.5 m, which means that events with smaller source radii could have high correlations even if they do not overlap (i.e. coincide spatially). This suggests that we are not able to separate small but not overlapping events, particularly those with larger stress-drops (see resolution limit in Figure 5.3). However, these events are small with low signal-to-noise ratios and high corner frequencies (Figure 5.2). To be able demonstrate that these low magnitude events are not overlapping, despite having high correlations, would require higher quality data with higher sampling-rates.

5.2 Repeating earthquake locations

We investigated the relationship between the repeating earthquake families and tectonic setting based on the focal mechanisms and hypocentres of earthquakes in each family. We divided the families into three groups: 1) families occurring within the Australian Plate, 2) families located on the subduction interface or 3) families occurring within the subducting Pacific Plate.

We used the ternary diagram (Frohlich, 1992) in Figure 4.9 to represent and colour the faulting type of each focal mechanism and compared the mean location of the families to the subduction plate interface model of Williams et al. (2013). If the family had a depth within 5 km of the subduction interface and thrust focal mechanism, we determined that the family was located on the subduction interface. Families with depths more than 5 km of the subduction interface, were assigned to be either in the Australian Plate or in the Pacific Plate, regardless of their faulting type. For families without focal mechanisms, we assigned them to be within the Australian or Pacific Plate based on their depth. We did not assign families to be on the interface using their depths only due to the uncertainties in the locations we obtained and the uncertainty in the subduction interface model. In other words, we used thrust focal mechanisms as a required condition for an earthquake to be assigned to the plate interface.

Using these criteria, 25 families are found to be located on the subduction interface, 19 within the Australian Plate and 181 in the Pacific Plate. Figure 5.4 shows the location of the repeating earthquake families on a map with the interface locking model and the modelled slow slip events (SSEs) from Wallace et al. (2012b). We may be underestimating the number of families on the interface, and additional work to constrain focal mechanisms using P-S ratios or waveform modelling may move some families from the Pacific Plate or Australian Plate to the interface grouping.

5.2.1 Families in the Australian Plate

19 of 225 families are located within the Australian Plate and are shown as blue circles in Figure 5.4. Five of the 19 families have a focal mechanism. Families CT13, DV12, MN14 and WP5 have strike-slip focal mechanisms and family TK27 exhibits normal faulting. Family CT13 located offshore Cape Turnagain, has the largest magnitude (M_L 2.4) of the Australian Plate families and family WP11 located near Waipukurau has the smallest magnitude of M_L 1.3. The depths of the Australian Plate families range between 17 and 6 km, all being located above the subduction interface model of

Figure 5.4: Map illustrating the locations of the repeating earthquake families in relation to the the interface locking model and the cumulative slip of slow slip events from Wallace et al. (2012b). Families in the Pacific Plate are shown in green, Australian Plate families in blue and families on the subduction interface in yellow. Red contours show the cumulative slip of slow slip events in mm, modified from Wallace (2020). The blue colours of the interseismic coupling represent a low coupling coefficient and red colours represent high coupling coefficient. The cross-section perpendicular to the subduction interface contains all the families.

Williams et al. (2013) (cross-section of Figure 5.4). 14 of the 19 families are located above the subduction interface inferred to be weakly coupled (creeping), one family (TK27) locates above the transition near Porangahau and five families (MN14, DV12, CT10, CT11 and CT13) above the locked interface. A few of the families are located nearby or exactly above the modelled SSEs. For example, families CT10, CT11 and CT13 are located above the edge of a sliding patch from the 2009 Riversdale SSE, family WP11 above the Manawatu SSEs and families KK08 and KK20 above East Coast SSEs.

5.2.2 Families in the Pacific Plate

Most of the families in our repeating earthquake catalogue are inferred to lie within the Pacific Plate (green circles in Figure 5.4). 45 of the 181 families within the Pacific Plate have a focal mechanism: five thrust, 21 strike-slip and 19 normal faulting focal mechanisms.

These families have a wide range of magnitudes and depths. Families PP01 and PP02 are the furthest offshore, have the largest magnitudes (M_L 3.6 and 3.1) and are deeper (> 40 km) than all other repeating earthquake families. The smallest magnitude of the Pacific Plate group is M_L 1.3 from family AK14 located near Akitio and the shallowest family is TR01 with a depth of 12 km located on the coast off Porangahau.

Families in the Pacific Plate are located beneath both the creeping and locked part of the subduction interface, with no observable distribution determined by the coupling state of the interface. In relation to the slow slip events, all families from the Toroa group (TR) are located close to the subduction interface and on the edge of the East Coast SSEs modelled slip. All the families from the Akitio group (AK) are also located close to the subduction interface, where a patch of the 2009 Riversdale slow slip event occurred (Figure 5.4). Other Pacific Plate families are located below the Kaimanawa and Manawatu SSEs, but with depths that clearly locate them below the subduction interface model. Temporal association to slow slip events is addressed in the next section.

Although several of the families within the Pacific Plate have depths close to the subduction interface modelled by Williams et al. (2013), they do not have a well-constrained focal mechanisms to determine if they were occurring on the interface.

5.2.3 Families on the subduction plate interface

25 families are located on the subduction interface with all of them presenting thrust focal mechanisms (yellow circles in Figure 5.4). Family TK44 has the largest magnitude of M_L 2.9, and family TK46 has the smallest magnitude of M_L 1.6. Both TK44 and TK46 families are located near the township of Porangahau.

Only families KK09 and HW04 are located on the creeping part of the subduction interface, the other 23 families are located near the transition from creeping to locked, near Porangahau. In relation to the slow slip events in the area, family HW04 is located where the deep Kaimanawa SSEs occurred (Wallace and Eberhart-Phillips, 2013) and families CT03, CT05, CT12 and AK13 are located in the area where the 2009 Riversdale SSE occurred. Further temporal analysis of interplate families and SSEs is discussed in the next section.

5.3 Repeating earthquakes and other phenomena

We analysed the spatial and temporal relationships between our catalogue of repeating earthquakes and nearby slow-slip episodes and large earthquakes (Figures 5.5 and 5.6). Table 5.1 contains a summary of the large earthquakes and slow slip events compared with the repeating earthquakes. The dates and location of the large earthquakes were obtained from the Institute of Geological and Nuclear Sciences (GNS) website² and the GeoNet website³. Start and end dates of some slow slip events were obtained via by Andrea Pérez (pers. comm., March 2021). Dates for the 2014/2015 Manawatu SSE, the 2006 and 2008 deep Kaimanawa SSEs and the 2009 Riversdale SSE were inferred using information and time-series from Wallace and Beavan, 2010; Wallace et al., 2012a; Eberhart-Phillips et al., 2013; Wallace et al., 2018 and Wallace, 2020. The slip contours used to make spatial comparisons are from Wallace (2020) and do not represent individual slow slip events, except for the 2009 Riversdale SSE. In this thesis we opted for using these cumulative contours to represent individual slow slip events because not all have published contours and they also migrate spatially. It is also important to consider that the contours have different slip limits (20 for Kaimanawa SSEs, 5 for Riversdale SSE and 100 for Manawatu and East Coast cumulative contours) for the spatial correlations. To enable more robust comparison between repeating

²<https://www.gns.cri.nz/Home/Learning/Science-Topics/Earthquakes/New-Zealand-Earthquakes/Where-were-NZs-largest-earthquakes>, last accessed April 2021

³<https://www.geonet.org.nz/>, last accessed April 2021.

earthquakes and slow slip events, a more detailed geodetic catalogue is required.

We took into consideration the duration of the families in our analysis and divided them into continual- and burst-type, as burst-type events are thought to be related to different seismic processes than continual-type (Igarashi et al., 2003; Kimura et al., 2006; Mesimeri and Karakostas, 2018). Burst-type repeating earthquakes are often associated with swarms and aftershocks of large earthquakes (Igarashi et al., 2003). Continual-type on the other hand, are associated with periodic loading and rupture of the same asperity and reflect the aseismic slip of the region (Nadeau and Johnson, 1998; Nadeau and McEvilly, 1999).

Previous studies have reported connections between repeating earthquakes and the occurrence of slow slip events (e.g. Gardonio et al., 2018; Kato et al., 2012, 2014; Chen et al., 2013) by analysing long repeating earthquake catalogues and GNSS time-series. There has been evidence of increases in the number of repeating earthquakes during slow slip events, suggesting that repeating earthquakes are sensitive to the occurrence of nearby slow slip events (Gardonio et al., 2018). Slow slip events in the Hikurangi margin are diverse: shallow (<15 km), short (<1 month) and frequent slow slip occurs in the northern and central margin and deeper (>30 km), longer (>1 year) and less frequent events occur in the southern margin (Wallace, 2020, and references therein).

The Riversdale 2009 slow slip event occurred in the strongly-coupled region of the Hikurangi margin, southeast of our area of study (Wallace et al., 2012b) (Figure 1.3). Three families (AK03, AK14 and AK15) can be spatially and temporally correlated with the Riversdale slow slip event (Figure 5.7). Family AK03 contains two events, both located within the SSE patch. The first event occurred while the SSE was in process (see time-series in Figure 5.7) and the second and most recent event from family AK03 occurred in 2017 with no clear temporal association with SSEs. Families AK14 and AK15 are burst-type families with events having hours of difference between them and occurring a week after the Riversdale SSE (see time-series of Figure 5.7). Close in time to families AK14 and AK15, several burst-type events occurred, although these are not located close to the Riversdale modelled slip patch. Two interplate events occurred close in time to the Riverside SSE, but with no strong spatial correlation to the slow slip patch. The burst-type interplate family TK53 occurred when the slow slip event was in process, but the events are not located near the slow slip patch. Interplate family HW04 occurred further north from the SSE and almost two months after the Riversdale SSE started (Figure 5.7).

The families from the geographical group Akitio (AK) (Figures 4.1 and 5.4) are located

Figure 5.5: Local magnitudes of continual- and burst-type repeating events through time with seismic and aseismic phenomena. Main figure: time-series with events from continual families as green circles and from burst-type families in pink. Displacement is shown for GeoNet GNSS PORA station in light blue. Slow slip events are shown as shaded areas and large earthquakes are marked as red dashed lines. Top histogram: Frequency of continual- and burst-type events through time (bars overlapping). Right histogram: Frequency of the local magnitudes for continual- and burst-type events (bars overlapping).

Figure 5.6: Cumulative moment of continual- and burst-type repeating events through time with seismic and aseismic phenomena. Cumulative moment for burst-type families is shown in pink, interplate continual-type in yellow, Pacific Plate continual-type in green and Australian Plate continual-type in blue. Slow slip events are shown as shaded areas and large events are marked as vertical lines of different colours according to the legend.

Table 5.1: Dates and locations of the seismic and aseismic phenomena to which we are spatially and temporarily comparing our catalogue of repeating earthquakes.

Event	Date	Location	Type	Reference
Manawatu SSE	2004 – 2005	Manawatu region	SSE	Wallace and Beavan (2010)
Kaimanawa SSE	2006	Beneath the Kaimanawa ranges	SSE	Wallace and Eberhart-Phillips (2013)
East Coast SSE	June 2006	South Hawkes Bay	SSE	Wallace and Beavan (2010)
East Coast SSE	August 2006	South Hawkes Bay	SSE	Wallace and Beavan (2010)
Gisborne Earthquake M6.8	2007-12-20	Offshore Gisborne	Earthquake	GNS website
East Coast SSE	March 2008	South Hawkes Bay	SSE	Wallace and Beavan (2010)
Kaimanawa SSE	2008	Beneath the Kaimanawa ranges	SSE	Wallace and Eberhart-Phillips (2013)
Dusky Sound Earthquake M7.8	2009-07-15	Fiordland	Earthquake	GNS website
Riversdale SSE	Dec 2009	Offshore Riversdale	SSE	Wallace et al. (2012b)
Darfield Earthquake M7.1	2010-09-04	Canterbury	Earthquake	GNS website
Manawatu SSE	2010 – 2011	Manawatu region	SSE	Wallace et al. (2012b)
Christchurch Earthquake M6.3	2011-02-22	Christchurch, Canterbury	Earthquake	GNS website
Cook Strait Earthquake M6.5	2011-07-21	Cook Strait	Earthquake	GNS website
East Coast SSE	Jun – Jul 2011	Cape Turnagain	SSE	Wallace et al. (2012b)
East Coast SSE	Aug – Sept 2011	Hawkes Bay	SSE	Wallace et al. (2012b)
East Coast SSE	Feb – Mar 2013	Cape Turnagain/Hawkes Bay	SSE	Wallace and Eberhart-Phillips (2013)
Grassmere Earthquake M6.6	2013-08-16	Seddon	Earthquake	GNS website
Eketahuna Earthquake M6.2	2014-01-20	Eketahuna	Earthquake	GNS website
Manawatu SSE	2014 – 2015	Manawatu region	SSE	Wallace et al. (2018)
Pongaroa Earthquake M5.7	2015-10-12	Pongaroa, near Porangahau	Earthquake	GeoNet catalogue
Te Araroa Earthquake M7.1	2016-09-02	Offshore East Cape	Earthquake	GNS website
Kaikōura Earthquake Mw 7.8	2016-11-13	Kaikōura	Earthquake	GNS website
East Coast SSE	Nov 2016	Hawkes Bay	SSE	Wallace et al. (2018)

where the Riversdale slip patch is and where an imbricate wedge of high reflectivity materials has been observed (Barker et al., 2009). But only the three previously mentioned families (AK03, AK14 and AK15) can be temporally correlated to the slow slip event. This indicates that the Akitio (AK) families are well correlated spatially with the Riversdale slow slip event. But there is not a strong temporal correlation, as only three of the 40 events from Akitio Group occur during or within a few days after the slow slip event. Jacobs et al. (2016) observed an increase in seismicity south of the slow slip event, which is outside our study area. As these observations are outside our area of study, we are not able to further relate the observations of Jacobs et al. (2016) to repeating earthquakes.

