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Seismological Characterisation of Slow Slip Regions at the Northern Hikurangi Margin, Aotearoa New Zealand

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I hereby declare that this thesis and the work reported herein was composed by and originated entirely from me. Information derived from the published and unpublished work of others has been acknowledged in the text and references are given in the list of sources.

Amy Woodward (2026)

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Abstract

Slow slip events (SSEs) accommodate strain at plate boundaries without ground shaking. Understanding SSEs is important for characterising hazards, but the mechanisms for their occurrence is debated. The Hikurangi margin hosts a diverse range of SSEs along a margin that varies significantly in subduction zone geometry, lithology, rheology, and hydrology. The northern Hikurangi margin experiences shallow (<20 km depth) SSEs. Most studies have focused on the offshore region at depths <15 km. The onshore, deeper part of the slow slip region is comparatively poorly constrained. I use a temporary, high-density onshore seismograph network to provide seismological constraints on this deeper part of the slow slip patch, and to test whether it is continuous with the better-characterised offshore shallow SSE source region. First, I create catalogues of 3,071 earthquakes and 86 focal mechanisms and investigate two previously unconstrained $\sim M_w 6$ SSEs. Earthquake clusters occur during each SSE, consistent with offshore observations of $\sim M_w 7$ SSEs. Second, I use local earthquake tomography to investigate how variations in wavespeeds relate to properties of the slow slip source regions. High V_P/V_S (~ 1.9) in the subducting slab and at the plate interface is in close agreement with existing plate interface models, and is likely due to the composition of the slab and the presence of fluids. I suggest that linear bands of seismicity, normal-faulting focal mechanisms, and high V_P are associated with subducted seamounts. I interpret these seamounts as relatively buoyant plate interface topography that results in localised uplift and normal faulting, enhanced fluid pathways from slab mantle to the plate interface, and the supply of fluids from the slab to surface hot springs, cold seeps, and mud volcanoes. I argue that fluid availability, modulated by subducted topography and plate interface roughness, is likely a key control on the occurrence and downdip limit of shallow SSEs.

Publications and datasets arising from this thesis

Chapter 3: **Woodward, A.**, I. Bastow, R. Bell et al. (2025). Local Earthquake and Focal Mechanism Catalogue for the Northern Hikurangi Margin December 2017–October 2018 [Dataset], doi:[10.5281/zenodo.15058318](https://doi.org/10.5281/zenodo.15058318).

Chapter 3–4: **Woodward, A.**, I. D. Bastow, R. Bell, L. Wallace, K. Jacobs, S. Henrys, et al. (2026). Seismological Characterization of Northern Hikurangi Margin Slow Slip Regions Associated with Normal Faults, Seamounts, and Seeps. *Journal of Geophysical Research: Solid Earth*, **131**, e2025JB032916, doi:[10.1029/2025JB032916](https://doi.org/10.1029/2025JB032916).

Chapter 5: **Woodward, A.**, I.D. Bastow, R. Bell, M. Paulatto, S. Henrys, K. Jacobs. (*in review*). Local earthquake tomography of the northern Hikurangi margin: plate roughness, fluid distributions, and controls on slow slip. *Journal of Geophysical Research: Solid Earth*.

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Nomenclature

Conventions, symbols, and abbreviations used in this thesis are defined here and also where relevant in the main text. Frequently referred to timescales are also defined.

Conventions

Backazimuth The angle in degrees measured clockwise from north from a seismograph station to the great circle arc between a station and an earthquake.

Epicentral distance The distance in degrees between a seismograph station and an earthquake along the great circle path.

Low Frequency Earthquakes Small magnitude earthquakes (typically $< M_W 2$) with a predominant frequency of ~ 0.2 Hz. Very low frequency earthquakes (VLFs) have a dominant frequency of 0.01–0.05 Hz with M_W 3–4.

Slow earthquakes A collective term for low or very low frequency earthquakes, tremor, slow slip events, all defined as transient slip that causes unobservable ground shaking.

Slow slip event (SSE) Slip along a fault that does not produce a seismological signal. These events are detected with geodetic data.

Teleseismic Earthquakes originating from an epicentral distance of at least 30° from a seismograph station.

Tremor Low amplitude, extended duration seismic signal typically with frequencies 2–8 Hz.

Tsunami earthquake An earthquake that results in a disproportionately large tsunami relative to its magnitude.

Symbols

Δ	Epicentral distance
f	Frequency
M_0	Seismic moment
M_L	Local earthquake magnitude
M_w	Moment magnitude
V_p	P-wave seismic velocity
V_s	S-wave seismic velocity
Q	Quality factor

List of Abbreviations

3-D	Three-dimensional
DWS	Derivative weight sum
FOCMEC	FOCal MECHANism code developed by Snoke (2003)
gCMT	global Centroid Moment Tensor project (Ekström et al., 2012)
GNSS	Global Navigation Satellite System
GPS	Global Positioning System
IASP91	International Association of Seismology and the Physics of Earth's Interior IASP91 Travel-Time Tables of Kennett and Engdahl (1991)
IRIS	Incorporated Research Institutions for Seismology
KT	Kermadec-Tonga
LET	Local earthquake tomography
Moho	Mohorovičić discontinuity
NIDFB	North Island Dextral Fault Belt

NonLinLoc	Non-linear location code of (Lomax et al., 2023)	2013)
NZ3D-FWI	New Zealand 3-D - Full Waveform Inversion experiment (Bell et al., 2019)	SAGE Seismological Facility for the Advancement of Geoscience
RMS	Root Mean Square	SSE Slow Slip Event
SAC	Seismic Analysis Code (Helffrich et al.,	TVZ Taupō Volcanic Zone

Key geological timescales mentioned throughout the thesis:

Holocene: 11.7 ka–Present
Pleistocene: 2.6 Ma–11.7 ka
Pliocene: 5.3–2.6 Ma
Miocene: 23–5.3 Ma
Oligocene: 33–23 Ma

CHAPTER 1

Introduction

1.1 Overview

The surface of the solid Earth is formed of rigid tectonic plates that move relative to each other. At these plate boundaries strain is accumulated and released on timescales of hundreds of years. Traditionally, strain release at plate boundaries was considered to occur through fast earthquakes, with centimetres to metres of slip over seconds to minutes, or through steady aseismic creep at millimetres to centimetres per year. Over the last 25 years, however, geodetic and seismological observations have shown that subduction margins accommodate strain through a spectrum of slip modes, including slow slip events (SSEs), low-frequency earthquakes (LFEs; typically 1–8 Hz), very-low-frequency earthquakes (VLFs; typically <0.1 Hz), and tremor (typically 2–8 Hz) (e.g., [Dragert et al., 2001](#); [Rogers and Dragert, 2003](#); [Obara, 2002](#); [Obara et al., 2004](#); [Ide et al., 2007](#); [Peng and Gomberg, 2010](#)). Understanding how strain accommodation is partitioned between fast earthquakes, slow slip, tremor, and aseismic creep is therefore important for assessing seismic hazard at subduction zones.

SSEs are thought to occur in conditionally stable regions, the transitional region between locked, unstable slip (velocity weakening) and creeping, stable slip (velocity strengthening); these behaviours depend on the frictional properties of the fault plane ([Schwartz and Rokosky, 2007](#)). Frictional properties are sensitive to temperature, mineralogy/lithology, the presence of fluids (pore fluid pressure), heterogeneities (e.g., fault plane asperities), and permeability ([Scholz, 1998](#); [Schwartz and Rokosky, 2007](#)). SSEs result in millimetre to centimetre-scale slip over days to years.

Broadly, SSEs can be categorised into two types based on the plate interface depth at which they occur: deep (≈ 30 – 50 km depth) and shallow (< 20 km depth) (e.g., [Dragert et al., 2001](#);

[Obara, 2002](#); [Wallace et al., 2016](#)). Deep SSEs have been well-constrained geodetically as they tend to occur in the onshore forearc, thus allowing for inversion of land-based Global Navigation Satellite System (GNSS) stations to determine their spatial and temporal extent (e.g., [Bartlow et al., 2011](#)). Deep SSEs typically occur downdip of the seismogenic zone or brittle-ductile transition, and are widely attributed to metamorphic dehydration reactions at 350–450°C that may increase pore fluid pressure (e.g., [Scholz, 1998](#); [Liu and Rice, 2007](#); [Peacock, 2009](#); [Audet et al., 2009](#); [Audet and Kim, 2016](#)). In contrast, shallow SSEs primarily occur in the offshore forearc, relying on constraints from seafloor pressure gauges, potentially distant land-based GNSS stations, or seismological proxies for slow slip, such as tremor or VLFs (e.g., [Woods et al., 2024](#)). Shallow SSEs have been observed at the northern Hikurangi margin, Costa Rica, Ecuador, Boso Peninsula, and the Ryukyu trench (e.g., [Wallace et al., 2012a](#); [Dixon et al., 2014](#); [Ozawa et al., 2007](#); [Nishimura, 2014](#); [Vallée et al., 2013](#)). Shallow slow slip source region temperatures vary widely, from ~12–230°C, suggesting that the thermal state is not the primary control ([Saffer and Wallace, 2015](#)), and there is no clear global correlation between shallow slow slip depth and temperature-controlled dehydration. Other mechanisms for shallow SSE initiation, such as overpressured fluids or a rough plate interface that reduces effective stress, have been invoked to explain their occurrence (e.g., [Bassett et al., 2014](#); [Bangs et al., 2023](#)).