Several slow slip events have occurred offshore between Cape Turnagain and southern Hawkes Bay: we refer to these events as East Coast slow slip events and we use the cumulative contour from Wallace (2020) to represent all of them. Figure 5.4 shows that most of the repeating earthquake families are located down-dip of the total modelled slip of East Coast SSEs from 2002 to 2016 from Wallace (2020). Few families in the Pacific and Australian Plates correlate spatially with the slip patch, but only families TR01, KK08 and KK04 have spatial and temporal correlations to East Coast SSE. Family TR01 is located within the Pacific Plate, with a similar depth to the subduction interface model by Williams et al. (2013) (see Figures 5.8 and 4.2) and contains four events. Two of the events in that family occur during two East Coast SSEs: the 2006 and 2008 South Hawkes Bay events. One of the three events in family KK08 has temporal correlation with the 2011 Cape Turnagain SSE. Family KK08 is not indicated in Figure 5.8 as only events occurring temporally close to the 2016 East Coast SSE are shown. Family KK04 is located on the edge of the modelled slip patch with two of four events correlating temporally to the 2013 and the 2016 slow slip events. The three mentioned families (TR01, KK08 and KK04) have more than one event that occurs during slow slip events, have depths similar to the subduction interface of Williams et al. (2013) and can be spatially correlated to the modelled slip patches. These temporal and spatial correlations suggest that some repeating earthquake families might have been triggered by the East Coast SSEs on more than one occasion.

The time-series in Figure 5.8 shows several events occurring during the 2016 East Coast SSE that are located down dip of the slip patch. These events are concentrated near the township of Porangahau and almost on the edge of the modelled slip patch. TK30 is the only interplate family occurring during the SSE and contains two events which occurred in 2010 and 2016. Families KK03, TK25 and TK26 occur in the days after the slow slip event and are also located down-dip the slip patch. Family TK27 is the only burst-type family observed temporally near the 2016 M7.8 Kaikōura earthquake

Figure 5.7: Time-series (top) and map (bottom) illustrating the repeating earthquakes that occur close in time to the 2009 Riversdale slow slip event. Events on the subduction interface are coloured in yellow and events within the Pacific Plate in green. The duration of the Riversdale slow slip event is indicated in green on the time-series and the spatial location of the modelled slip patch is marked in red contours on the map. The labelled events are the ones mentioned in the text.

and the triggered East Coast SSE (time-series in Figure 5.8). Burst-type repeating earthquakes are frequently interpreted as arising from stress changes due to large nearby events (Igarashi et al., 2003). Yao et al. (2021) identified a strong increase in seismicity 8 days after the Kaikōura earthquake around Porangahau which temporally correlates with five of the repeating earthquakes occurring during the slow slip event (including TK30 and TK27). However, we do not observe an increase in burst-type or continual-type families after the duration of the SSE. Wallace et al. (2017) identified repeating earthquakes probably driven by the 2016 East Coast SSE, with 155 of their events also in the repeating earthquake catalogue obtained in this study (Laura Wallace, pers. comm., May 2021). When comparing the locations of repeating earthquakes, both catalogues show concentration of events near the township of Porangahau and station ANWZ. A notable difference between the two catalogues is the concentration of events near Eketāhuna in the catalogue of Wallace et al. (2017) which are not present in our repeating earthquake catalogue. Another difference is the number of families in their catalogue with more than ten events, in contrast with the largest family in our catalogue having nine events. We were not able to do a more detailed comparison as we were given the list of all events not divided by family.

In terms of cumulative moment, Figure 5.6 shows an increase in cumulative moment from Pacific Plate continual-type families for the 2008 South Hawkes Bay SSE, from Australian Plate continual families for the 2011 June-July Cape Turnagain SSE, and from continual-type interplate families for the 2013 and 2016 East Coast SSEs.

Manawatu slow slip events last between 1–2 years, occur every 4–5 years and are located at depths >25 km on the subduction interface (Wallace and Beavan, 2010; Wallace et al., 2018; Wallace, 2020). In this thesis, we use the cumulative contour from Wallace (2020) to represent all of the individual slow slip events. Figure 5.4 shows that most of our repeating earthquake families are located up-dip of the Manawatu slow slip patch contours. For example, Figure 5.9 shows that the events occurring close in time to the 2010/2011 Manawatu SSE are mostly occurring up-dip of the slip-patch, except for family RH01. Family RH01 is spatially correlated with the slip patch and has one event occurring before the SSE and the other event during (see time-series in Figure 5.9). Several repeating earthquakes are correlated temporally to this slow slip event, but only few are located on the outside of the slip patch contours and only family RH01 is well correlated spatially to the region of highest slip. This does not necessarily mean that the occurrence of these repeating earthquakes is not related to the SSE. As we mentioned in Section 5.1, the network upgrade of 2008 might be the cause of the apparent increase in identified events. Therefore, it is not possible to clearly indicate if repeating earthquakes during these years are casually linked to slow slip events.

Figure 5.8: Time-series (top) and map (bottom) illustrating the repeating earthquakes that occur close in time to the 2016 triggered South Hawkes Bay slow slip event. Events on the interface are coloured in yellow, events within the Pacific Plate are green and events in the Australian Plate are blue. The duration of the East Coast slow slip event is indicated in grey on the time-series and the spatial location of the modelled slip patch is marked in red contours on the map. The 2016 M7.8 Kaikōura earthquake is indicated with a red vertical line. The labelled events are the ones mentioned through the text.

As can be observed in the top histogram from Figures 5.5 and 5.9, the number of events identified decreases just before the 2010/2011 Manawatu slow slip event. Figure 5.6 shows an almost constant cumulative moment during this SSE for both the burst-type events and the continual-type interplate and Pacific Plate events. We observe an increase in the continual-type Australian Plate cumulative moment, most likely due to the events occurring during the SSE (Figure 5.9). For the 2014/2015 event, the cumulative moment of all types of events stays constant during the slow slip. We do not observe an decrease of seismicity before the 2014/2015 SSE analogous to that seen before the 2010/2011 and there are insufficient data to analyse the earlier 2004/2005 SSE. We were not able to identify a network or catalogue related explanation for the decrease of events. It is also interesting to note that most of the burst-type earthquakes occur between late 2008 and mid 2010, but do not show a clear spatial distribution (e.g. Figure 5.9). Therefore, we are not able to correlate them with known spatial phenomena and characteristics of the Hikurangi margin such as coupling of the interface, slow slip events, large earthquakes in the area or faults.

Deep Kaimanawa SSEs occurred in 2006 and 2008 beneath the Kaimanawa Ranges, down-dip of the shallow East Coast SSEs Wallace and Eberhart-Phillips (2013). The repeating earthquake families that correlate spatially to the SSEs are mostly from the Hastings West (HW) and Hastings East (HE) groups. Most of the families from these groups are located within the Pacific Plate, but close to the interface. We observe an increase of burst-type earthquakes from the Hasting East (HE) Group in August 2008, a month after the end date we inferred for the 2008 Kaimanawa SSE (see time-series of Figure 5.10). Family HW02 is the only family that correlates spatially and has one event occurring during the slow slip.

Although tremor has been identified in our study area (Ide, 2012; Romanet and Ide, 2019) near Cape Turnagain, we do not observe a clear temporal relationship between tremor occurrence and number of events or cumulative moment.

Repeating earthquakes have been found to be triggered by nearby large earthquakes (e.g. Chen et al., 2013; Kimura et al., 2006; Templeton et al., 2008; Nadeau et al., 1995; Igarashi et al., 2003). We analysed the temporal relationship between our repeating earthquake catalogue and the largest earthquakes that occurred in New Zealand during our study period (Figures 5.8 and 5.6). The only clear temporal relationship was with the M7.8 Kaikōura earthquake, which occurred in November 2016. As described earlier, we found 10 events in the month following the Kaikōura earthquake (Figure 5.8), two from a burst-type family (TK27). Additionally, we observe a small increase in the cumulative moment in interplate and Pacific Plate families after the Kaikōura

Figure 5.9: Time-series (top) and map (bottom) illustrating the repeating earthquakes that occur close in time to the 2010/2011 Manawatu slow slip event. Events on the interface are coloured in yellow, events within the Pacific Plate are green and events in the Australian Plate are blue. The duration of the 2010/2011 Manawatu slow slip event is indicated in yellow on the time-series and the spatial location of the modelled slip patch is marked in red contours on the map. The labelled events are the ones mentioned in the text.

Figure 5.10: Time-series (top) and map (bottom) illustrating the repeating earthquakes that occur close in time to the 2008 Kaimanawa slow slip event. Events on the interface are coloured in yellow and events within the Pacific Plate are green. No events in the Australian Plate are observed in this time period. The duration of the 2008 Kaimanawa slow slip event is indicated in light red on the time-series and the spatial location of the modelled slip patch is marked in red contours on the map. The labelled events are the ones mentioned in the text.

earthquake and during the triggered 2016 East Coast SSE. This increase in cumulative moment results from the events showed in Figure 5.8.

There is a significant increase in cumulative moment from interplate families in late 2012 that has no temporal correlation with any known seismic or aseismic occurrences (Figure 5.6). We also observe a step in the cumulative moment of Pacific continual-type families in late 2012 and of continual interplate families in mid 2018 that also correlate with displacements of ~ 20 mm at the GNSS station PORA time signal 5.6. From the top histogram in Figure 5.5, we observe an increase in the frequency of the burst-type events. While this increase is not as big as the events between late 2008 and mid 2010, it clearly stands out from the previous years. There are no reported seismic or aseismic phenomena that correlate temporarily with the observations mentioned in this paragraph.

There are a variety of slow slip events that occur in our area of study near Porangahau. Even though some repeating earthquakes correlate spatially and temporally with slow slip events, Figure 5.4 shows that most of the repeating events are located outside the slip patches, except for the 2009 Riversdale SSE. Taikorai Group (TK), located near the township of Porangahau, has the largest number of families and they are mostly located on the transition from strongly- to weakly-coupled subduction interface. The scarce and sparse temporal and spatial correlations between slow slip events and repeating earthquakes indicate that repeating earthquakes are not solely driven by slow slip events. But there is also evidence that slow slip events in the Hikurangi margin have triggered seismicity (Delahaye et al., 2009; Jacobs et al., 2016), burst-type repeating earthquakes (Shaddock and Schwartz, 2019; Warren-Smith et al., 2018) and repeating earthquakes triggered by a slow slip event (Wallace et al., 2017). While some repeating earthquakes may be triggered by dynamic or static stress changes for nearby SSEs or earthquakes, others may not be, because they have not experienced sufficient stress accumulation since their last rupture to be triggered again. The triggering of repeating earthquakes would entail, based on the elastic rebound theory Reid (1910), that the shear stress accumulated on the asperity patch is close to τ_1 (Figures 2.3 and 2.4) when the stress perturbation due to slow slip events occur.

5.4 Seismic moment–recurrence interval relationship

Nadeau and Johnson (1998) developed a scaling relationship for repeating earthquakes in California using seismic moment and recurrence interval. This relationship is based on the presumption that repeating earthquakes are the result of cyclic loading (Section 2.3), which means that events with larger magnitudes are expected after longer recurrence interval times. The relationship for repeating earthquake families occurring at Parkfield, California stands as:

$$\log(Tr) = 4.85 + 0.17 \log(M_0) \quad (5.2)$$

Here Tr is the recurrence interval in seconds and M_0 is the seismic moment in dyne-cm obtained using the relationship derived by Hanks and Kanamori (1979) (Equation 3.11). This seismic moment–recurrence interval relationship appears to hold in other regions including Japan, Taiwan, Tonga-Kermadec-Vanuatu, and Mexico (e.g. Chen et al., 2007; Chen and Lapusta, 2009; Yu, 2013; Schaff and Richards, 2011; Dominguez et al., 2016; Uchida et al., 2019).

As mentioned in Section 2.3.1, burst-type repeating earthquakes cannot be explained by constant loading-rate models due to their short family duration and inter-event times. For this reason we only used continual-type repeating earthquake families, which we established as having a family duration larger than a year. We established this duration based on arguments presented by Mesimeri and Karakostas (2018) and Hughes et al. (2021). This criterion yields 116 continual-type families and 107 burst-type families (Figure 5.5). Figure 5.11 shows the moment–recurrence interval relationship of the continual-type repeating earthquake families near Porangahau, separated based on location, along with results from different studies. To allow comparison of $Tr - M_0$ relationships, we normalized the recurrence interval by the ratio of plate convergence rates between California and the Hikurangi margin, using a convergence rate of 40 mm/yr for the region near Porangahau and 23 mm/yr for Parkfield (Chen et al., 2007). Then, we fitted a linear relationship to the logarithms of Tr and M_0 by minimising the L1 norm. Table 5.2 summarises the regression models of the calculated $Tr - M_0$ relationships with 95% confidence intervals.

The families located on the subduction interface present a linear relationship with slope similar to the relationship from Parkfield data, but the 95% confidence intervals of the regression model are significantly higher than the ones obtained by Nadeau

Figure 5.11: Relationship between moment and recurrence interval for repeating earthquake families near Porangahau and other published studies. Families near Porangahau are separated based on whether they are located in the Australian Plate (blue), within the Pacific Plate (green) or on the subduction interface (yellow). The yellow line represents the fitted regression line for the continual-type Porangahau interplate families and the dashed line the scaling relationship from Parkfield data Nadeau and Johnson (1998); Chen et al. (2007). Figure modified from Uchida (2019).

and Johnson (1998). The families located in the Australian Plate and Pacific Plate yield regression models considerably different from the Nadeau and Johnson (1998) relationship, especially the families in the Pacific Plate which are fitted by an almost horizontal line with slope close to zero and large 95%.

Even though the uncertainty associated with the regression models is high, the repeating earthquakes near Porangahau present similar values to some of the NE Japan (blue circles) and Central Taiwan (red triangles) repeating earthquakes (Figure 5.11). This suggests that the calculated values of recurrence interval and seismic moment for repeating earthquakes near Porangahau are within the ranges of global observations and the processes responsible for the recurrence of repeating earthquakes near Porangahau are similar to the processes from other settings (Chen et al., 2007). However, this does not mean that the relationships derived by Nadeau and Johnson (1998) using repeating earthquakes in California holds perfectly for repeating earthquakes near Porangahau, as verified by the values and quality of the linear fits (see Table 5.2). Therefore, results from this and other relationships (e.g. Equation 5.3) derived from repeating earthquakes in other settings need to be interpreted with caution.

Table 5.2: Summary of the seismic moment–recurrence interval relationship identified for repeating earthquakes near Porangahau compared to the identified by Nadeau and Johnson (1998). Intercepts are in seconds (s) to allow comparison to the relationship derived by Nadeau and Johnson (1998). 95% confidence intervals for the gradient and intercept are included in the brackets.

Location	Intercept	Gradient
Australian Plate	5.88 [5.32–6.44]	0.12 [−0.67–0.91]
Interface	5.085 [4.47–5.69]	0.16 [−0.06–0.38]
Pacific Plate	8.20 [7.95–8.45]	0.006 [−0.10–0.11]
Nadeau and Johnson (1998): Parkfield, California	4.85 ± 0.16	0.17 ± 0.009

5.5 Slip–rate relationships

The cumulative slip associated with repeating earthquakes is thought to reflect the slip history of the aseismic slip in the region surrounding the asperities (Nadeau and Johnson, 1998; Igarashi et al., 2003; Chen et al., 2007; Yu, 2013). Therefore, several studies have used repeating earthquakes to estimate the spatio-temporal distribution of interplate creep. Repeating earthquakes have been used to estimate creep on faults in several studies (e.g. Nadeau and Johnson, 1998; Uchida et al., 2003; Chen et al., 2007; Igarashi et al., 2003; Li et al., 2011; Yu, 2013; Mavrommatis et al., 2015; Dominguez et al., 2016). Nadeau and Johnson (1998) developed a relationship between slip (d) and moment (M_0) based on repeating earthquakes in California:

$$\log(d) = -2.36 + 0.17\log(M_0) \quad (5.3)$$

where slip, d is given in centimetres and M_0 in dyne-cm. We applied this relationship to 19 continual-type families located on the subduction interface. We then divided the obtained slip value by the duration of each family in years and converted units to mm to obtain a slip rate in mm/yr. The slip-rate values are shown in Table 5.3 their distribution is shown in Figure 5.12.

At the Hikurangi margin, the Pacific Plate is subducting under the Australian Plate at convergence rate that decreases from ~ 54 mm/yr near offshore Raukumara Peninsula to ~ 32 mm/yr offshore Wairarapa. For the region near Porangahau, the convergence rate at the trench is ~ 40 mm/yr. Family TK44, located near the township of Porangahau, has a remarkably high slip-rate value of 40 mm/yr in comparison to the rest of the families, which average 14 mm/yr (Figure 5.12). Most of the families are located close to each other near the township of Porangahau, which means that we are not able to observe spatial variations in slip-rates. The values we calculated, except for

family TK44, are lower than the convergence rate. As mentioned in Section 5.4, the results obtained using the relationships proposed by Nadeau and Johnson (1998) must be interpreted with caution as they were developed using earthquakes in California. Moreover, although the linear recurrence interval-seismic moment fitting model for interplate families has a slope similar to that proposed by Nadeau and Johnson (1998), the 95% confidence intervals are wide. This suggests the possibility that the repeating earthquakes near Porangahau might not be capturing the whole slip-rate when using the relationship developed by Nadeau and Johnson (1998).

Other proposed repeating earthquake models take into consideration an aseismic component or variable slip-rates. The model proposed by Beeler et al. (2001), contains an aseismic component in the asperities, where asperities continuously creep and strain harden until reaching a failure threshold and the models by Mavrommatis et al. (2015) and Nomura et al. (2017) include variable slip-rates. Future modelling of slip of repeating earthquakes near Porangahau considering an aseismic component within the asperities or variable slip-rates, could complement the results obtained using Nadeau and Johnson (1998)'s relationship.

Regarding the convergence rate near Porangahau, the ~ 40 mm/yr previously mentioned includes the margin-normal motion and the margin-parallel motion. According to Nicol and Wallace (2007), at the Hikurangi margin, $\sim 85\%$ of the margin-normal motion is accommodated on the plate interface and $\geq 60\%$ of the margin-parallel motion is accommodated by strike-slip faulting and vertical-axis rotations in the upper plate. We decomposed the ~ 40 mm/yr plate convergence vector into margin-normal and margin-parallel motion using an angle of 51° between the Pacific Plate motion vector and the trench offshore Porangahau. Adding the 65% of the normal-margin motion and 40% of the parallel-margin motion results in ~ 28 mm/yr being accommodated on the plate interface near Porangahau. Although a value of ~ 28 mm/yr of plate convergence is closer than ~ 40 mm/yr to the average slip-rate obtained from repeating earthquakes, these results need to be interpreted with caution due to the poorly constrained values used (e.g. percentages of normal and parallel motion, and angle between the trench and motion vector). As not all the plate convergence motion is assumed to be on the subduction interface (Nicol et al., 2007; Nicol and Wallace, 2007) and other structures might also accommodate part of the convergence motion, comparison between slip-rates from repeating earthquakes and plate converge is limited.

The slip-rate value of family TK44 is close to the ~ 40 mm/yr convergence rate, but it is also important to note that this family only has two events and has the shortest family duration of all the families with calculated slip-rates. As mentioned in Section 5.1, the

Figure 5.12: Box plot illustrating the distribution of slip-rate values obtained from repeating earthquakes near Porangahau. Slip-rates were calculated using the slip-seismic moment relationship derived by Nadeau and Johnson (1998). Each circle represents a family, with outlier family TK44 in red.

completeness of the GeoNet catalogue affects the completeness of the repeating earthquakes catalogue, including the duration of the families. The matched-filter analysis that is currently in process will identify any missing events within families, providing with reliable family duration's to further analyse slip-rates.

Table 5.3: Slip-rates calculated from the slip-seismic moment relationship derived by Nadeau and Johnson (1998). N indicates the number of events in the family.

Family ID	N	Slip-rate (mm/yr)	Family Duration (yr)
CT3	2	11	7.6
CT5	3	7	12.0
DV5	3	13	7.3
DV9	4	14	6.7
HW4	5	12	7.0
KK9	3	14	4.8
TK2	3	8	9.7
TK13	2	20	6.2
TK18	2	13	5.6
TK19	3	7	8.7
TK22	2	9	10.0
TK23	4	11	7.4
TK30	2	14	6.7
TK33	2	12	6.7
TK40	2	23	4.7
TK42	2	16	4.4
TK43	2	16	6.9
TK44	2	40	3.3
TK46	4	18	3.5

5.6 Conclusions and future work

We successfully identified 532 repeating earthquakes occurring between 2004 and 2018 near Porangahau using a multisegment normalized cross-correlation threshold of 0.95 on three or more stations. The identified repeating earthquakes were clustered into 225 families using hierarchical clustering and then separated into 16 geographical groups. The final repeating earthquake catalogue includes precise absolute and relative locations, local magnitudes calibrated with moment magnitude and high-quality focal mechanisms. We are aware of the effects of the completeness of the GeoNet catalogue in the duration and recurrence intervals of the identified repeating earthquake families. Events identified with the matched-filter, that is currently in process, will provide better constrained duration and interval times.

With precise locations and well-constrained focal mechanisms, we determined the repeating earthquake families to occur within the Pacific Plate, Australian Plate or on the subduction interface. Most of the families are located in the Pacific Plate and although some are close to the subduction interface model of Williams et al. (2013), most do not have a thrust focal mechanism required to conclude that they might be on the subduction interface. Additional work to constrain focal mechanisms using P-S ratios or waveform modelling may move some families from the Pacific Plate or Australian Plate to the interface grouping. Most of the interplate families belong to Taikorai group (TK) and are located in the transition between locked and creeping, near the township of Porangahau.

We observe that most of our repeating earthquake families are located down-dip of the East Coast SSEs and up-dip of the Manawatu SSEs, outside the modelled slip contours. The Akitio (AK) families are well correlated spatially with the Riversdale slow slip event as they are located where the 2009 Riversdale slip patch is. Moreover, there is not a strong temporal correlation, as only three of the 40 events from Akitio Group occur during or just after the slow slip event. The scarce and sparse temporal and spatial correlations between slow slip events and repeating earthquakes, indicate that repeating earthquakes are not solely driven by slow slip events. Although we are not able to discriminate between triggered and non-triggered repeating earthquakes, studies have observed triggered seismicity (Delahaye et al., 2009; Jacobs et al., 2016) and burst-type repeating earthquakes (Shaddox and Schwartz, 2019; Warren-Smith et al., 2018) due to slow slip events in the Hikurangi Margin. Temporal and spatial correlations are strongly dependent on the start and end dates of slow slip events and on the contours plotted. The actual slip patches of slow slip events might encompass

smaller or larger areas, which adds uncertainty to the possible correlations between slow slip events and repeating earthquakes. Overall, there are no clearly constrained spatial or temporal patterns between slow slip events and repeating earthquakes for the region near Porangahau.

In relation to large earthquakes, we identified 10 events in the month following the Kaikōura earthquake and observe a small increase in the cumulative moment of interplate and Pacific Plate families after the Kaikōura earthquake and during the triggered 2016 East Coast SSE.

From the recurrence interval–seismic moment relationship, we observe that the linear fit of continual-type interplate events has a slope similar to that of the relationship derived by Nadeau and Johnson (1998), but with high uncertainties. For the slip-rates obtained, we are not able to perform a spatial analysis due to the clustered distribution of interplate families. Excluding family TK44, the slip rates obtained are significantly lower than the plate converge rate for the region. There is the possibility that by using the relationships from Nadeau and Johnson (1998), repeating earthquakes near Porangahau are not representing the whole slip-rate of the plate interface. Other proposed repeating models, take into consideration an aseismic component or variable slip-rates. Future modelling of slip during repeating earthquakes near Porangahau might consider including an aseismic component or variable slip-rates (Beeler et al., 2001; Mavromatis et al., 2015; Nomura et al., 2017) the amount of motion accommodated on the plate interface. Repeating earthquakes do not show a distribution determined by the coupling state of the interface.

This study confirms the existence of repeating earthquakes near Porangahau and provides the first comprehensive spatiotemporal analysis of repeating earthquakes in relation to seismic and aseismic phenomena near Porangahau. The insights gained from this study lay the groundwork for future work constraining processes of stress accumulation at the creeping to locked transition zone near Porangahau.

Bibliography

- Abercrombie, R. E. (2014). Stress drops of repeating earthquakes on the San Andreas Fault at Parkfield. *Geophysical Research Letters*, 41(24):8784–8791.
- Aki, K. (1984). Asperities, barriers, characteristic earthquakes and strong motion prediction. *Journal of Geophysical Research*, 89(B7):5867–5872.
- Allmann, B. P. and Shearer, P. M. (2009). Global variations of stress drop for moderate to large earthquakes. *Journal of Geophysical Research: Solid Earth*, 114(1):1–22.
- Arnold, R. and Townend, J. (2007). A Bayesian approach to estimating tectonic stress from seismological data. *Geophysical Journal International*, 170(3):1336–1356.
- Aster, B. Y. R. C. and Scott, J. (1993). Comprehensive characterization of waveform similarity in microearthquake data sets. *Bulletin of the Seismological Society of America*, 83(4):1307–1314.
- Avouac, J. P. (2015). From geodetic imaging of seismic and aseismic fault slip to dynamic modeling of the seismic cycle. *Annual Review of Earth and Planetary Sciences*, 43:233–271.
- Ballance, P. F. (1976). Evolution of the Upper Cenozoic magmatic arc and plate boundary in northern New Zealand. *Earth and Planetary Science Letters*, 28(3):356–370.
- Bar-Joseph, Z., Gifford, D. K., and Jaakkola, T. S. (2001). Fast optimal leaf ordering for hierarchical clustering. *Bioinformatics*, 17.
- Barker, D. H., Sutherland, R., Henrys, S., and Bannister, S. (2009). Geometry of the Hikurangi subduction thrust and upper plate, North Island, New Zealand. *Geochemistry, Geophysics, Geosystems*, 10(2).
- Barnes, P. M., Lamarche, G., Bialas, J., Henrys, S., Pecher, I., Netzeband, G. L., Greinert, J., Mountjoy, J. J., Pedley, K., and Crutchley, G. (2010). Tectonic and geological framework for gas hydrates and cold seeps on the Hikurangi subduction margin, New Zealand. *Marine Geology*, 272:26–48.
- Bassett, D., Sutherland, R., Henrys, S., Stern, T., Scherwath, M., Benson, A., Toulmin, S., and Henderson, M. (2010). Three-dimensional velocity structure of the northern Hikurangi margin, Raukumara, New Zealand: Implications for the growth