The northern Hikurangi margin is unusual because its short trench-to-coastline distance means that shallow SSEs straddle the onshore and offshore forearc. Here, SSEs primarily occur at <20 km depth, though occasionally deeper ([Wallace et al., 2012a](#)), providing a unique opportunity to combine onshore and offshore geologic and geophysical constraints on SSEs and the physical properties of their source regions. The source region of the offshore portion of the slow slip patch has been particularly well-characterised, through, for example, seafloor drilling and controlled-source seismic surveys (e.g., [Bell et al., 2019](#); [Gase et al., 2023](#)). The onshore portion (>10–15 km depth) of the shallow SSE patch has been comparatively little studied, due to the depth limitations of drilling or offshore active source seismic surveys. This thesis makes use of a temporary, dense onshore seismograph network, together with the permanent GeoNet seismic and GNSS data, to investigate three linked research questions:

1. What is the character of the deeper (>10 km) part of the SSE zone, and how does it compare with the well-constrained shallow zone?
2. How do microseismicity and SSEs interact and evolve, spatially and temporally?
3. What controls the downdip limit of shallow SSEs?

These questions are addressed using data from a temporary, dense onshore seismograph network (Figure 1.1), together with data from the permanent GeoNet seismograph and GNSS stations and offshore seismograph networks (*GNS Science, 2019, 2021; Wallace et al., 2014*). The onshore network consisted of 50 seismometers deployed near Gisborne from December 2017 to October 2018, making it possible to seismologically constrain the deeper, onshore portion of the shallow Northern Hikurangi slow slip region. First, a new local earthquake catalogue is derived to map seismicity in the upper plate, along the plate interface and in the subducting slab. Second, focal mechanisms are used to determine faulting styles; in particular, we consider the role of normal faulting in allowing fluids to migrate upwards from the slab, through the upper plate to the surface. Third, local earthquake tomography is used to determine variations in V_P and V_P/V_S centred on the onshore region, providing constraints on plate margin structure, including slab hydration and plate roughness. We consider the spatial and temporal evolution of earthquakes in relation to time-dependent slow slip models, and assess whether this evolution is controlled by the slow slip patch or by persistent structural and/or fluid-related heterogeneity. Together, these observations test whether the deeper onshore structure of the northern Hikurangi margin is continuous with the better-characterised offshore portion, and whether the occurrence and downdip limit of shallow SSEs is controlled by changes in plate roughness and fluid availability.

1.2 SSEs

During SSEs, episodic slip occurs faster than the ambient plate convergence rate, but too slowly to generate clear radiated seismic waves on standard seismograms (e.g., *Hirose et al., 1999; Obara et al., 2004; Schwartz and Rokosky, 2007; Peng and Gomberg, 2010*). Compared to earthquakes, SSEs have lower slip rates, propagation velocities, stress drops, and longer durations.

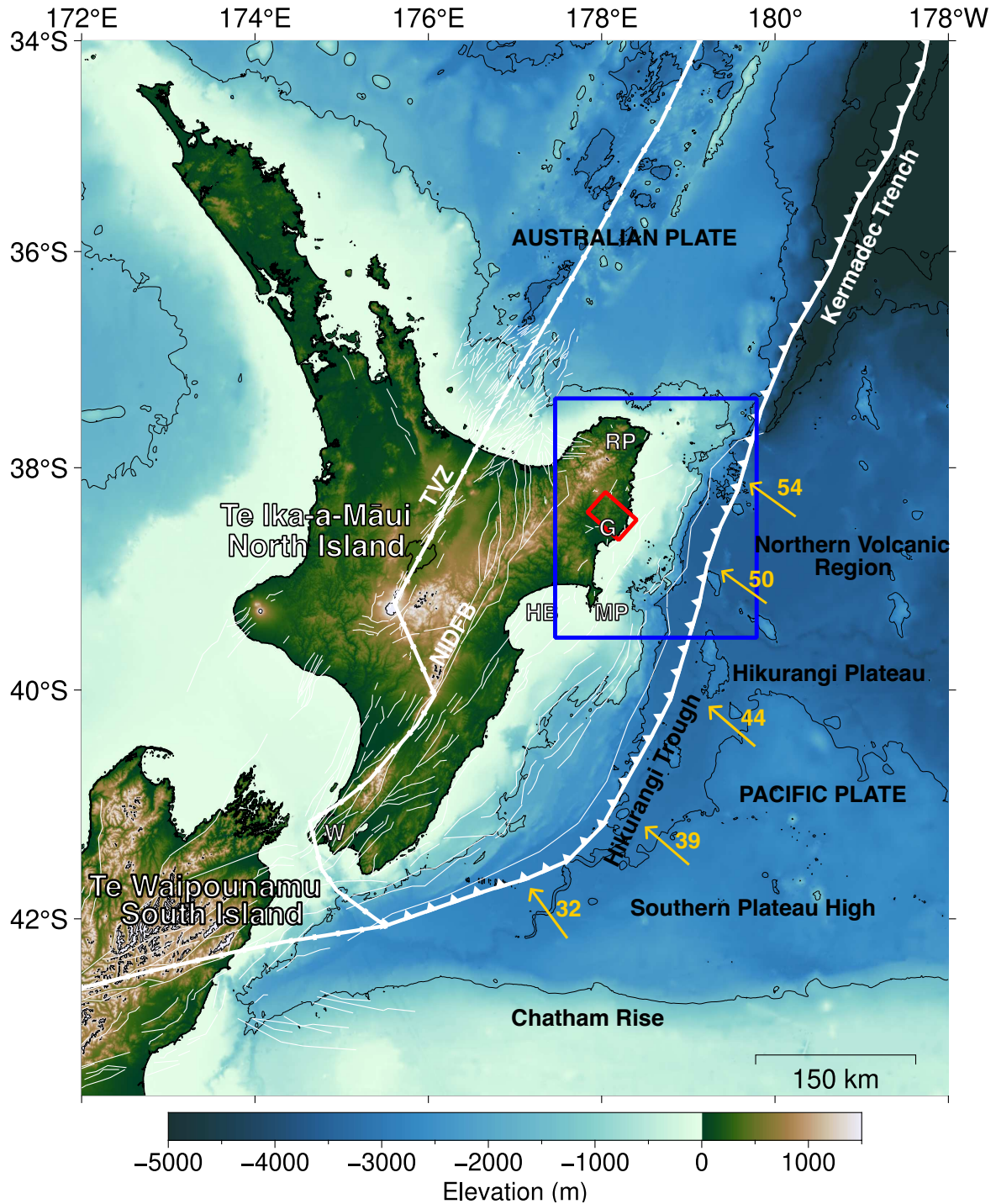


Figure 1.1: Tectonic setting of North Island and the Hikurangi margin and key geologic features. Blue box: study region in this thesis. Red box: temporary seismograph network near Gisborne. Contours: 1,500 m elevation/depth intervals. Abbreviations: TVZ, Taupō Volcanic Zone; NIDFB, North Island Dextral Fault Belt; RP, Raukūmara Peninsula; G, Gisborne; MP, Mahia Peninsula; HB, Hawke Bay; W, Wellington. Yellow arrows: plate convergence rate (mm/yr; *Wallace et al.*, 2012a). Thick white lines: plate boundaries (*Bird*, 2003). Thin white lines: faults (*Seebeck et al.*, 2022). Bathymetry data: *Tozer et al.* (2019).

Following observations of SSEs in the early 2000s, installation of geodetic instruments has demonstrated that SSEs occur globally at subduction zones, with up to tens of centimetres of movement over days to years. SSEs have primarily been observed in circum-Pacific subduction zones (e.g., [Douglas et al., 2005](#); [Wallace and Beavan, 2006](#); [McCaffrey et al., 2008](#); [Bartlow et al., 2014](#)), because they host the largest earthquakes and SSEs and are often the most heavily instrumented regions, but were first observed in continental strike-slip settings (as discussed in [Bürgmann, 2018](#)), and have more recently been observed on normal faults in rifting settings (e.g., [Wood et al., 2025](#)).

SSEs can have an equivalent moment magnitude to $M_W \sim 7$ earthquakes, potentially accommodating a large proportion of the overall slip budget. There have been observations of SSEs before megathrust earthquakes (preslip; e.g., [Dixon et al., 2014](#); [Ozawa et al., 2007](#); [Kato et al., 2012](#); [Ito et al., 2013](#)), and SSEs relieving stress increases on faults from nearby earthquake ruptures (afterslip; [Bürgmann, 2018](#)). SSEs may be triggered by static stress changes from earthquakes ([Wallace et al., 2017](#); [Araki et al., 2017](#)), and SSEs may in turn trigger large magnitude earthquakes (e.g., [Radiguet et al., 2016](#)). For example, two SSEs migrated towards the nucleation point of the 2011 Tōhoku-Oki earthquake prior to rupture, suggesting that slow slip unlocked the plate interface (e.g., [Dixon et al., 2014](#); [Ozawa et al., 2007](#); [Kato et al., 2012](#); [Ito et al., 2013](#)). In the central Ecuador margin, SSEs occur in a region dominated by interseismic creep, characterised by low seismic coupling, though likely within smaller locked patches within the creeping region ([Vallée et al., 2013](#)).