- of continental crust by subduction erosion and tectonic underplating. *Geochemistry, Geophysics, Geosystems*, 11(10).
- Beanland, S. and Haines, J. (1998). The kinematics of active deformation in the North Island, New Zealand, determined from geological strain rates. *New Zealand Journal of Geology and Geophysics*, 41(4):311–323.
- Beeler, N. M., Lockner, D. L., and Hickman, S. H. (2001). A simple stick-slip and creep-slip model for repeating earthquakes and its implication for microearthquakes at Parkfield. *Bulletin of the Seismological Society of America*, 91(6):1797–1804.
- Beyreuther, M., Barsch, R., Krischer, L., Megies, T., Behr, Y., and Wassermann, J. (2010). ObsPy: A python toolbox for seismology. *Seismological Research Letters*, 81(3):530–533.
- Boese, C. M., Townend, J., Smith, E., and Stern, T. (2012). Microseismicity and stress in the vicinity of the Alpine Fault, central Southern Alps, New Zealand. *Journal of Geophysical Research: Solid Earth*.
- Bohnhoff, M., Wollin, C., Domigall, D., Küperkoch, L., Martínez-Garzón, P., Kwiatek, G., and Malin, P. E. (2017). Repeating Marmara Sea earthquakes: Indication for fault creep. *Geophysical Journal International*, 210(1):332–339.
- Bormann, P., Baumbach, M., Bock, G., Grosser, H., Choy, G. L., and Boatwright, J. (2009). Seismic sources and source parameters. *New Manual of Seismological Observatory Practice (NMSOP-1)*, pages 1–94.
- Brune, J. N. (1970). Tectonic stress and the spectra of seismic shear waves from earthquakes. *Journal of Geophysical Research*, 75(26):4997–5009.
- Bürgmann, R., Kogan, M. G., Steblov, G. M., Hilley, G., Levin, V. E., and Apel, E. (2005). Interseismic coupling and asperity distribution along the Kamchatka subduction zone. *Journal of Geophysical Research: Solid Earth*, 110(7):1–17.
- Chamberlain, C. J., Hopp, C. J., Boese, C. M., Warren-Smith, E., Chambers, D., Chu, S. X., Michailos, K., and Townend, J. (2018). EQcorrscan: Repeating and near-repeating earthquake detection and analysis in python. *Seismological Research Letters*, 89(1):173–181.
- Chamberlain, C. J., Townend, J., and Thomas, A. (2019). A repeating earthquake catalog for New Zealand. In *SSA Annual Meeting Abstracts*.
- Chen, K. H., Bürgmann, R., and Nadeau, R. M. (2013). Do earthquakes talk to each other? Triggering and interaction of repeating sequences at Parkfield. *Journal of Geophysical Research: Solid Earth*, 118(1):165–182.
- Chen, K. H., Bürgmann, R., Nadeau, R. M., Chen, T., and Lapusta, N. (2010). Post-seismic variations in seismic moment and recurrence interval of repeating earthquakes. *Earth and Planetary Science Letters*, 299(1-2):118–125.

- Chen, K. H., Nadeau, R. M., and Rau, R. J. (2007). Towards a universal rule on the recurrence interval scaling of repeating earthquakes? *Geophysical Research Letters*, 34(16):1–5.
- Chen, K. H., Nadeau, R. M., and Rau, R. J. (2008). Characteristic repeating earthquakes in an arc-continent collision boundary zone: The Chihshang fault of eastern Taiwan. *Earth and Planetary Science Letters*, 276(3-4):262–272.
- Chen, T. and Lapusta, N. (2009). Scaling of small repeating earthquakes explained by interaction of seismic and aseismic slip in a rate and state fault model. *Journal of Geophysical Research: Solid Earth*, 114(1):1–12.
- Clark, K., Howarth, J., Litchfield, N., Cochran, U., Turnbull, J., Dowling, L., Howell, A., Berryman, K., and Wolfe, F. (2019). Geological evidence for past large earthquakes and tsunamis along the Hikurangi subduction margin, New Zealand. *Marine Geology*, 412(October 2018):139–172.
- Clark, K. J., Hayward, B. W., Cochran, U. A., Wallace, L. M., Power, W. L., and Sabaa, A. T. (2015). Evidence for past subduction earthquakes at a plate boundary with widespread upper plate faulting: Southern Hikurangi margin, New Zealand. *Bulletin of the Seismological Society of America*.
- Davy, B. and Wood, R. (1994). Gravity and magnetic modelling of the Hikurangi Plateau. *Marine Geology*, 118(1-2):139–151.
- Deichmann, N. and Garcia-Fernandez, M. (1992). Rupture geometry from high-precision relative hypocentre locations of microearthquake clusters. *Geophysical Journal International*, 110:501–517.
- Delahaye, E. J., Townend, J., Reyners, M. E., and Rogers, G. (2009). Microseismicity but no tremor accompanying slow slip in the Hikurangi subduction zone, New Zealand. *Earth and Planetary Science Letters*.
- DeMets, C., Gordon, R. G., and Argus, D. F. (2010). Geologically current plate motions. *Geophysical Journal International*, 181:1–80.
- Dominguez, L. A., Taira, T., and Santoyo, M. A. (2016). Spatiotemporal variations of characteristic repeating earthquake sequences along the Middle America Trench in Mexico. *Journal of Geophysical Research: Solid Earth*, 121(12):8855–8870.
- Doser, D. I. and Webb, T. H. (2003). Source parameters of large historical (1917–1961) earthquakes, North Island, New Zealand. *Geophysical Journal International*, 152:795–832.
- Eberhart-Phillips, D. and Bannister, S. (2015). 3-D imaging of the northern Hikurangi subduction zone, New Zealand: variations in subducted sediment, slab fluids and slow slip. *Geophysical Journal International*, 201(2):838–855.
- Eberhart-Phillips, D., Bannister, S., and Reyners, M. (2017). New Zealand Wide Model 2.1 Seismic Velocity Model For New Zealand (Version 2.1) [Data set].

- Eberhart-Phillips, D., Reyners, M., Faccenda, M., and Naliboff, J. (2013). Along-strike variation in subducting plate seismicity and mantle wedge attenuation related to fluid release beneath the North Island, New Zealand. *Physics of the Earth and Planetary Interiors*, 225:12–27.
- Ellsworth, W. L. (1995). Characteristic earthquakes and long-term earthquake forecasts: Implications of central California seismicity. *Urban Disaster Mitigation: The Role of Engineering and Technology*, pages 1–14.
- Frohlich, C. (1992). Triangle diagrams: ternary graphs to display similarity and diversity of earthquake focal mechanisms. *Physics of the Earth and Planetary Interiors*, 75(1-3):193–198.
- Frohlich, C. and Apperson, K. D. (1992). Earthquake focal mechanisms, moment tensors, and the consistency of seismic activity near plate boundaries. *Tectonics*, 11(2):279–296.
- Fry, B., Chao, K., Bannister, S., Peng, Z., and Wallace, L. (2011). Deep tremor in New Zealand triggered by the 2010 Mw8.8 Chile earthquake. *Geophysical Research Letters*, 38(15):1–5.
- Gao, D. and Kao, H. (2020). Optimization of the match-filtering method for robust repeating earthquake detection: The multisegment cross-correlation approach. *Journal of Geophysical Research: Solid Earth*, 125(7):1–19.
- Gardonio, B., Marsan, D., Socquet, A., Bouchon, M., Jara, J., Sun, Q., Cotte, N., and Campillo, M. (2018). Revisiting slow slip events occurrence in Boso Peninsula, Japan, combining GPS data and repeating earthquakes analysis. *Journal of Geophysical Research: Solid Earth*, 123(2):1502–1515.
- Geller, R. J. and Mueller, C. S. (1980). Four similar earthquakes in central California. *Geophysical Research Letters*, 7(10):821–824.
- Gibowicz, S. J. and Lasocki, S. (2007). Analysis of shallow and deep earthquake doublets in the Fiji-Tonga-Kermadec region. *Pure and Applied Geophysics*, 164(1):53–74.
- Gutenberg, B. and Richter, C. F. (1942). Earthquake magnitude, intensity, energy, and acceleration. *Bulletin of the Seismological Society of America*, 32(3):163–191.
- Hanks, T. C. and Kanamori, H. (1979). A moment magnitude scale. *Journal of Geophysical Research*, 84(B5):2348–2350.
- Hatakeyama, N., Uchida, N., Matsuzawa, T., and Nakamura, W. (2017). Emergence and disappearance of interplate repeating earthquakes following the 2011 M9.0 Tohoku-oki earthquake: Slip behavior transition between seismic and aseismic depending on the loading rate. *Journal of Geophysical Research: Solid Earth*, 122(7):5160–5180.
- Havskov, J. and Ottemöller, L. (1999). SeisAn earthquake analysis software. *Seismological Research Letters*, 70(5):532–534.

- Havskov, J. and Ottemoller, L. (2010). *Routine Data Processing in Earthquake Seismology*. Springer.
- Hayward, T. W. and Bostock, M. G. (2017). Slip behavior of the Queen Charlotte Plate boundary before and after the 2012, MW 7.8 Haida Gwaii earthquake: Evidence from repeating earthquakes. *Journal of Geophysical Research: Solid Earth*, 122(11):8990–9011.
- Hughes, L., Chamberlain, C. J., Townend, J., and Amanda, M. (2021). A Repeating Earthquake Catalog from 2003 to 2018 for the Raukumara Peninsula , Northern Hikurangi Subduction Margin , New Zealand. *Geochemistry, Geophysics, Geosystems*, pages in–press.
- Hull, A. G. (1990). Tectonics of the 1931 hawke’s bay earthquake. *New Zealand Journal of Geology and Geophysics*, 33(2):309–320.
- Ide, S. (2012). Variety and spatial heterogeneity of tectonic tremor worldwide. *Journal of Geophysical Research: Solid Earth*, 117(B03302).
- Igarashi, T. (2010). Spatial changes of inter-plate coupling inferred from sequences of small repeating earthquakes in Japan. *Geophysical Research Letters*, 37(20):1–5.
- Igarashi, T., Matsuzawa, T., and Hasegawa, A. (2003). Repeating earthquakes and interplate aseismic slip in the northeastern Japan subduction zone. *Journal of Geophysical Research: Solid Earth*, 108(B5):1–9.
- Ikari, M. J., Wallace, L. M., Rabinowitz, H. S., Savage, H. M., Hamling, I. J., and Kopf, A. J. (2020). Observations of laboratory and natural slow slip Events: Hikurangi subduction zone, New Zealand. *Geochemistry, Geophysics, Geosystems*, 21(2):1–19.
- Jacobs, K. M., Savage, M. K., and Smith, E. C. (2016). Quantifying seismicity associated with slow slip events in the Hikurangi margin, New Zealand. *New Zealand Journal of Geology and Geophysics*, 59(1):58–69.
- Johnson, L. R. (2010). An earthquake model with interacting asperities. *Geophysical Journal International*, 182(3):1339–1373.
- Kamp, P. J. (1999). Tracking crustal processes by FT thermochronology in a forearc high (Hikurangi margin, New Zealand) involving Cretaceous subduction termination and mid-Cenozoic subduction initiation. *Tectonophysics*, 307(3-4):313–343.
- Kaneko, Y., Avouac, J. P., and Lapusta, N. (2010). Towards inferring earthquake patterns from geodetic observations of interseismic coupling. *Nature Geoscience*, 3(5):363–369.
- Kato, A., Igarashi, T., and Obara, K. (2014). Detection of a hidden Boso slow slip event immediately after the 2011 Mw 9.0 Tohoku-Oki earthquake, Japan. *Geophysical Research Letters*, 41(16):5868–5874.

- Kato, A., Obara, K., Igarashi, T., Tsuruoka, H., Nakagawa, S., and Hirata, N. (2012). Propagation of slow slip leading up to the 2011Mw 9.0 Tohoku-Oki earthquake. *Science*, 335:705–708.
- Kimura, H., Kasahara, K., Igarashi, T., and Hirata, N. (2006). Repeating earthquake activities associated with the Philippine Sea plate subduction in the Kanto district, central Japan: A new plate configuration revealed by interplate aseismic slips. *Tectonophysics*, 417(1-2):101–118.
- Lamb, S. (2011). Cenozoic tectonic evolution of the New Zealand plate-boundary zone: A paleomagnetic perspective. *Tectonophysics*, 509(3-4):135–164.
- Langridge, R. M., Ries, W. F., Litchfield, N. J., Villamor, P., Van Dissen, R. J., Barrell, D. J., Rattenbury, M. S., Heron, D. W., Haubrock, S., Townsend, D. B., Lee, J. M., Berryman, K. R., Nicol, A., Cox, S. C., and Stirling, M. W. (2016). The New Zealand Active Faults Database. *New Zealand Journal of Geology and Geophysics*, 59(1):86–96.
- Lay, T. and Wallace, T. (1995). *Modern Global Seismology*. Academic Press.
- Lengliné, O. and Marsan, D. (2009). Inferring the coseismic and postseismic stress changes caused by the 2004 Mw = 6 Parkfield earthquake from variations of recurrence times of microearthquakes. *Journal of Geophysical Research: Solid Earth*, 114(10):1–19.
- Li, L., Chen, Q. F., Cheng, X., and Niu, F. (2007). Spatial clustering and repeating of seismic events observed along the 1976 Tangshan fault, north China. *Geophysical Research Letters*, 34(23):4–9.
- Li, L., Chen, Q. F., Niu, F., and Su, J. (2011). Deep slip rates along the Longmen Shan fault zone estimated from repeating microearthquakes. *Journal of Geophysical Research: Solid Earth*, 116(9):1–14.
- Lomax, A., Virieux, J., Volant, P., and Berge-Thierry, C. (2000). Probabilistic earthquake location in 3D and layered models - Introduction of a Metropolis-Gibbs method and comparison with linear locations. In *Advances in Seismic Event Location*.
- Loveless, J. P. and Meade, B. J. (2011). Spatial correlation of interseismic coupling and coseismic rupture extent of the 2011 MW = 9.0 Tohoku-oki earthquake. *Geophysical Research Letters*, 38(17):2–6.
- Marone, C., Vidale, J. E., and Ellsworth, W. L. (1995). Fault healing inferred from time dependent variations in source properties of repeating earthquakes. *Geophysical Research Letters*, 22(22):3095–3098.
- Martin, J. and Crowley, J. L. (1995). Comparison of Correlation Techniques. *Intelligent Autonomous Systems: IAS-4 : Proceedings of the International Conference*, pages 86–93.