The underlying mechanism that promotes SSE occurrence is an area of ongoing research, but is widely considered to be the manifestation of transitional fault zone stability, between conditions for seismic versus aseismic behaviour ([Scholz, 2002](#); [Heslot et al., 1994](#)). The processes that contribute to this transitional property likely vary with depth and between subduction zones, but may be due to frictional properties of rocks, low fault rigidity, low effective stress (frequently associated with elevated pore fluid pressures), or a combination of these factors ([Kodaira et al., 2004](#); [Liu and Rice, 2007](#); [Ito and Obara, 2006](#)). There are some commonalities observed in shallow and deep slow slip settings: tremor has been observed in both, suggesting the presence of geometric or frictional heterogeneity; high fluid pressures

have been invoked to explain occurrence, linked to frictional and rheological heterogeneity ([Ghosh et al., 2012](#); [Ando et al., 2012](#); [Song et al., 2009](#); [Audet et al., 2009](#)). The differences between shallow and deep SSEs are explored in the following section.

1.2.1 Deep SSEs

Deep-seated SSEs (>30–50 km depth) are observed geodetically as transient deformations and seismologically as tremor or low-frequency signals ([Dragert et al., 2001](#); [Gomberg, 2010](#)). They are often accompanied by non-volcanic tremor, and are sometimes referred to jointly as episodic tremor and slip (ETS) ([Rogers and Dragert, 2003](#)). Deep SSEs and/or tremor have been observed in Cascadia, Nankai (south-west Japan), central Mexico, Costa Rica, and the southern Hikurangi margin (e.g., [Dragert et al., 2001](#); [Radiguet et al., 2011](#); [Ide, 2012](#); [Fry et al., 2011](#)). These events typically occur close to, but just downdip of, the brittle-ductile transition between the subducting slab and overlying crust and mantle wedge ([Audet and Kim, 2016](#)). The extent of the seismogenic zone, and therefore the depth below which deep SSEs initiate, depends on material properties and pore fluid pressure around the plate interface ([Scholz, 1998](#)).

Deep SSE occurrence is generally attributed to elevated fluid pressure at the plate interface and/or reduced friction in fault gouges, based on numerical modelling and seismic observations (e.g., [Kodaira et al., 2004](#); [Liu and Rice, 2007](#); [Audet et al., 2009](#); [Song et al., 2009](#)). Metamorphic dehydration reactions are considered a key driver of these conditions, by increasing fluids and reducing permeability ([Liu and Rice, 2007](#); [Peacock, 2009](#); [Song et al., 2009](#); [Audet et al., 2009](#)). The abundance of subduction zone fluids at this depth is primarily controlled by metamorphic dehydration reactions, which depend on pressure-temperature conditions (e.g., [Peacock et al., 2011](#)). In some subduction zones, such as Cascadia, there is an apparent gap between the seismogenic zone and slow slip region ([Audet and Kim, 2016](#)). High V_P/V_S is commonly observed in subducting slabs, above which tremor and LFEs cluster, possibly associated with dehydration of the plate and high pore fluid pressure (e.g., at 40–45 km depth in Cascadia, Nankai and Mexico [Peacock, 2009](#); [Fagereng and Diener, 2011](#); [Brantut et al., 2011](#)). Deep SSEs may also be influenced by mantle wedge serpentinisation,

which affects permeability and fluid pathways and concentrations ([Katayama et al., 2012](#)). Deep SSE occurrence is therefore likely due to a combination of hydrological and petrological conditions, influenced by pressure-temperature conditions.

1.2.2 Shallow SSEs

Shallow SSEs (<20 km depth) have been observed at the Ryukyu trench, Boso Peninsula, central Ecuador, and northern Hikurangi margin ([Ozawa et al., 2007](#); [Obana and Kodaira, 2009](#); [Kato et al., 2012](#); [Vallée et al., 2013](#); [Nishimura, 2014](#)). At the northern Hikurangi margin, Costa Rica, Ecuador, Boso Peninsula, and Ryukyu trench, SSEs have been recorded with equivalent moment magnitude $M_w = 6.0\text{--}7.0$ with a duration of weeks to months. The updip limit of shallow SSEs is poorly constrained due to the challenges of instrumenting close to the trench, though shallow slow slip may continue all the way to the trench ([Wallace et al., 2016](#)). At the northern and central Japan and Costa Rica trenches, slow earthquake signatures (VLFs and tremor) have been observed within the seismogenic zone that has hosted large magnitude earthquakes ([Dixon et al., 2014](#); [Kato et al., 2012](#); [Ito et al., 2013](#); [Matsuzawa et al., 2015](#); [Yamashita et al., 2015](#)). In contrast, the Ecuador and Hikurangi margin host SSEs in the depth range dominated by interseismic creep, though there may be small locked patches within these mostly creeping zones ([Wallace et al., 2012a](#); [Vallée et al., 2013](#)). At the Boso Peninsula, Hikurangi margin, Japan trench, and Ecuador, SSEs are frequently accompanied by bursts of tremor, VLFs, and/or microseismicity ([Wallace et al., 2012a](#); [Ozawa et al., 2007](#); [Kato et al., 2012](#); [Vallée et al., 2013](#); [Todd et al., 2018](#)).

Shallow SSE occurrence has widely been attributed to overpressured fluids and low effective stress, combined with rock frictional heterogeneity, though specific conditions and associated seismic signatures vary widely between subduction zones ([Kodaira et al., 2004](#); [Kitajima and Saffer, 2012](#); [Bassett et al., 2014](#); [Bell et al., 2014](#)). During shallow SSEs, temporal variations in stress ratio (a parameter that describes the relative magnitude of the minimum, intermediate, and maximum principal stresses), seismicity rates, and anisotropy have been linked to fluid migration and fault valving processes ([Warren-Smith et al., 2019](#); [Nishikawa et al., 2021](#); [Zal et al., 2020](#)). Shallow slow slip patches are not consistently correlated with temperature,

porosity, or metamorphic dehydration reactions ([Saffer and Wallace, 2015](#)). For example, clay dehydration is the largest source of mineral-bound water in the 80–150°C temperature range of shallow slow slip regions, yet these fluids are concentrated downdip of SSEs in northern Japan and Costa Rica, within the slow slip patch at the northern Hikurangi margin, and updip of VLFs at the Nankai margin ([Saffer and Wallace, 2015](#)). Similarly, porosity in shallow slow slip regions ranges 2–40%, suggesting that rock rigidity may contribute to the conditions required for slow slip but cannot account for its occurrence alone ([Kitajima and Saffer, 2012](#); [Ellis et al., 2015](#)). Shallow slow slip source region lithologic and geometric heterogeneity occurs on cm-to-km length-scales, associated with interbedded volcanoclastic sediments and seamount subduction ([Barnes et al., 2020](#)). It is difficult to explain the occurrence of shallow SSEs with a single unifying theory; the Hikurangi margin is an ideal target for understanding the source region of shallow SSEs because it is possible to integrate onshore and offshore geologic and geophysical constraints.

1.3 Tectonic and geologic background

1.3.1 Formation and structure of the Hikurangi margin

At the Hikurangi margin, the Pacific Plate (including the Hikurangi Plateau) subducts beneath the Australian Plate at the east coast of North Island, Aotearoa New Zealand. Subduction initiated 24–30 Ma ([Ballance, 1976](#); [Pettinga, 1982](#)). The Hikurangi margin forms the southern part of the 3,000 km Tonga-Kermadec-Hikurangi subduction system. The southern Hikurangi margin lacks volcanism, and terminates where the boundary transitions to the Marlborough Fault System between the North and South Islands, then to the strike slip Alpine Fault system to the south (Figure 1.1; [Wallace et al. \(2012b\)](#)). The subducting slab has been imaged to >400 km depth in western North Island ([Williams et al., 2013](#); [Henrys et al., 2025](#)). The Taupō Volcanic Zone is an active volcanic arc, west of the forearc (Figure 1.1).

For at least the last few Myr, the east coast of the North Island has been rotating 3–4°/yr relative to the Australian Plate, enhancing back-arc rifting in the Taupō Volcanic Zone (TVZ) and transpression of the southern North Island (Figure 1.1; [Nicol et al., 2007](#)). This rotation

has resulted in considerable along-strike variation in convergence rate, from 60 mm/yr in the north to 20 mm/yr in the south. Despite the oblique relative plate motion, convergence is largely margin-normal along the entire margin; margin-parallel plate motion is accommodated through rotation and the North Island Dextral Fault Belt (NIDFB), which displays strike slip motion (Figure 1.1; [Wallace et al., 2004](#)).

Geological and palaeoseismological data suggest that ten great subduction zone earthquakes have occurred in the last 7,000 years ([Henrys et al., 2020](#)), though historical records only cover the last 170 years and lack earthquakes larger than M_w 7.2 ([Wallace et al., 2009](#); [Henrys et al., 2020](#)). Crustal strain accumulation from GPS stations suggests that the northern Hikurangi plate interface could host an earthquake with $M_w \geq 8.0$ ([Henrys et al., 2020](#)), though recent studies have suggested that SSEs may accommodate all accumulated strain ([Carlson et al., 2026](#)).