- Materna, K., Taira, T., and Bürgmann, R. (2018). Aseismic transform fault slip at the Mendocino triple junction from characteristically repeating earthquakes. *Geophysical Research Letters*, 45(2):699–707.
- Mavrommatis, A. P., Segall, P., Uchida, N., and Johnson, K. M. (2015). Long-term acceleration of aseismic slip preceding the Mw 9 Tohoku-oki earthquake: Constraints from repeating earthquakes. *Geophysical Research Letters*, 42(22):9717–9725.
- McKenzie, D. P. (1969). The relationship between fault plane solutions for earthquakes and the directions of the principal stresses. *Bulletin of the Seismological Society of America*, 59(2):591–601.
- Mesimeri, M. and Karakostas, V. (2018). Repeating earthquakes in western Corinth Gulf (Greece): Implications for aseismic slip near locked faults. *Geophysical Journal International*, 215(1):659–676.
- Mochizuki, K., Sutherland, R., Henrys, S., Bassett, D., Van Avendonk, H., Arai, R., Kodaira, S., Fujie, G., Yamamoto, Y., Bangs, N., and Barker, D. (2019). Recycling of depleted continental mantle by subduction and plumes at the Hikurangi Plateau large igneous province, southwestern Pacific Ocean. *Geology*, 47:795–798.
- Mortimer, N., Campbell, H. J., Tulloch, A. J., King, P. R., Stagpoole, V. M., Wood, R. A., Rattenbury, M. S., Sutherland, R., Adams, C. J., Collot, J., and Seton, M. (2017). Zealandia: Earth’s hidden Continent. *GSA Today*, 27(3-4):27–35.
- Mortimer, N. and Parkinson, D. (1996). Hikurangi Plateau: A Cretaceous large igneous province in the southwest Pacific Ocean. *Journal of Geophysical Research: Solid Earth*, 101(B1):687–696.
- Müllner, D. (2011). Modern hierarchical, agglomerative clustering algorithms. *arXiv preprint arXiv:1109.2378*, pages 1–29.
- Nadeau, R., Antolik, M., Johnson, P. A., Foxall, W., and McEvilly, T. V. (1994). Seismological studies at Parkfield III: microearthquake clusters in the study of fault-zone dynamics. *Bulletin of the Seismological Society of America*, 84(2):247–263.
- Nadeau, R. M., Foxall, W., and McEvilly, T. V. (1995). Clustering and periodic recurrence of microearthquakes on the San Andreas fault at Parkfield, California. *Science*, 267(5197):503–507.
- Nadeau, R. M. and Johnson, L. R. (1998). Seismological studies at Parkfield VI: moment release rates and estimates of source parameters for small repeating earthquakes. *Bulletin of the Seismological Society of America*, 88(3):790–814.
- Nadeau, R. M. and McEvilly, T. V. (1999). Fault slip rates at depth from recurrence intervals of repeating microearthquakes. *Science*, 285(5428):718–721.
- Nadeau, R. M. and McEvilly, T. V. (2004). Periodic pulsing of characteristic microearthquakes on the San Andreas Fault. *Science*, 303(5655):220–222.

- Nicol, A., Mazengarb, C., Chanier, F., Rait, G., Uruski, C., and Wallace, L. (2007). Tectonic evolution of the active Hikurangi subduction margin, New Zealand, since the Oligocene. *Tectonics*, 26(4):1–24.
- Nicol, A. and Wallace, L. M. (2007). Temporal stability of deformation rates: Comparison of geological and geodetic observations, Hikurangi subduction margin, New Zealand. *Earth and Planetary Science Letters*, 258(3-4):397–413.
- Nomura, S., Ogata, Y., and Nadeau, R. M. (2014). Space-time model for repeating earthquakes and analysis of recurrence intervals on the San Andreas Fault near Parkfield, California. *Journal of Geophysical Research: Solid Earth*, 119.
- Nomura, S., Ogata, Y., Uchida, N., and Matsu’ura, M. (2017). Spatiotemporal variations of interplate slip rates in northeast Japan inverted from recurrence intervals of repeating earthquakes. *Geophysical Journal International*, 208:468–481.
- Peng, Z., Vidale, J. E., Marone, C., and Rubin, A. (2005). Systematic variations in recurrence interval and moment of repeating aftershocks. *Geophysical Research Letters*, 32(15):3–6.
- Petersen, T., Gledhill, K., Chadwick, M., Gale, N. H., and Ristau, J. (2011). The New Zealand national seismograph network. *Seismological Research Letters*, 82(1):9–20.
- Poupinet, G., Ellsworth, W. L., and Frechet, J. (1984). Monitoring velocity variations in the crust using earthquake doublets: An application to the Calaveras Fault, California. *Journal of Geophysical Research*, 89(4):5719–5731.
- Proakis, J. and Dimitris, M. (1990). *Digital Signal Processing: principles, devices and applications*. Prentice-Hall.
- Rao, K. D. and Swamy, M. N. (2018). *Digital signal processing: Theory and practice*. Springer.
- Rau, R. J., Chen, K. H., and Ching, K. E. (2007). Repeating earthquakes and seismic potential along the northern Longitudinal Valley fault of eastern Taiwan. *Geophysical Research Letters*, 34(24):1–6.
- Reid, H. (1910). *The California Earthquake of April 18, 1906, Report of the State Earthquake Investigation Commission*. Carnegie Institution of Washington.
- Rhoades, D. A., Christophersen, A., Bourguignon, S., Ristau, J., and Salichon, J. (2020a). A depth-dependent local magnitude scale for New Zealand earthquakes consistent with moment magnitude. *Bulletin of the Seismological Society of America*, 111(2):1–11.
- Rhoades, D. A., Christophersen, A., Bourguignon, S., Ristau, J., and Salichon, J. (2020b). A Depth-Dependent Local Magnitude Scale for New Zealand Earthquakes Consistent with Moment Magnitude. *Bulletin of the Seismological Society of America*, 111(2).

- Richter, C. F. (1935). An instrumental earthquake magnitude scale. *Bulletin of the Seismological Society of America*, 25(1):1–32.
- Ristau, J. (2013). Update of regional moment tensor analysis for earthquakes in New Zealand and adjacent offshore regions. *Bulletin of the Seismological Society of America*, 103(4):2520–2533.
- Ristau, J., Harte, D., and Salichon, J. (2016). A Revised Local Magnitude (M_L) Scale for New Zealand Earthquakes. *Bulletin of the Seismological Society of America*, 106(2):398–407.
- Robinson, R. (1994). Shallow subduction tectonics and fault interaction: The Weber, New Zealand, earthquake sequence of 1990-1992. *Journal of Geophysical Research*, 99(B5):9663–9679.
- Romanet, P. and Ide, S. (2019). Ambient tectonic tremors in Manawatu, Cape Turnagain, Marlborough, and Puysegur, New Zealand. *Earth, Planets and Space*, 71(59).
- Rubin, A. M. (2002). Using repeating earthquakes to correct high-precision earthquake catalogs for time-dependent station delays. *Bulletin of the Seismological Society of America*, 92(5):1647–1659.
- Rubin, A. M., Gillard, D., and Got, J. L. (1999). Streaks of microearthquakes along creeping faults. *Nature*, 400(6745):635–641.
- Rubinstein, J. L. and Ellsworth, W. L. (2010). Precise estimation of repeating earthquake moment: Example from parkfield, california. *Bulletin of the Seismological Society of America*, 100(5A):1952–1961.
- Rubinstein, J. L., Ellsworth, W. L., Chen, K. H., and Uchida, N. (2012). Fixed recurrence and slip models better predict earthquake behavior than the time-and slip-predictable models: 1. Repeating earthquakes. *Journal of Geophysical Research: Solid Earth*, 117(2):1–23.
- Schaff, D. P. and Beroza, G. C. (2004). Coseismic and postseismic velocity changes measured by repeating earthquakes. *Journal of Geophysical Research: Solid Earth*, 109(10).
- Schaff, D. P., Beroza, G. C., and Shaw, B. E. (1998). Postseismic response of repeating aftershocks. *Geophysical Research Letters*, 25(24):4549–4552.
- Schaff, D. P., Bokelmann, G. H. R., Beroza, G. C., Waldhauser, F., and Ellsworth, W. L. (2002). High-resolution image of Calaveras Fault seismicity. *Journal of Geophysical Research: Solid Earth*, 107(B9):5–1.
- Schaff, D. P. and Richards, P. G. (2011). On finding and using repeating seismic events in and near China. *Journal of Geophysical Research: Solid Earth*, 116(3):1–20.

- Schaff, D. P. and Richards, P. G. (2014). Improvements in magnitude precision, using the statistics of relative amplitudes measured by cross correlation. *Geophysical Journal International*, 197(1):335–350.
- Schmittbuhl, J., Karabulut, H., Lengliné, O., and Bouchon, M. (2016). Long-lasting seismic repeaters in the Central Basin of the Main Marmara Fault. *Geophysical Research Letters*, 43(18):9527–9534.
- Schwartz, D. P. and Coppersmith, K. J. (1984). Fault behavior and characteristic earthquakes: examples from the Wasatch and San Andreas fault zones. *Journal of Geophysical Research*, 89(B7):5681–5698.
- Shaddox, H. R. and Schwartz, S. Y. (2019). Subducted seamount diverts shallow slow slip to the forearc of the northern Hikurangi subduction zone, New Zealand. *Geology*, 47(5):415–418.
- Shimazaki, K. and Nakata, T. (1980). Time-Predictable recurrence model for large earthquakes. *Geophysical Research Letters*, 4(4):279–282.
- Stein, S. and Wysession, M. (2003). *An Introduction to Seismology, Earthquakes, and Earth Structure*. John Wiley & Sons.
- Stirling, M., McVerry, G., Gerstenberger, M., Litchfield, N., Van Dissen, R., Berryman, K., Barnes, P., Wallace, L., Villamor, P., Langridge, R., Lamarche, G., Nodder, S., Reyners, M., Bradley, B., Rhoades, D., Smith, W., Nicol, A., Pettinga, J., Clark, K., and Jacobs, K. (2012). National seismic hazard model for New Zealand: 2010 update. *Bulletin of the Seismological Society of America*, 102(4):1514–1542.
- Templeton, D. C., Nadeau, R. M., and Bürgmann, R. (2008). Behavior of repeating earthquake sequences in Central California and the implications for subsurface fault creep. *Bulletin of the Seismological Society of America*, 98(1):52–65.
- Trugman, D. T. and Shearer, P. M. (2017). GrowClust: A Hierarchical clustering algorithm for relative earthquake relocation, with application to the Spanish Springs and Sheldon, Nevada, earthquake sequences. *Seismological Research Letters*, 88(2):379–391.
- Turner, R. C., Nadeau, R. M., and Bürgmann, R. (2013). Aseismic slip and fault interaction from repeating earthquakes in the Loma Prieta aftershock zone. *Geophysical Research Letters*, 40(6):1079–1083.
- Uchida, N. (2015). Postseismic response of repeating earthquakes around the 2011 Tohoku-oki earthquake: Moment increases due to the fast loading rate. *Journal of Geophysical Research: Solid Earth*, pages 3782–3803.
- Uchida, N. (2016). Periodic slow slip triggers megathrust zone earthquakes in north-eastern Japan. *Science*, 351(6272):488–492.
- Uchida, N. (2019). Detection of repeating earthquakes and their application in characterizing slow fault slip. *Progress in Earth and Planetary Science*, 6(1).

- Uchida, N. and Bürgmann, R. (2019). Repeating Earthquakes. *Annual Review of Earth and Planetary Sciences*, 47:305–332.
- Uchida, N., Kalafat, D., Pinar, A., and Yamamoto, Y. (2019). Repeating earthquakes and interplate coupling along the western part of the North Anatolian Fault. *Tectonophysics*, 769(July):228185.
- Uchida, N. and Matsuzawa, T. (2011). Coupling coefficient, hierarchical structure, and earthquake cycle for the source area of the 2011 off the Pacific coast of Tohoku earthquake inferred from small repeating earthquake data. *Earth, Planets and Space*, 63(7):675–679.
- Uchida, N., Matsuzawa, T., Ellsworth, W. L., Imanishi, K., Okada, T., and Hasegawa, A. (2007). Source parameters of a M4.8 and its accompanying repeating earthquakes off Kamaishi, NE Japan: Implications for the hierarchical structure of asperities and earthquake cycle. *Geophysical Research Letters*, 34(20):3–7.
- Uchida, N., Matsuzawa, T., Ellsworth, W. L., Imanishi, K., Shimamura, K., and Hasegawa, A. (2012). Source parameters of microearthquakes on an interplate asperity off Kamaishi, NE Japan over two earthquake cycles. *Geophysical Journal International*, 189(2):999–1014.
- Uchida, N., Matsuzawa, T., Hasegawa, A., and Igarashi, T. (2003). Interplate quasi-static slip off Sanriku, NE Japan, estimated from repeating earthquakes. *Geophysical Research Letters*, 30(15):1–4.
- Uchida, N., Matsuzawa, T., Hirahara, S., and Hasegawa, A. (2006). Small repeating earthquakes and interplate creep around the 2005 Miyagi-oki earthquake (M=7.2). *Earth, Planets and Space*, 58(12):1577–1580.
- Uchida, N., Nakajima, J., Hasegawa, A., and Matsuzawa, T. (2009). What controls interplate coupling?: Evidence for abrupt change in coupling across a border between two overlying plates in the NE Japan subduction zone. *Earth and Planetary Science Letters*, 283(1-4):111–121.
- Van Dissen, R., Seebeck, H., Litchfield, N., Barnes, P., Nicol, A., Langridge, R., Barrell, D. J., Villamor, P., Ellis, S., Rattenbury, M., Bannister, S., Gerstenberger, M., Ghisetti, F., Sutherland, R., Fraser, S., Nodder, S., Stirling, M., Humphrey, J., Bland, K. J., Howell, A., Mountjoy, J., Moon, V., Stahl, T., Spinardi, F., Townsend, D., Clark, K., Hamling, I., Cox, S., W, d. L., Wopereis, P., Johnston, M. R., Morgenstern, R., Eccles, J. D., Little, T., Fry, B., Griffin, J., Townend, J., Mortimer, N., Alcaraz, S., Massiot, C., Rowland, J., Muirhead, J., Upton, P., Hirschberg, H., and Lee, J. (2021). Development of the New Zealand Community Fault Model – version 1.0. In *NZSEE*.
- Vidale, J. E., Ellsworth, W. L., Cole, A., and Marone, C. (1994). Variations in rupture process with recurrence interval in a repeated small earthquake. *Nature*, 368(April):624–626.

- Virtanen, P., Gommers, R., Oliphant, T. E., Haberland, M., Reddy, T., Cournapeau, D., Burovski, E., Peterson, P., Weckesser, W., Bright, J., van der Walt, S. J., Brett, M., Wilson, J., Millman, K. J., Mayorov, N., Nelson, A. R., Jones, E., Kern, R., Larson, E., Carey, C. J., Polat, , Feng, Y., Moore, E. W., VanderPlas, J., Laxalde, D., Perktold, J., Cimrman, R., Henriksen, I., Quintero, E. A., Harris, C. R., Archibald, A. M., Ribeiro, A. H., Pedregosa, F., van Mulbregt, P., Vijaykumar, A., Bardelli, A. P., Rothberg, A., Hilboll, A., Kloeckner, A., Scopatz, A., Lee, A., Rokem, A., Woods, C. N., Fulton, C., Masson, C., Häggström, C., Fitzgerald, C., Nicholson, D. A., Hagen, D. R., Pasechnik, D. V., Olivetti, E., Martin, E., Wieser, E., Silva, F., Lenders, F., Wilhelm, F., Young, G., Price, G. A., Ingold, G. L., Allen, G. E., Lee, G. R., Audren, H., Probst, I., Dietrich, J. P., Silterra, J., Webber, J. T., Slavič, J., Nothman, J., Buchner, J., Kulick, J., Schönberger, J. L., de Miranda Cardoso, J. V., Reimer, J., Harrington, J., Rodríguez, J. L. C., Nunez-Iglesias, J., Kuczynski, J., Tritz, K., Thoma, M., Newville, M., Kümmerer, M., Bolingbroke, M., Tartre, M., Pak, M., Smith, N. J., Nowaczyk, N., Shebanov, N., Pavlyk, O., Brodtkorb, P. A., Lee, P., McGibbon, R. T., Feldbauer, R., Lewis, S., Tygier, S., Sievert, S., Vigna, S., Peterson, S., More, S., Pudlik, T., Oshima, T., Pingel, T. J., Robitaille, T. P., Spura, T., Jones, T. R., Cera, T., Leslie, T., Zito, T., Krauss, T., Upadhyay, U., Halchenko, Y. O., and Vázquez-Baeza, Y. (2020). SciPy 1.0: fundamental algorithms for scientific computing in Python. *Nature Methods*, 17(3):261–272.
- Waldhauser, F. and Ellsworth, W. L. (2000). A Double-difference Earthquake location algorithm: Method and application to the Northern Hayward Fault, California. *Bulletin of the Seismological Society of America*, 90(6):1353–1368.
- Waldhauser, F., Ellsworth, W. L., Schaff, D. P., and Cole, A. (2004). Streaks, multiplets, and holes: High-resolution spatio-temporal behavior of Parkfield seismicity. *Geophysical Research Letters*, 31(18):2–5.
- Wallace, L. M. (2020). Slow Slip Events in New Zealand. *Annual Review of Earth and Planetary Sciences*, 48:1–8.
- Wallace, L. M., Barnes, P., Beavan, J., Van Dissen, R., Litchfield, N., Mountjoy, J., Langridge, R., Lamarche, G., and Pondard, N. (2012a). The kinematics of a transition from subduction to strike-slip: An example from the central New Zealand plate boundary. *Journal of Geophysical Research: Solid Earth*, 117(B02405).
- Wallace, L. M., Bartlow, N., Hamling, I., and Fry, B. (2014). Quake clamps down on slow slip. *Geophysical Research Letters*, 41(24):8840–8846.
- Wallace, L. M. and Beavan, J. (2010). Diverse slow slip behavior at the Hikurangi subduction margin, New Zealand. *Journal of Geophysical Research: Solid Earth*, 115(B12402).
- Wallace, L. M., Beavan, J., Bannister, S., and Williams, C. (2012b). Simultaneous long-term and short-term slow slip events at the Hikurangi subduction margin, New Zealand: Implications for processes that control slow slip event occurrence, duration, and migration. *Journal of Geophysical Research B: Solid Earth*, 117(B11402).

- Wallace, L. M., Beavan, J., McCaffrey, R., and Darby, D. (2004). Subduction zone coupling and tectonic block rotations in the North Island, New Zealand. *Journal of Geophysical Research: Solid Earth*, 109(B12406):1–21.
- Wallace, L. M. and Eberhart-Phillips, D. (2013). Newly observed, deep slow slip events at the central Hikurangi margin, New Zealand: Implications for downdip variability of slow slip and tremor, and relationship to seismic structure. *Geophysical Research Letters*, 40(20):5393–5398.
- Wallace, L. M., Hreinsdóttir, S., Ellis, S., Hamling, I., D’Anastasio, E., and Denys, P. (2018). Triggered slow slip and afterslip on the southern Hikurangi subduction zone following the Kaikōura earthquake. *Geophysical Research Letters*, 45(10):4710–4718.
- Wallace, L. M., Kaneko, Y., Hreinsdóttir, S., Hamling, I., Peng, Z., Bartlow, N., D’Anastasio, E., and Fry, B. (2017). Large-scale dynamic triggering of shallow slow slip enhanced by overlying sedimentary wedge. *Nature Geoscience*, 10(10):765–770.
- Wallace, L. M., Reyners, M., Cochran, U., Bannister, S., Barnes, P. M., Berryman, K., Downes, G., Eberhart-Phillips, D., Fagereng, A., Ellis, S., Nicol, A., McCaffrey, R., Beavan, R. J., Henrys, S., Sutherland, R., Barker, D. H., Litchfield, N., Townend, J., Robinson, R., Bell, R., Wilson, K., and Power, W. (2009). Characterizing the seismogenic zone of a major plate boundary subduction thrust: Hikurangi Margin, New Zealand. *Geochemistry, Geophysics, Geosystems*, 10(10).
- Walsh, D., Arnold, R., and Townend, J. (2009). A Bayesian approach to determining and parametrizing earthquake focal mechanisms. *Geophysical Journal International*, 176:235–255.
- Warren-Smith, E., Fry, B., Kaneko, Y., and Chamberlain, C. J. (2018). Foreshocks and delayed triggering of the 2016 MW7.1 Te Araroa earthquake and dynamic reinvigoration of its aftershock sequence by the MW7.8 Kaikōura earthquake, New Zealand. *Earth and Planetary Science Letters*, 482:256–276.
- Webb, T. H. and Anderson, H. (1998). Focal mechanisms of large earthquakes in the North Island of New Zealand: slip partitioning at an oblique active margin. *Geophysical Journal International*, 134(1):40–86.
- Wiemer, S. and Wyss, M. (2000). Minimum magnitude of completeness in earthquake catalogs: Examples from Alaska, the Western United States, and Japan. *Bulletin of the Seismological Society of America*, 90(4):859–869.
- Williams, C. A., Eberhart-Phillips, D., Bannister, S., Barker, D. H., Henrys, S., Reyners, M., and Sutherland, R. (2013). Revised Interface Geometry for the Hikurangi Subduction Zone, New Zealand. *Seismological Research Letters*, 84(6):1066–1073.
- Yamashita, Y., Shimizu, H., and Goto, K. (2012). Small repeating earthquake activity, interplate quasi-static Slip, and interplate coupling in the Hyuga-nada, southwestern Japan subduction zone. *Geophysical Research Letters*, 39(8):1–5.

- Yao, D., Peng, Z., Kaneko, Y., Fry, B., and Meng, X. (2021). Dynamic triggering of earthquakes in the North Island of New Zealand following the 2016 Mw 7.8 Kaikōura earthquake. *Earth and Planetary Science Letters*, 557:116723.
- Yao, D., Walter, J. I., Meng, X., Hobbs, T. E., Peng, Z., Newman, A. V., Schwartz, S. Y., and Protti, M. (2017). Detailed spatiotemporal evolution of microseismicity and repeating earthquakes following the 2012 Mw 7.6 Nicoya earthquake. *Journal of Geophysical Research: Solid Earth*, 122(1):524–542.
- Yu, W. C. (2013). Shallow-focus repeating earthquakes in the tonga-kermadec-vanuatu subduction zones. *Bulletin of the Seismological Society of America*, 103(1):463–486.
- Yu, W. c. and Wen, L. (2012). Deep-focus repeating earthquakes in the Tonga-Fiji subduction zone. *Bulletin of the Seismological Society of America*, 102(4):1829–1849.
- Zechar, J. D. and Nadeau, R. M. (2012). Predictability of repeating earthquakes near Parkfield, California. *Geophysical Journal International*, 190(1):457–462.
- Zhang, J., Richards, P. G., and Schaff, D. P. (2008). Wide-scale detection of earthquake waveform doublets and further evidence for inner core super-rotation. *Geophysical Journal International*, 174(3):993–1006.

Appendices

A.1 Catalogue of repeating earthquakes near Porangahau

Table A.1: Summary of the repeating earthquake families identified near Porangahau in this study. The first two letters of the family ID correspond to the geographic group (Section 4.1). N is the number of events in each family and FM means focal mechanism. SS is for strike-slip faults, REV for reverse (thrust) faults and NOR for normal faults. Location is where the family was located, in the Australian Plate (AUS), in the Pacific Plate (PAC) or on the subduction interface (INTER).

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
AH1	-40.701	176.857	24.4	2.7	2	SS	PAC
AK1	-40.513	176.352	19.2	2.8	2	–	PAC
AK10	-40.542	176.424	18.5	1.9	2	–	PAC
AK11	-40.509	176.345	19.2	2.3	2	SS	PAC
AK12	-40.508	176.267	21.0	1.7	2	–	PAC
AK13	-40.605	176.533	18.4	2.3	2	REV	INTER
AK14	-40.564	176.460	15.3	1.3	2	–	PAC
AK15	-40.562	176.457	14.9	1.5	2	–	PAC
AK16	-40.492	176.461	21.7	1.6	2	–	PAC
AK17	-40.499	176.464	21.7	2.1	2	SS	PAC
AK18	-40.496	176.463	22.0	1.9	3	–	PAC
AK19	-40.695	176.646	14.9	2.3	2	–	PAC
AK2	-40.703	176.602	14.4	1.7	2	–	PAC
AK20	-40.581	176.475	15.8	1.9	2	–	PAC
AK3	-40.488	176.336	19.5	1.5	2	–	PAC
AK4	-40.625	176.583	14.3	1.5	2	–	PAC
AK5	-40.560	176.381	16.7	1.8	2	–	PAC
AK6	-40.640	176.563	14.5	1.8	2	–	PAC
AK7	-40.512	176.346	19.5	1.5	2	–	PAC
AK8	-40.511	176.349	19.3	2.7	2	SS	PAC
AK9	-40.571	176.459	15.8	2.2	2	–	PAC
CT1	-40.539	176.669	15.2	1.9	3	–	PAC
CT10	-40.587	176.696	7.7	2.1	2	–	AUS
CT11	-40.591	176.700	8.4	1.6	2	–	AUS
CT12	-40.596	176.706	8.4	2.8	2	REV	INTER
CT13	-40.590	176.697	8.6	2.4	2	SS	AUS
CT14	-40.480	176.616	14.6	2.0	2	–	PAC