1.3.2 The Hikurangi Plateau

The Hikurangi Plateau is a Cretaceous Large Igneous Province (LIP), formed between ~96 and 122 Ma as part of the Hikurangi-Manihiki-Ontong Java Plateau ([Mortimer and Parkinson, 1996](#); [Reyners et al., 2011](#)). Oceanic LIP crust is typically 1.5–4 times thicker than normal oceanic crust ([Coffin and Eldholm, 1994](#)). Rifting, then seafloor spreading at the Osborn Trough, resulted in the separation of the Hikurangi and Manihiki Plateaux ([Worthington et al., 2006](#)). The Hikurangi Plateau met and eventually underplated the Gondwana convergent margin at 100 Ma ([Davy et al., 2008](#); [Hoernle et al., 2020](#)). The southern limit of the Hikurangi Plateau is marked by the Chatham Rise (Figure 1.1). Today, the western Hikurangi Plateau is 7–12 km thick, and is subducting beneath the Australian plate ([Davy et al., 2008](#); [Mochizuki et al., 2019](#); [Bassett et al., 2023](#)). The subduction of the Plateau has resulted in a subaerial forearc and broad upper plate deformation ([Litchfield et al., 2007](#); [Nicol et al., 2007](#)).

At the northern Hikurangi margin, sediment cover overlying the Hikurangi Plateau is <1 km thick, with numerous seamounts protruding above the sediment (Figure 1.1 Northern Volcanic Region; [Barnes et al., 2010](#)). Dredge samples and drilling demonstrate that the plateau is primarily formed of 96–118 Ma tholeiitic basaltic basement ([Hoernle et al., 2010](#)). The

basaltic basement is overlain by Cretaceous clastic sedimentary rocks, as well as chalks and mudstones of Late Cretaceous to Early Oligocene age ([Barnes et al., 2020](#); [Davy et al., 2008](#)). Seamounts, formed 52–99 Ma, are formed of alkali basalt ([Hoernle et al., 2010](#)) and younger Cenozoic volcanoes ([Timm et al., 2010](#)). Drilling data shows that the seamounts are formed of interbedded smectite clay-rich volcanoclastic conglomerate, silty volcanic claystone, and basaltic material at the seamount flank, with vesicular basalt and volcanoclastic conglomerate at the top ([Gase et al., 2023](#)). The active outer accretionary wedge is <20 km wide, attributed to tectonic erosion due to subducting seamounts ([Lewis et al., 1998](#)). In contrast, the southern Hikurangi margin has ~5 km of sediment cover, which would bury any seamounts if present (Figure 1.1 Southern Plateau High), and the accretionary wedge is 30–70 km wide ([Barnes and De Lépinay, 1997](#)). Subducting seamounts enhance plate roughness, giving rise to variable stress and fluid pressure conditions and heterogeneous lithologies and frictional properties that have been linked with SSE recurrence ([Shaddox and Schwartz, 2019](#); [Barnes et al., 2020](#)). They also damage the upper plate, producing enhanced fracture networks that provide pathways for fluid migration ([Arai et al., 2017](#); [Bangs et al., 2023](#)). Several studies have characterised fluid pressure variations through the SSE cycle, observing an accumulation and release of fluid pressure before and during SSEs ([Warren-Smith et al., 2019](#); [Zal et al., 2020](#)).

The subduction of the relatively buoyant Hikurangi Plateau has resulted in a subaerial forearc and widespread uplift of the east coast of North Island ([Litchfield et al., 2007](#)). Consequently, at the east coast, the plate interface is at ~12 km depth and the margin is 40–120 km offshore. In comparison, coastlines at other subduction zones are 100–200 km from the margin, requiring different instrumentation to provide geophysical constraints. The unusually shallow plate interface and short distance between the trench and coastline at the Hikurangi margin mean that land-based instrumentation can be used to provide constraints on the shallow portion of the slow slip zone, making the Hikurangi margin a unique natural laboratory in which to study these processes. Unique to the Hikurangi margin is the subaerial forearc, which hosts numerous surface seeps, allowing direct sampling of fluids released at a range of depths, and providing constraints on fluid concentrations and compositions with depth ([Reyes et al., 2022](#)).

1.3.3 Hikurangi margin slow slip events

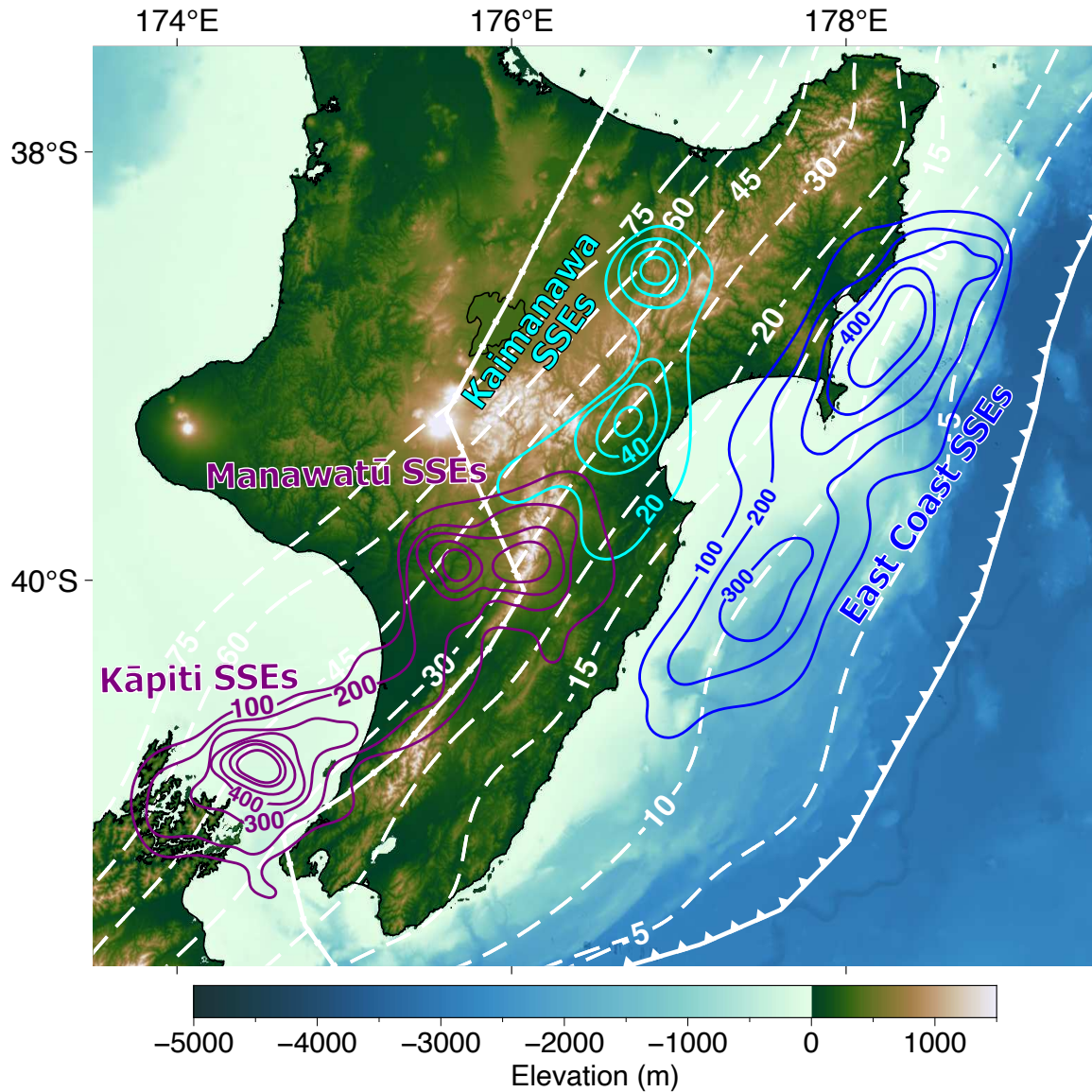


Figure 1.2: Cumulative slow slip in North Island from 2002–2014 (Wallace, 2020). Southern Hikurangi margin SSEs (pink contours): 100 mm intervals for Kāpiti and Manawātū SSEs. Central Hikurangi margin SSEs (cyan contours): 20 mm intervals for Kaimanawa SSEs in 2006 and 2008. Northern Hikurangi margin SSEs (blue contours): 100 mm intervals for East Coast SSEs. White solid lines: plate boundaries (Bird, 2003). White dashed lines: plate interface depth (km; Williams *et al.*, 2013).

Constraints on SSEs along the Hikurangi margin, obtained from the national continuously operating Global Position System (cGPS) instruments, Interferometric Synthetic Aperture Radar (InSAR), temporary deployments of absolute pressure gauges (APGs) and borehole instruments, demonstrate a diverse range of SSE characteristics in terms of depth, duration, magnitude, and recurrence (e.g., Wallace and Beavan, 2010; Woods *et al.*, 2024; Wallace, 2020).

There is significant along-strike variation in subducting sediment thickness and seamount subduction, which affects lithology, fluid pressure, and permeability of the megathrust, in turn controlling variations in plate coupling, seismicity distribution, and slow slip ([Henry et al., 2025](#)). Hikurangi margin SSEs occur just below the downdip limit of decadal interseismic coupling (based on the GPS velocities averaged over the GNSS deployment since 1995, Figure 1.4), which varies considerably along-margin ([Wallace and Beavan, 2010](#)). Interseismic coupling ranges between 0 and 1, based on the short-term creep rate and the long-term averaged slip rate. An interseismic coupling value of 0 represents a fault that is creeping at the long-term slip rate; an interseismic coupling value of 1 represents a fault that has no creep in the interseismic period. The SSEs at the Hikurangi margin can be broadly categorised based on depth, magnitude, and recurrence interval. Figure 1.2 shows the northern, central, and southern Hikurangi margin SSEs, referred to as the East Coast, Kaimanawa and Manawatū/Kāpiti SSEs respectively. The variation in SSEs and earthquake behaviour makes the Hikurangi margin well-suited to studying the conditions and mechanisms that may induce SSEs at a range of depths.

Table 1.1: Summary of along-margin variation in SSE and margin properties. Data from [Wallace et al. \(2009, 2012a\)](#); [Wallace \(2020\)](#).

	Northern margin	Central margin	Southern margin
Sediment thickness	<1 km	Transition	5 km
SSE recurrence interval	6–24 months	Observed in 2006 and 2008	1–2 / 4–5 years
SSE depth	< 15 km	30–75 km	15–50 km
SSE duration	2–3 weeks	1–2 months	1–2 years
SSE magnitude	6.0–7.0	6.5–6.6	≥ 7.0
Plate coupling	Weakly coupled with some locked patches offshore	Transition	Almost 100% coupled during inter-SSE; creeping on decadal timescales
Convergence rate	50–54 mm/yr	40–50 mm/yr	32–42 mm/yr

East Coast SSEs at the northern and central Hikurangi margin

At the northern Hikurangi margin (the focus of this thesis), SSEs occur within a region characterised by low interseismic coupling, high fluid content, and pronounced structural

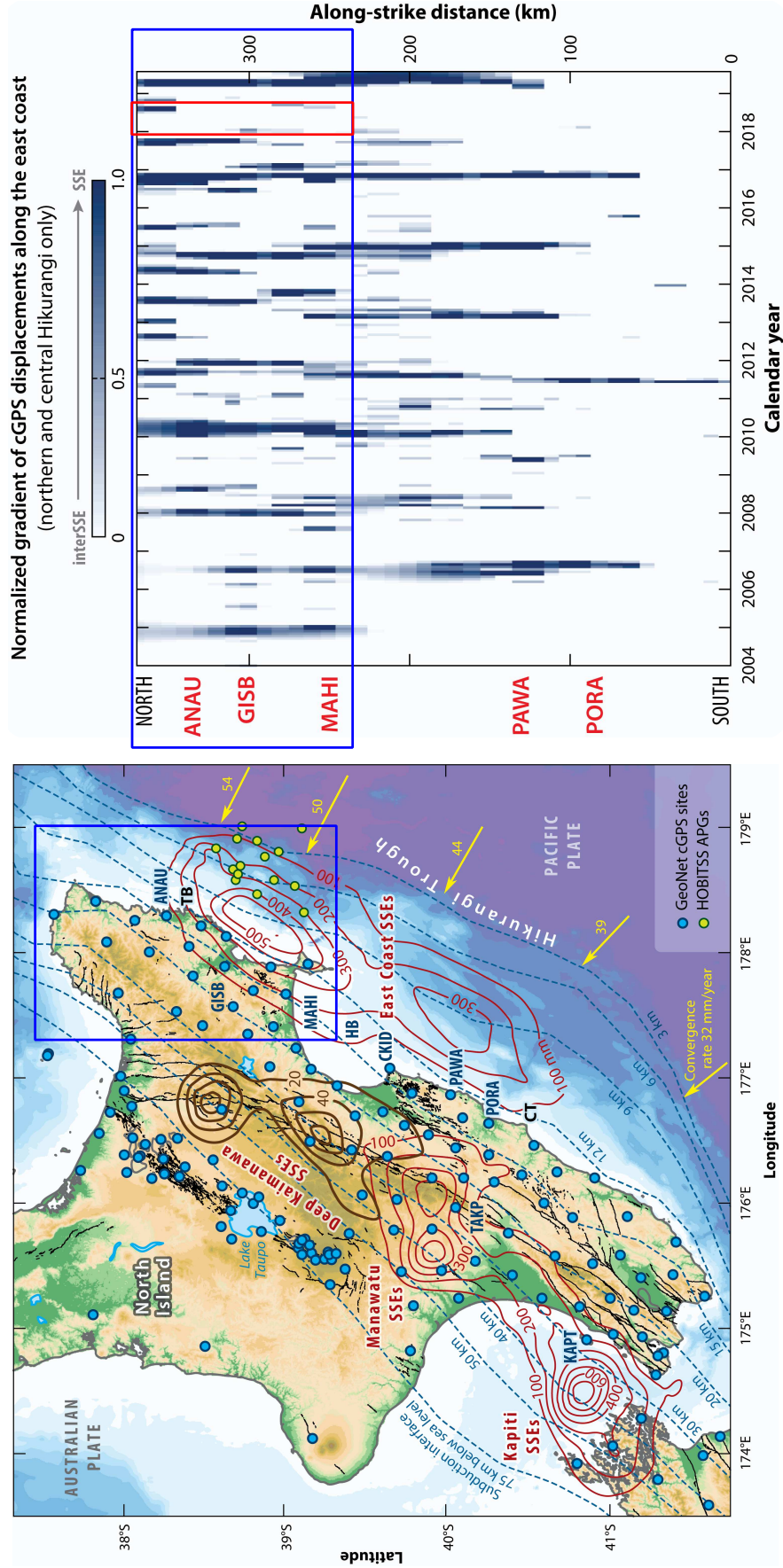


Figure 1.3: Figure adapted from Wallace (2020). Left) Cumulative slow slip patches 2002–2014. Blue circles: GPS sites. Labelled GPS sites are shown in plot on right. Abbreviations: APG, absolute pressure gauge; cGPS, continuously operating global positioning system; CKID, Cape Kidnappers; CT, Cape Turnagain; HB, Hawke Bay; TB, Tolaga Bay. Blue box: study region of this thesis. Right) Shallow east coast SSEs 2004–mid-2019. Darker colours correspond to faster rates of position change in cGPS data, i.e., the occurrence of SSEs. Blue box: study region of this thesis. Red box: time period of this thesis.

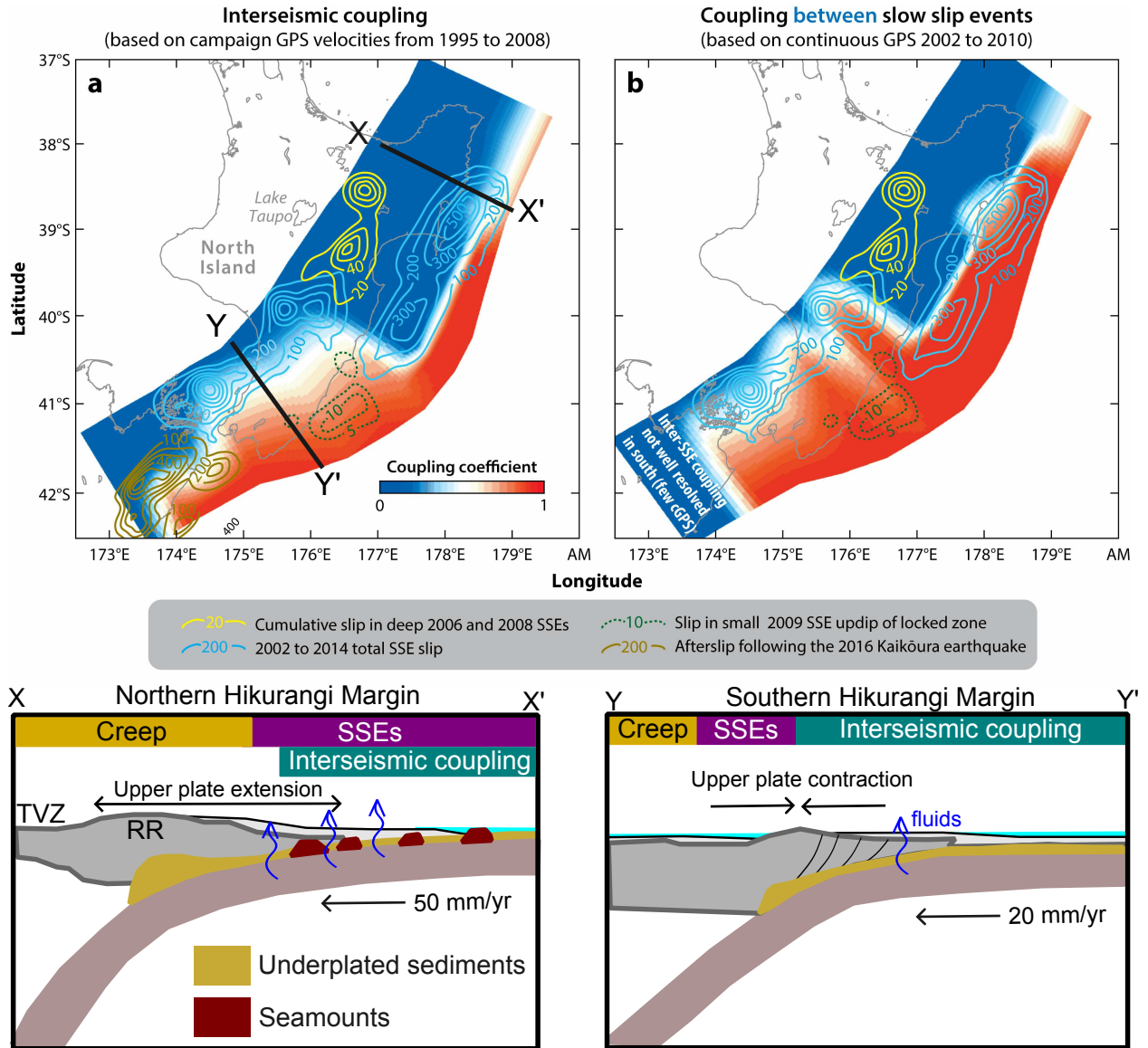


Figure 1.4: Panels a) and b) adapted from *Wallace (2020)*; *Henrys et al. (2025)*. a) Coupling coefficient 0: creeping at long-term slip rate. Coupling coefficient 1: no creep. Coupling is determined from GPS velocities between 1995 and 2008. b) Coupling coefficient determined from GPS velocities for the periods between SSEs. Schematics show cross-sections through the northern Hikurangi margin (XX') and the northern Hikurangi margin (YY'). Abbreviations: RR, Raukūmara Range; TVZ, Taupō Volcanic Zone.