Table A.1 continued from previous page

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
CT2	-40.539	176.687	15.1	1.6	2	–	PAC
CT3	-40.561	176.647	13.1	2.2	2	REV	INTER
CT4	-40.519	176.690	14.9	2.0	4	NOR	PAC
CT5	-40.564	176.646	13.9	2.4	3	REV	INTER
CT6	-40.532	176.681	15.0	1.5	3	NOR	PAC
CT7	-40.522	176.694	14.8	2.0	2	–	PAC
CT8	-40.517	176.618	15.1	1.9	2	–	PAC
CT9	-40.514	176.617	16.1	2.0	2	–	PAC
DV1	-40.632	175.871	29.1	2.3	2	–	PAC
DV10	-40.292	176.291	23.1	1.7	2	–	PAC
DV11	-40.385	175.790	31.3	1.6	2	–	PAC
DV12	-40.404	176.064	17.5	1.8	2	SS	AUS
DV13	-40.386	176.149	27.1	1.7	2	–	PAC
DV14	-40.512	175.872	30.2	2.1	2	–	PAC
DV15	-40.254	176.337	26.0	1.7	2	–	PAC
DV16	-40.249	176.225	26.1	1.7	2	–	PAC
DV2	-40.227	176.287	43.1	2.3	2	NOR	PAC
DV3	-40.439	175.955	30.9	2.7	2	NOR	PAC
DV4	-40.428	175.852	26.0	1.7	2	–	PAC
DV5	-40.454	176.023	25.0	2.5	3	REV	INTER
DV6	-40.209	176.186	24.8	1.6	2	–	PAC
DV7	-40.459	176.018	24.5	2.0	2	REV	INTER
DV8	-40.344	176.133	30.5	2.9	3	NOR	PAC
DV9	-40.457	176.020	24.8	2.4	4	REV	INTER
HE1	-39.634	176.763	28.3	2.1	2	–	PAC
HE10	-39.649	176.759	28.8	2.5	2	SS	PAC
HE11	-39.624	176.747	27.8	1.8	2	–	PAC
HE12	-39.627	176.747	27.7	2.2	2	–	PAC
HE13	-39.628	176.767	27.8	1.9	2	–	PAC
HE14	-39.620	176.749	28.4	1.8	2	REV	PAC
HE15	-39.625	176.757	27.4	2.5	2	SS	PAC
HE16	-39.657	176.747	27.4	1.8	2	–	PAC
HE2	-39.592	176.816	27.5	1.9	2	NOR	PAC
HE3	-39.630	176.755	27.5	2.4	2	SS	PAC
HE4	-39.625	176.765	28.2	1.6	2	NOR	PAC
HE5	-39.626	176.749	28.0	1.9	2	–	PAC

Table A.1 continued from previous page

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
HE6	-39.669	176.820	27.6	2.1	2	–	PAC
HE7	-39.663	176.748	26.9	2.2	2	–	PAC
HE8	-39.635	176.753	27.9	2.0	2	–	PAC
HE9	-39.629	176.770	28.0	1.8	2	–	PAC
HW1	-39.675	176.588	27.7	2.2	4	SS	PAC
HW2	-39.653	176.545	28.4	1.9	6	–	PAC
HW3	-39.653	176.546	28.7	1.9	4	–	PAC
HW4	-39.654	176.541	28.4	2.2	5	REV	INTER
HW5	-39.647	176.541	28.4	1.6	2	–	PAC
KK1	-40.100	176.872	16.1	1.6	6	–	PAC
KK10	-39.982	176.758	20.7	1.6	2	SS	PAC
KK11	-40.023	176.771	18.3	2.4	3	SS	PAC
KK12	-39.978	176.763	20.2	1.8	2	–	PAC
KK13	-40.017	176.811	16.6	1.9	3	–	PAC
KK14	-39.861	176.937	18.0	1.7	2	–	PAC
KK15	-39.866	176.866	16.1	1.7	2	–	PAC
KK16	-39.831	176.961	19.0	1.8	2	–	PAC
KK17	-39.660	177.022	21.8	1.8	2	–	PAC
KK18	-39.935	176.886	17.5	2.0	2	SS	PAC
KK19	-40.076	176.743	17.7	2.2	2	–	PAC
KK2	-40.023	176.772	18.0	2.1	4	SS	PAC
KK20	-40.038	177.126	11.3	2.2	2	–	AUS
KK21	-39.994	176.895	20.3	1.6	2	–	PAC
KK22	-39.779	176.932	19.0	2.1	2	–	PAC
KK23	-40.000	176.902	20.6	2.5	2	REV	PAC
KK24	-39.916	176.893	18.2	1.9	2	–	PAC
KK25	-39.916	176.888	18.4	1.9	3	–	PAC
KK26	-39.845	176.933	19.7	1.9	2	–	PAC
KK27	-39.874	176.961	16.5	2.2	2	–	PAC
KK28	-39.898	177.037	11.7	2.4	2	–	AUS
KK29	-39.895	177.037	11.4	1.8	2	–	AUS
KK3	-40.021	176.751	18.3	1.6	2	–	PAC
KK30	-40.142	176.923	11.3	2.0	2	–	AUS
KK31	-40.009	176.768	19.0	1.9	2	–	PAC
KK32	-40.022	176.851	17.9	2.5	2	–	PAC
KK33	-39.742	177.075	6.3	2.1	2	–	AUS

Table A.1 continued from previous page

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
KK4	-40.125	176.943	12.2	2.1	4	–	AUS
KK5	-39.727	177.146	11.5	2.0	2	–	AUS
KK6	-40.022	176.764	18.3	1.6	5	SS	PAC
KK7	-40.019	176.761	19.2	1.5	2	–	PAC
KK8	-40.047	177.141	11.0	2.0	3	–	AUS
KK9	-39.769	177.144	10.2	1.8	3	REV	INTER
MN1	-40.329	176.434	19.1	1.8	4	–	PAC
MN10	-40.362	176.369	26.4	2.3	2	NOR	PAC
MN11	-40.414	176.328	23.9	2.1	2	–	PAC
MN12	-40.285	176.495	24.4	3.1	2	NOR	PAC
MN13	-40.382	176.286	30.9	1.8	2	–	PAC
MN14	-40.340	176.312	16.6	2.0	2	SS	AUS
MN15	-40.406	176.294	23.7	1.4	2	–	PAC
MN16	-40.377	176.395	25.7	1.9	2	–	PAC
MN17	-40.376	176.392	25.9	1.8	2	–	PAC
MN18	-40.394	176.287	22.7	3.1	2	–	PAC
MN2	-40.389	176.362	20.4	2.1	2	SS	PAC
MN3	-40.342	176.429	20.8	1.9	2	–	PAC
MN4	-40.385	176.284	22.9	1.7	5	NOR	PAC
MN5	-40.330	176.434	19.2	1.6	4	–	PAC
MN6	-40.423	176.327	24.5	1.6	2	–	PAC
MN7	-40.320	176.419	25.9	1.8	2	–	PAC
MN8	-40.417	176.331	24.5	2.5	3	NOR	PAC
MN9	-40.340	176.439	26.1	2.2	2	–	PAC
MT1	-40.900	176.451	14.9	2.1	2	–	PAC
MT2	-40.647	176.158	18.7	1.4	2	–	PAC
MT3	-40.751	175.992	25.5	1.9	2	–	PAC
MT4	-40.648	176.154	18.5	1.4	2	–	PAC
MT5	-40.718	176.302	23.1	1.5	2	–	PAC
MT6	-40.695	176.338	23.6	1.8	2	NOR	PAC
OU1	-40.146	176.643	19.3	1.9	9	SS	PAC
OU10	-40.138	176.648	19.0	2.0	2	–	PAC
OU2	-40.158	176.631	19.0	1.6	3	–	PAC
OU3	-40.153	176.637	19.9	1.5	2	–	PAC
OU4	-40.135	176.647	20.0	2.0	3	–	PAC
OU5	-40.154	176.634	20.2	1.6	2	–	PAC

Table A.1 continued from previous page

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
OU6	-40.144	176.647	19.7	1.7	2	–	PAC
OU7	-40.145	176.629	19.9	1.8	2	–	PAC
OU8	-40.158	176.630	20.8	1.9	2	–	PAC
OU9	-40.140	176.644	19.6	1.7	3	–	PAC
PP1	-40.404	177.335	47.2	3.1	2	–	PAC
PP2	-40.408	177.349	43.1	3.7	2	–	PAC
PP3	-40.373	177.055	49.3	2.9	2	REV	PAC
RH1	-39.992	175.839	38.1	2.7	2	NOR	PAC
RH2	-39.949	176.189	30.7	1.9	2	NOR	PAC
TK1	-40.306	176.691	16.7	2.0	2	–	PAC
TK10	-40.307	176.687	16.4	1.4	2	–	PAC
TK11	-40.312	176.701	15.6	1.5	2	–	PAC
TK12	-40.299	176.675	13.9	1.8	3	–	PAC
TK13	-40.278	176.566	17.3	2.8	2	REV	INTER
TK14	-40.323	176.700	15.8	1.9	3	–	PAC
TK15	-40.322	176.700	16.4	2.7	2	–	PAC
TK16	-40.335	176.691	15.0	1.6	2	–	PAC
TK17	-40.355	176.622	16.7	1.8	6	–	PAC
TK18	-40.354	176.553	16.5	2.0	2	REV	INTER
TK19	-40.354	176.548	16.4	1.8	3	REV	INTER
TK2	-40.335	176.637	15.8	2.2	3	REV	INTER
TK20	-40.280	176.564	17.1	1.8	3	–	PAC
TK21	-40.353	176.677	15.7	1.7	2	–	PAC
TK22	-40.337	176.539	16.6	2.4	2	REV	INTER
TK23	-40.278	176.562	16.9	2.2	4	REV	INTER
TK24	-40.368	176.502	18.3	1.7	2	–	PAC
TK25	-40.306	176.593	15.6	2.8	3	SS	PAC
TK26	-40.306	176.753	15.0	2.1	2	–	PAC
TK27	-40.281	176.713	12.5	2.3	3	NOR	AUS
TK28	-40.301	176.743	13.1	1.9	2	–	PAC
TK29	-40.296	176.741	14.1	1.7	2	–	PAC
TK3	-40.325	176.518	18.3	1.7	2	SS	PAC
TK30	-40.283	176.741	12.5	2.4	2	REV	INTER
TK31	-40.326	176.519	17.6	1.8	3	–	PAC
TK32	-40.303	176.686	16.4	1.7	3	–	PAC
TK33	-40.396	176.625	15.8	2.2	2	REV	INTER

Table A.1 continued from previous page

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
TK34	-40.306	176.725	14.9	1.5	2	–	PAC
TK35	-40.276	176.560	16.8	2.0	3	SS	PAC
TK36	-40.311	176.699	16.2	2.3	2	–	PAC
TK37	-40.314	176.697	15.9	1.6	2	–	PAC
TK38	-40.324	176.604	17.9	1.9	2	–	PAC
TK39	-40.370	176.509	18.1	2.1	2	–	PAC
TK4	-40.304	176.747	14.6	1.7	2	–	PAC
TK40	-40.327	176.703	16.1	2.7	2	REV	INTER
TK41	-40.335	176.625	17.3	2.1	2	–	PAC
TK42	-40.375	176.522	17.9	1.9	2	REV	INTER
TK43	-40.395	176.627	15.5	2.6	2	REV	INTER
TK44	-40.360	176.497	18.7	3.0	2	REV	INTER
TK45	-40.303	176.682	16.3	1.7	2	–	PAC
TK46	-40.312	176.744	14.8	1.7	4	REV	INTER
TK47	-40.321	176.688	16.0	2.1	2	–	PAC
TK48	-40.285	176.707	13.0	1.6	2	–	AUS
TK49	-40.264	176.739	13.6	1.8	2	–	PAC
TK5	-40.323	176.770	14.1	2.7	2	REV	INTER
TK50	-40.301	176.727	14.2	1.6	2	–	PAC
TK51	-40.220	176.569	17.6	1.9	2	SS	PAC
TK52	-40.316	176.774	14.5	1.7	3	–	PAC
TK53	-40.308	176.767	15.0	2.1	2	REV	INTER
TK54	-40.305	176.682	16.4	2.2	2	NOR	PAC
TK55	-40.283	176.677	13.9	1.9	2	SS	PAC
TK56	-40.289	176.646	16.5	1.7	3	–	PAC
TK57	-40.295	176.637	16.4	1.4	2	–	PAC
TK58	-40.302	176.714	13.7	1.7	2	–	PAC
TK59	-40.362	176.544	15.8	2.0	3	–	PAC
TK6	-40.326	176.768	14.1	2.7	3	REV	INTER
TK60	-40.287	176.713	12.6	1.7	2	–	AUS
TK61	-40.271	176.630	15.2	1.4	2	–	PAC
TK62	-40.312	176.588	17.9	2.5	2	SS	PAC
TK7	-40.358	176.623	16.5	2.1	5	NOR	PAC
TK8	-40.313	176.737	14.4	1.6	2	NOR	PAC
TK9	-40.314	176.739	15.0	1.4	2	–	PAC
TR1	-40.361	176.851	12.5	2.1	4	–	PAC

Table A.1 continued from previous page

Family ID	Latitude (°)	Longitude (°)	Depth (km)	M_L	N	FM	Location
TR2	-40.305	176.874	13.5	1.7	2	–	PAC
TR3	-40.484	176.807	12.9	2.1	2	–	PAC
TT1	-39.706	176.861	38.3	2.4	2	NOR	PAC
TT2	-39.733	176.834	41.1	2.6	2	REV	PAC
WP1	-39.926	176.643	17.6	2.3	2	–	PAC
WP10	-39.922	176.623	23.6	1.9	2	–	PAC
WP11	-40.108	176.537	8.7	1.4	2	–	AUS
WP12	-39.986	176.637	22.8	1.6	2	–	PAC
WP13	-40.074	176.417	30.8	2.0	2	–	PAC
WP14	-40.076	176.442	29.2	2.5	2	REV	PAC
WP2	-40.002	176.509	23.8	1.7	2	–	PAC
WP3	-40.077	176.429	28.7	1.7	2	–	PAC
WP4	-40.044	176.447	29.5	1.5	2	–	PAC
WP5	-40.091	176.342	16.8	2.4	2	SS	AUS
WP6	-40.038	176.490	28.0	2.2	2	NOR	PAC
WP7	-40.097	176.395	30.5	1.5	2	–	PAC
WP8	-39.991	176.676	13.4	1.6	2	–	AUS
WP9	-40.099	176.552	18.3	1.6	2	–	PAC

A.2 Stations corrections from magnitude inversion

Table A.2: Summary of the stations corrections values obtained from the magnitude inversion.