heterogeneity, including complex upper plate fracture networks and subducted topography (Wallace, 2020; Bangs *et al.*, 2023). Shallow SSEs (≤ 20 km depth), referred to as the East Coast SSEs, recur every 6–24 months and last from weeks to several months (Wallace, 2020). The upper plate at the northern Hikurangi margin is in a state of extension (Townend *et al.*, 2012; Behboudi *et al.*, 2023; Henrys *et al.*, 2025), creating fractures that permit upward fluid migration. The subducting Pacific Plate has relatively thin sediment cover (≤ 1 km; Barnes *et al.*, 2010; Ghisetti *et al.*, 2016) and a rough plate interface where seamounts protrude above the sediment (Pedley *et al.*, 2010; Barnes *et al.*, 2020). The Hikurangi Plateau forms the portion of the Pacific Plate that subducts at the Hikurangi Trough (Mortimer and Parkinson, 1996); it is considerably thicker than the 4–9 km thick oceanic crust subducting at the northeastern end of the Hikurangi-Tonga-Kermadec subduction system (Davy *et al.*, 2008; Mochizuki *et al.*, 2019). The plateau is an ~ 11 km-thick large igneous province of tholeiitic basalt (Bassett *et al.*, 2023), formed 98–118 Ma, with numerous 52–99 Ma alkali basalt seamounts protruding from its surface (Figure 1.1; e.g., Hoernle *et al.*, 2010). Subduction of the relatively buoyant plateau has caused uplift of the east coast of the North Island (Litchfield *et al.*, 2007), resulting in a subaerial forearc (Figure 1.1).

Tsunami earthquakes generate much larger wave heights than expected, given their magnitude and lack of intense ground shaking (Kanamori, 1972). In 1947, two tsunami earthquakes occurred with magnitudes of M_w 6.9–7.1 that resulted in unusually large tsunami inundation heights of 6 and 10 m (Bell *et al.*, 2014). The behaviour of these tsunami earthquakes was attributed to unusually long rise times and very slow rupture speeds, likely over a subducting seamount (Bell *et al.*, 2014). Tsunami earthquakes are characterised by a lack of high-frequency energy and long source durations, associated with transitional fault behaviour (Bilek and Lay, 2002). Similar physical conditions likely account for the occurrence of SSEs and tsunami earthquakes.

Shallow SSEs at the central and northern Hikurangi margin, where the majority of slip occurs at depths < 15 km, are typically accompanied by up to 1 cm of coastal subsidence (Douglas *et al.*, 2005; Wallace, 2020). These SSEs occur every 6–24 months, depending on their magnitude, with a duration of weeks to months. Large magnitude SSEs, with equivalent $M_w > 7.0$ occurred

in 2014 and 2019 ([Wallace et al., 2016](#); [Woods et al., 2024](#)), typically every 1–2 years. Smaller magnitude SSEs, with equivalent moment magnitude of M_w 6.0–6.5, occur every 6–12 months.

Kaimanawa SSEs at the central Hikurangi margin

The Kaimanawa SSEs occur downdip of the well-documented, shallow, central Hikurangi margin East Coast SSEs, up to 75 km depth ([Wallace and Eberhart-Phillips, 2013](#)). Such events were recorded in 2006 and 2008, with equivalent M_w 6.6 for each event. They had a duration of 2–3 months and 2–3 cm slip. Swarms of tremor and VLFs have been detected downdip of the slow slip source region, associated with plate interface shearing, during and between geodetic observations of slow slip ([Romanet and Ide, 2019](#); [Aden-Antoniów et al., 2024](#)). The Kaimanawa SSEs are comparatively little studied relative to the East Coast and Kapiti and Manawatū SSEs.

Kapiti and Manawatū SSEs at the southern Hikurangi margin

The Kapiti and Manawatū SSEs recur every 4–5 years at the southern Hikurangi margin, with a duration of 1–2 years ([Wallace, 2020](#)). They have a similar equivalent M_w to the shallow SSEs (M_w 7.0), with >30 cm slip on the plate boundary. The Manawatū SSEs typically occur at 15–50 km depth, sometimes following a Kapiti SSE, which typically occurs at 25–50 km depth ([Wallace and Beavan, 2010](#); [Bartlow et al., 2014](#)). There are suggestions that the Kapiti and Manawatū SSEs follow an overall 10-year cycle of recurring slip with propagation of SSEs across the patches, though this is currently limited by the 20-year SSE catalogue ([Wallace and Beavan, 2006](#)). The locus of slip has migrated during SSEs, along strike and dip, with periods of slower and faster slip ([Wallace and Beavan, 2010](#)). The Kapiti SSEs have recurred approximately every 5 years, except for the 2016–2018 SSE, which was initiated ‘early’ following the 2016 M_w 7.8 Kaikōura earthquake ([Wallace et al., 2018](#)). The deeper, longer duration SSEs here show similar characteristics to those observed in Japan (e.g., [Hirose et al., 1999](#); [Hirose and Obara, 2005](#); [Ozawa et al., 2007](#)).

The southern Hikurangi margin displays strong interseismic coupling to ~40 km depth, which transitions sharply to a creeping megathrust at the central margin ([Wallace et al., 2004](#); [Wallace and Beavan, 2010](#)). As temperatures increase with depth, and as sediment thickness increases,

conditions for the deeper SSEs may arise ([Henrys et al., 2025](#)). The locked portion of the plate interface shows resistive zones, correlated with areas of greater friction and compressional strain ([Heise et al., 2023](#)).

Higher V_P , lower V_P/V_S , higher resistivity, and higher Q , compared to the northern Hikurangi margin, have been interpreted as fluid-poor basement rock ([Bassett et al., 2014](#); [Eberhart-Phillips and Fry, 2017](#); [Bassett et al., 2023](#); [Heise et al., 2023](#)). Surface seeps show lower flow rates and longer residence times, suggestive of a lower permeability forearc compared to the north ([Reyes et al., 2022](#)). The southwards transition to a compressional upper plate regime may account for the decreased permeability and change in geochemistry (e.g., [Barnes et al., 2019](#)). [Henrys et al. \(2025\)](#) suggested that reduced upper plate vertical permeability may increase upper plate and interface fluid pressures, accounting for the deeper interseismic coupling compared to the north, thus resulting in deeper slow slip patches.

1.4 Thesis outline

1.4.1 Research questions and objectives

Building on the motivation outlined above, this thesis focuses on the northern Hikurangi margin, where the shallowest and most frequent Hikurangi SSEs occur. Recent advances in understanding have come from ocean drilling ([Wallace et al., 2019](#); [Gase et al., 2023](#)), dense ocean-bottom seismograph networks (e.g., [Yarce et al., 2019](#); [Todd et al., 2018](#); [Shaddox and Schwartz, 2019](#); [Mochizuki et al., 2021](#)), 3-D seismic data ([Bell et al., 2019](#); [Arai et al., 2020](#); [Bangs et al., 2023](#)), geodetic data ([Wallace, 2020](#)), active-source seismic data ([Jacobs et al., 2020](#); [Bassett et al., 2022, 2023](#)), magnetotelluric data ([Heise et al., 2023](#)), and electromagnetic data ([Chesley et al., 2021, 2023](#)). With the exception of magnetotelluric and geodetic constraints, many of these observations are concentrated offshore. However, the 2002–2014 cumulative SSE distribution (Figure 1.2) suggests that the slow slip region extends to depths of ~20 km depth, beneath the onshore part of the Hikurangi margin.

This thesis forms part of the NZ3D-FWI project, which was designed to characterise the onshore portion of the slow slip region using a dense temporary network of onshore seismograph

stations. Instrumentation is detailed further in Chapter 2. The primary aim of this thesis is to use seismological observations to constrain plate margin architecture, fluid distribution, and microseismicity in the deeper onshore portion of the northern Hikurangi slow slip region.

1.4.2 Thesis structure

Chapter 2 details the seismograph networks used, including documentation of the NZ3D-FWI temporary deployment of 50 land-based seismometers. This chapter also details the permanent broadband and short-period seismometers and the GNSS stations, operated by GeoNet, used in this study.

Chapter 3 presents the methods used to create a local earthquake catalogue for the northern Hikurangi margin, using automated and manual earthquake detection methods. The earthquakes are used to derive a new 1-D velocity model for the region. Focal mechanisms are determined. This chapter addresses research questions 1 and 2.

Chapter 4 uses the focal mechanism and earthquake distributions derived in Chapter 3 to analyse seismicity distributions and subduction zone architecture. Links with slow slip events, subducted seamounts, fluid migration, and surface seeps are explored, using GNSS displacement data. Secondly, the earthquake catalogue is used to investigate spatiotemporal links with SSEs that occurred during the SSE deployment. This chapter addresses research questions 1, 2, and 3.

Chapter 5 uses the earthquake catalogue derived in Chapter 3, as well as existing catalogues, to derive a 3-D velocity model for the region using local earthquake tomography. We then use the tomography models to consider variations in fluid distributions in the margin and how this controls SSE distribution and downdip limits. This chapter addresses research questions 1, 2 and 3.

Chapter 6 summarises the main findings presented in this thesis and considers areas for future work.

CHAPTER 2

Networks and data

2.1 Introduction

This thesis uses several seismological methods to characterise the northern Hikurangi margin. An earthquake and focal mechanism catalogue is developed to analyse spatiotemporal relationships between SSEs and seismicity (Chapters 3 and 4). Local earthquake tomography is then used to derive V_P and V_P/V_S models for the same region (Chapter 5). These analyses use broadband and short period data from temporary and permanent seismograph networks.

Chapters 3 and 4 use data from a temporary onshore network deployed between December 2017 and October 2018 ([Bell et al., 2019](#)), together with permanent GeoNet stations ([GNS Science, 2021](#)). Chapter 5 uses the earthquake catalogue developed in the earlier chapters, supplemented by earthquakes recorded during a temporary offshore deployment ([Wallace et al., 2014](#); [Yarce et al., 2019](#)).

Table 2.1: Seismograph networks used in this study.

Code	Duration	Data centre	Instrument	DOI
NZ	Permanent	GeoNet	Güralp CMG-3ESPC	doi: 10.21420/G19Y-9D40
NZ	Permanent	GeoNet	LE-3DliteMkII / L4C-3D	doi: 10.21420/G19Y-9D40
3C	2017–2018	IRISDMC	Güralp CMG-6TD	doi: 10.7914/SN/3C.2017
YH	2014–2015	IRISDMC	Reftek	doi: 10.7914/SN/YH.2014

2.2 NZ3D-FWI project

Until recently, studies of SSEs at the northern Hikurangi margin focused on the offshore region (≤ 15 km), where the shallowest and most frequent events occur. Consequently, the

offshore region is comparatively well-constrained by controlled-source seismic surveys and ocean-bottom seismometer deployments (e.g., [Bell et al., 2010](#); [Arai et al., 2020](#); [Bassett et al., 2025](#)). In contrast, the onshore region is less well resolved, particularly with respect to microseismicity, fault networks, margin architecture, and velocity structure. Although SSEs also occur beneath the onshore margin, their downdip limit and source region properties are poorly constrained.

This thesis forms part of the NZ3D-FWI experiment, which was designed to improve constraints on the slow slip region across both the offshore and onshore northern Hikurangi margin ([Bell et al., 2019](#)).

The NZ3D-FWI project acquired four datasets:

- the Gisborne network: 50 land-based CMG–6TD seismometers (used in Chapters 3–5; Figure 2.1);
- the Waimata Valley network: 143 short-period DATA-CUBE3 and GSX3 instruments;
- a 3-D multi-channel hydrophone streamer dataset comprising four 6 km-long streamers (used in Chapter 4); and
- 99 ocean-bottom seismometers.

The ocean-bottom seismometers from the NZ3D-FWI project have been used to generate P-wave velocity models and investigate fluid distribution for an offshore 3-D volume ([Davy et al., 2021](#); [Frahm, 2025](#)). We make use of the onshore Gisborne network to study the portion of the plate interface downdip of the most frequent, shallow SSEs, where some SSEs nevertheless occur.

2.2.1 Gisborne network

Fifty seismometers were shipped from the NERC Geophysical Equipment Facility SEIS-UK; 49 were operational on arrival. From December 2017 to October 2018, 49 Gralp CMG–6TD 30 s instruments were deployed (Figure 2.1). The network covered an area of $\sim 15 \times 30$ km, with an average station spacing of ~ 2 km.

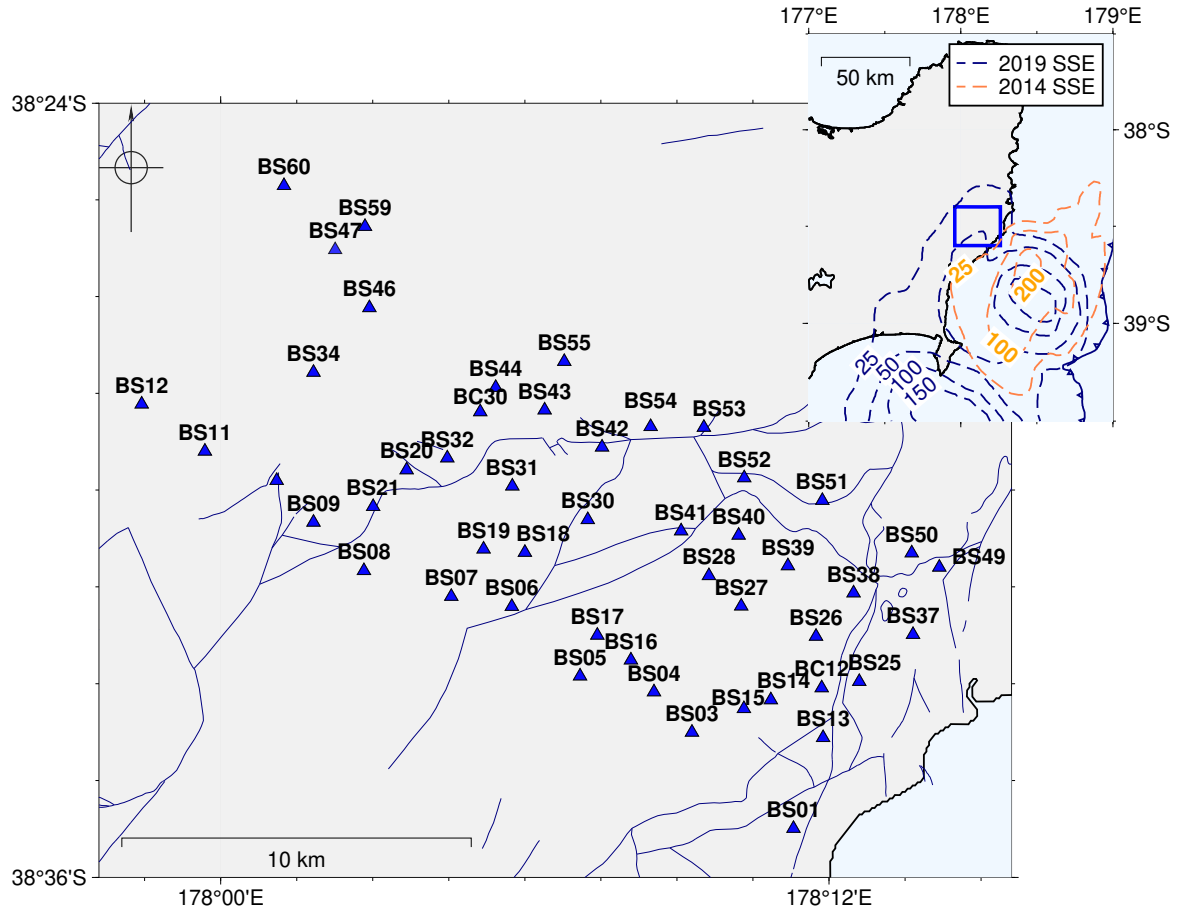


Figure 2.1: Temporary seismograph station deployment. Blue triangles: seismometers with station codes. Blue lines: faults (*Mazengarb and Speden, 2000*) Inset map: location of temporary network (blue rectangle) in study region. Blue line: Hikurangi margin. Dashed contours: 2014 and 2019 SSEs (*Wallace et al., 2016; Woods et al., 2024*).

The network was designed primarily for full-waveform inversion of controlled-source data, and was therefore much denser than a typical passive-source deployment designed for imaging crust or mantle structure. This dense spacing also provided an unusually high-resolution record of local seismicity, which is the focus of this thesis.