Station and channel	Station correction	Station and channel	Station correction
ANWZ EHE	-1.900	MTW EHN	-1.133
ANWZ EHN	-1.809	NGZ EHE	-1.671
ARHZ EHE	-0.530	NGZ EHN	-1.507
ARHZ EHN	-0.536	NMHZ EHE	-1.581
BFZ HHE	-1.214	NMHZ EHN	-1.641
BFZ HHN	-1.259	OGWZ EHE	-1.186
BHHZ EHE	-0.139	OGWZ EHN	-1.114
BHHZ EHN	-0.132	PAWZ EHE	-0.601
BHW EHN	-1.238	PAWZ EHN	-0.783
BHW HHE	-0.814	PLWZ EHE	-1.208
BHW HHN	-1.152	PLWZ EHN	-0.942
BKZ HHE	-1.124	PNHZ EHE	-1.095
BKZ HHN	-1.163	PNHZ EHN	-0.705
CAW EHE	-0.594	POWZ EHE	-1.577
CAW EHN	-0.451	POWZ EHN	-1.473
CKHZ EHE	-0.789	PRHZ EHE	-1.469
CKHZ EHN	-0.981	PRHZ EHN	-1.363
CPWZ EHE	-0.559	PRWZ EHE	-1.696
CPWZ EHN	-0.525	PRWZ EHN	-1.393
DVHZ EHE	-1.041	PXZ HHE	-1.564
DVHZ EHN	-0.892	PXZ HHN	-1.660
FWVZ HHE	-0.913	RATZ HHE	-1.681
FWVZ HHN	-1.207	RATZ HHN	-1.327
HATZ EHE	-1.197	RITZ EHE	-0.658
HATZ EHN	-1.285	RITZ EHN	-0.864
HOWZ EHE	-1.135	TIWZ EHE	-1.066
HOWZ EHN	-1.323	TIWZ EHN	-1.117
KAHZ EHE	-1.217	TMWZ EHE	-1.705
KAHZ EHN	-0.983	TMWZ EHN	-1.195
KATZ EHE	-0.443	TRWZ EHE	-1.001
KATZ EHN	-1.194	TRWZ EHN	-0.831
KIW EHE	-1.300	TSZ HHE	-1.519
KIW EHN	-1.200	TSZ HHN	-1.498
KNZ HHE	-0.841	TUVZ EHE	-1.691
KNZ HHN	-0.476	TUVZ EHN	-1.604
KRHZ EHE	-1.424	TWVZ EHE	-0.875
KRHZ EHN	-1.352	TWVZ EHN	-1.124
KRVZ EHE	-1.294	VRZ HHN	-1.425
KRVZ EHN	-1.474	WAZ HHE	-1.747
MCHZ EHE	-1.128	WAZ HHN	-1.825
MCHZ EHN	-0.992	WEL HHE	-1.552
MOVZ EHE	-1.045	WEL HHN	-1.177
MOVZ EHN	-1.056	WNVZ EHE	-1.097
MRZ HHE	-1.445	WNVZ EHN	-1.047
MRZ HHN	-1.416	WPHZ EHE	-0.807
MSWZ EHE	-0.812	WPHZ EHN	-0.624
MSWZ EHN	-0.529	WPVZ HHN	-1.478
MTVZ EHE	-1.894	WTVZ EHE	-1.451
MTVZ EHN	-1.904	WTVZ EHN	-1.514
MTW EHE	-1.151		

A.3 Focal Mechanism Solutions

Table A.3: Table with the final focal mechanism solutions. ID is the GeoNet ID. The error calculated for each focal mechanism is in degrees.

ID	Strike	Dip	Rake	Error (°)
2863270	340.3	78.2	88.6	29.6
2961331	333.9	51.7	-154.6	29.2
2962789	120.9	81.2	114.0	31.4
2969615	161.0	77.8	-42.9	30.1
2982096	33.4	88.3	125.5	27.7
3016047	319.2	55.4	-169.7	28.7
3056579	118.2	51.4	-129.5	27.4
3056834	216.9	71.9	-153.6	27.1
3068492	298.5	46.0	-100.8	30.7
3066608	145.9	72.8	18.8	27.9
3118804	193.9	88.0	-144.6	26.6
3125550	86.4	37.2	46.1	29.4
3135048	121.9	87.2	-118.6	32.3
3140077	102.1	83.3	-59.3	28.7
3145133	107.3	47.8	-70.9	25.0
3151973	145.8	60.1	-43.3	29.5
3153127	111.9	45.4	-32.8	31.3
3157224	300.6	89.9	-31.9	29.5
3166416	214.8	61.6	90.7	27.4
3168746	247.6	80.7	-167.6	26.7
3180041	110.4	75.2	-38.1	29.8
3204179	28.1	83.7	95.0	27.8
3256247	123.4	65.1	7.4	28.6
3275134	69.4	66.6	-132.2	28.8
3294401	235.7	56.4	-163.3	28.9
3294406	35.2	70.7	122.8	26.8
3300769	86.4	35.1	-44.6	25.2
3334320	67.3	14.0	3.6	26.2
3419251	196.2	74.9	-61.2	31.4
3444687	161.0	81.3	-33.8	29.6
3493247	167.8	36.2	-9.9	27.5
3619577	335.6	58.4	-150.5	23.4

Table A.3 continued from previous page

2012p158837	143.2	45.6	36.6	31.4
2012p659872	22.2	84.8	77.4	20.8
2013p135714	24.8	71.4	92.8	28.4
2013p472502	240.0	19.9	116.9	32.3
2013p691776	152.4	57.8	30.4	30.3
2013p763995	227.4	14.5	135.5	24.2
2013p931321	55.5	88.6	145.0	32.3
2014p150497	39.6	31.5	-104.6	31.1
2014p836248	237.3	10.6	140.5	28.4
2015p020648	336.2	37.3	143.7	29.8
2015p267492	193.8	39.3	-100.7	27.1
2015p497791	161.7	77.9	-24.9	32.1
2015p539855	355.0	74.1	18.7	26.9
2015p897797	13.2	88.5	85.1	28.8
2016p239623	83.5	39.2	28.9	30.3
2016p367364	255.8	77.5	-129.6	29.6
2016p407264	350.3	74.5	36.9	29.5
2016p548022	357.7	82.7	85.5	31.9
2016p625356	2.7	84.0	87.0	23.9
2016p742264	193.9	63.2	-80.2	27.0
2016p767511	229.7	21.7	109.7	30.6
2016p854353	235.4	62.7	-143.1	27.9
2016p882825	347.4	88.4	127.3	24.4
2016p886198	165.3	82.9	-128.2	25.3
2016p957315	83.3	87.3	-148.7	27.3
2017p190305	25.1	44.3	75.1	28.1
2017p271473	28.3	66.8	87.9	28.3
2017p363282	216.7	30.9	110.9	27.9
2017p363577	197.3	31.7	90.8	26.6
2017p514329	333.0	15.4	-158.0	27.8
2017p516586	26.4	85.7	87.8	28.2
2017p784816	211.4	31.9	95.8	22.6
2017p924347	36.5	69.9	-133.6	23.7
2018p069063	343.0	77.6	36.3	29.9
2018p072548	123.0	46.0	-3.5	32.9
2018p072991	164.9	88.5	-54.0	28.5

Table A.3 continued from previous page

2018p073185	28.4	71.8	103.3	29.4
2018p073198	128.6	31.2	18.5	20.0
2018p153134	67.6	66.0	-71.0	33.8
2018p364597	337.4	88.6	8.6	27.7
2018p443045	251.0	76.6	177.6	30.3
2018p479843	110.1	37.3	31.3	27.2
2018p552196	227.9	86.9	-150.0	25.3

A.4 Stereonets and PT axis contour plots of the final focal mechanisms catalogue

This appendix contains the stereonets (left) and PT contour plots (right) of well-constrained events. Solid black lines indicate the focal mechanism solution. Blue points are compressions and red points are dilatations. Orange and green contours denote the posterior distribution of the P- and T-axes in orange and green, respectively.

2015p897797

2016p239623

2016p367364

2016p407264

2016p548022

2016p625356

2016p742264

2016p767511

2016p854353

2016p882825

2016p886198

2016p957315

2017p190305

2017p271473

2017p363282

2018p069063

2018p072548

2018p072991

2018p073185

2018p073198

2018p153134

2018p364597

2018p443045

2018p479843

2018p552196

B.1 NonLinLoc input file

```
# =====
# NonLinLoc programs control file
# NonLinLoc Version 7.0 - March 2018
# =====

# =====
# Generic control file statements
# =====
CONTROL 3 54321
# output filename root
# Layer 2DGrid
TRANS SIMPLE -41.7638 172.9037 140
# For NZ3D model.
# Faultmap
MAPTRANS SIMPLE -41.7638 172.9037 140
#NZ3D downloaded from https://zenodo.org/record/1043558#.XYLQtEEzZKg
# NZ3D uses an East Negative co-ordinate system. The flags in the Vel2Grid3D.c
# file need to be checked - none should be set.
# =====
# END of Generic control file statements
# =====

# =====
# Vel2Grid3D control file statements
# =====
# output filename root
VGOUT VEL/Porangahau_3D
# NZ3D
# wave type (VGTYPE wave_type (P, S))
#VGTYPE P
VGTYPE S
VGINP VEL/donna_nz3d_2.1_S.inp SIMUL2K 0 0 0
VGGRID 500 500 100 -200.0 -550.0 -3.0 1.0 1.0 1.0 SLOW_LEN
# =====
```

```
# END of Vel2Grid control file statements
# =====

# =====
# Grid2Time control file statements
# =====
GTFILES  VEL/Porangahau_3D TIME/Porangahau_3D P
GTMODE  GRID3D ANGLES_YES
INCLUDE  stations_nllloc.in
GT_PLFD  1.0e-3  1
# =====
# END of Grid2Time control file statements
# =====

# =====
# NLLoc control file statements
# =====
LOCSIG  ODPS VUW
LOCFILES IN/*.nll NLLoc_OBS TIME/Porangahau_3D OUT/located
LOCHYPOUT SAVE_NLLoc_ALL
LOCSEARCH  OCT 10 10 10 0.01 10000 5000 1 0
LOCGRID  500 500 100 -200 -550 -3.0 1.0 1.0 1.0 PROB_DENSITY SAVE
LOCMETH  EDT_OT_WT 9999.0 4 -1 -1 -1 6 -1.0 0
LOCGAU  0.2 10.0
LOCGAU2  0.07 0. 0.5
LOCPHASEID  P  P p G PN PG
LOCPHASEID  S  S s G SN SG
LOCPHSTAT  9999.0 -1 9999.0 1.0 1.0 9999.9 -9999.9 9999.9
LOCANGLES  ANGLES_YES 5
LOCMAG  ML_HB 1.0 1.0 0.0029
LOCQUAL2ERR 0.05 0.1 0.2 0.6 99999.9 99999.9 99999.9 99999.9 99999.9 99999.9
# =====
# END of NLLoc control file statements
# =====
```

B.2 GrowClust input file

```
***** GrowClust Control File *****
*****
*
*****
***** Event list *****
*****
* evlist_fmt (0 = evlist, 1 = phase, 2 = GrowClust, 3 = HypoInverse)
1
* fin_evlist (event list file name)
IN/evlist.txt
*
*****
***** Station list *****
*****
* stlist_fmt (0 = SEED channel, 1 = station name)
1
* fin_stlist (station list file name)
IN/stlist.txt
*
*****
***** XCOR data *****
*****
* xcordat_fmt (0 = binary, 1 = text), tdif_fmt (21 = tt2-tt1, 12 = tt1-tt2)
1 12
* fin_xcordat
IN/xcorr_data.txt
*
*****
*** Velocity Model / Travel Time Tables ***
*****
* fin_vzmdl (input vz model file)
IN/vzmodel.txt
* fout_vzfine (output, interpolated vz model file)
TT/vzfine.txt
* fout_pTT (output travel time table, P phase)
```

```

TT/tt.pg
* fout_sTT (output travel time table, S phase)
TT/tt.sg
*
*****
***** Travel Time Table Parameters *****
*****
* vpvs_factor rayparam_min (-1 = default)
  1.732          -1
* tt_dep0  tt_dep1  tt_ddep
  0.          180.    1.
* tt_del0  tt_del1  tt_ddel
  0.          400.    2.
*
*****
***** GrowClust Algorithm Parameters *****
*****
* rmin  delmax rmsmax
* previous: 0.6 80 0.5
  0.5    100    0.5
* rpsavgmin, rmincut  ngoodmin  iponly
  0          0          0          0
*
*****
***** Output files *****
*****
* nboot  nbranch_min
  10          1
* fout_cat (relocated catalog)
OUT/out.growclust_cat
* fout_clust (relocated cluster file)
OUT/out.growclust_clust
* fout_log (program log)
OUT/out.growclust_log
* fout_boot (bootstrap distribution)
OUT/out.growclust_boot
*****
*****

```