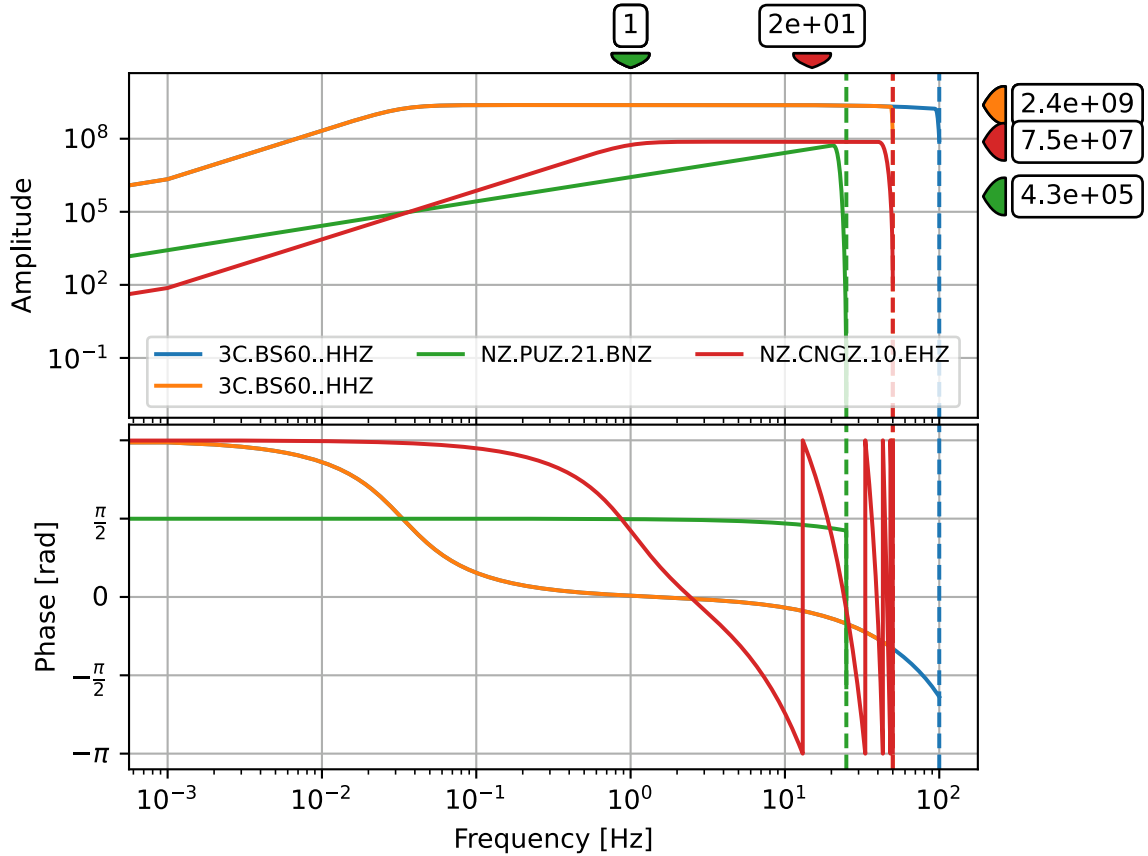


Figure 2.2: Amplitude response (V/m/s) of each instrument type used in this study. Station BS60 in blue: temporary network sampling at 200 Hz (December 2017–February 2018). Station BS60 in orange: temporary network sampling at 100 Hz (February 2018–October 2018). Station PUZ in green: permanent broadband instrument. Station CNGZ in red: permanent short period instrument.

The instruments had flat responses in the frequency range 0.03–100 Hz (Figure 2.2). The network recorded at 200 Hz from December 2017 to February 2018 and at 100 Hz from February to October 2018. For each instrument, a 60 cm deep hole in soil was back-filled with sand for thermal insulation.

The instruments were powered either by either two 12 V car batteries, or one 12V car battery and a 20 W solar panel. Data were recovered in February 2018, July 2018, and at decommissioning in October 2018. Stations for which data recovery was incomplete are

Table 2.2: Summary of recording periods for stations with partial data recovery.

Station	Comment
BS26	Operational December 2017–February 2018. Partial recovery of data from February 2018–October 2018.
BS43	Operational December 2017–February 2018. Partial recovery of data from February 2018–October 2018.
BS46	Operational February 2018–July 2018 due to stolen solar panel/battery.
BS47	Operational December 2017–February 2018 due to pig-related damage.
BS16	Operational December 2017–February 2018 and July 2018–October 2018.

detailed in Table 2.2. Of the land-based instruments, 87% were oriented to true north. The remainder were corrected prior to data archiving as detailed by *Bell et al. (2019)*.

All NZ3D-FWI land-based seismograph data are hosted at https://doi.org/10.7914/SN/3C_2017, with network code 3C, and are made available through the International Federation of Digital Seismograph Networks (FDSN).

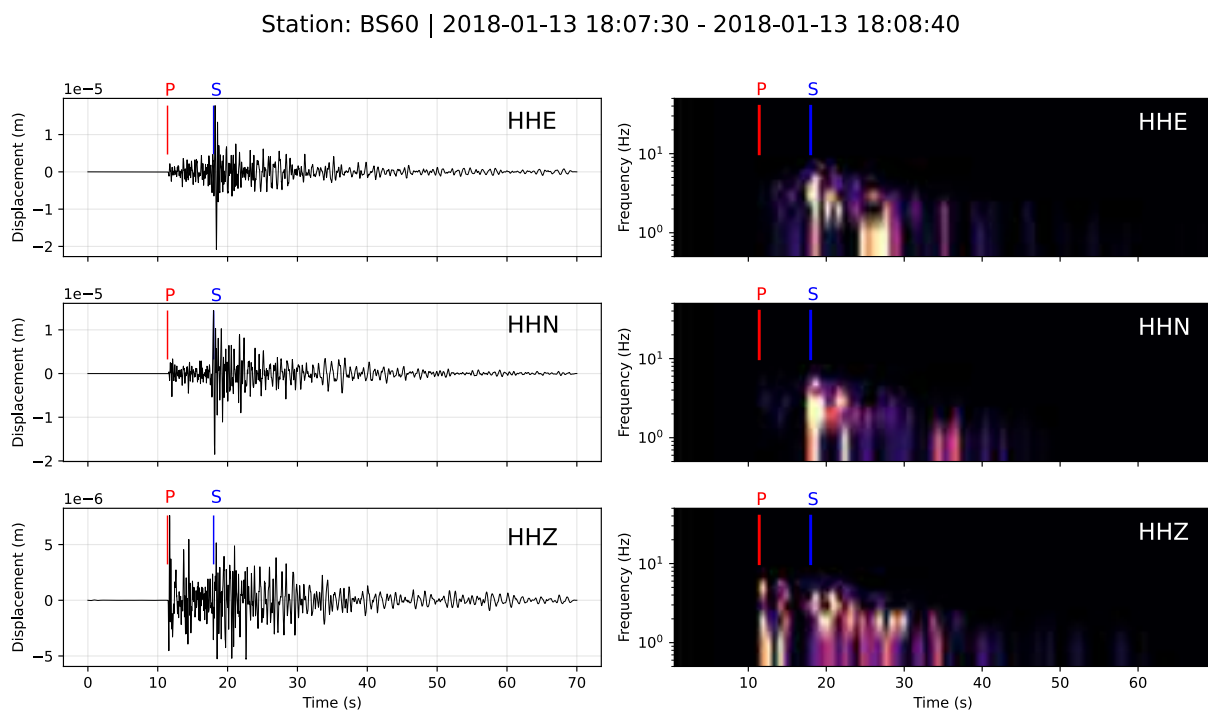


Figure 2.3: Example three-component seismographs (left column) and corresponding spectrograms (right column) recorded at station BS60, from the temporary Gisborne network. Data are Butterworth bandpass filtered with corner frequencies of 1 and 14 Hz. Seismograph data are shown for east (top row), north (middle row), vertical (bottom row) components. Example P- and S- picks are labelled.

2.3 Instrumentation in Aotearoa New Zealand

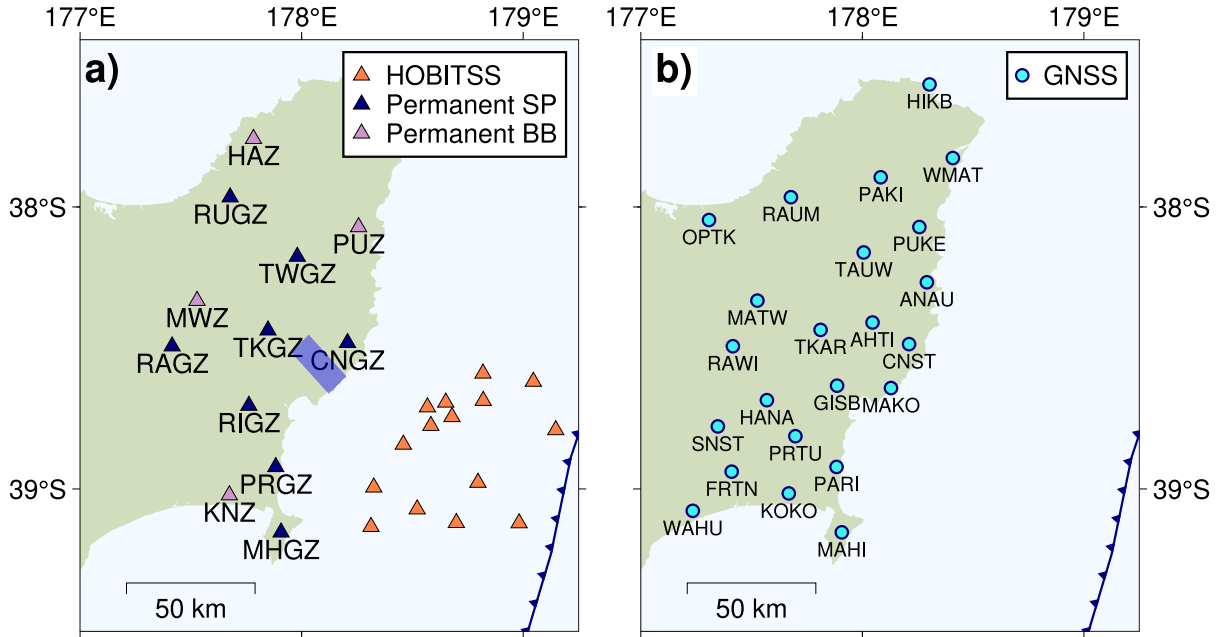


Figure 2.4: a) Seismograph stations used to supplement the Gisborne network. Orange triangles: temporary HOBITSS network (*Wallace et al., 2014*) Navy triangles: permanent short-period instruments (SP). Pink triangles: permanent broadband instruments (BB). Shaded blue rectangle: location of temporary Gisborne network. b) Cyan circles: GNSS stations used in this thesis.

2.3.1 Permanent seismograph networks

To extend spatial coverage beyond the temporary network, data from the permanent GeoNet network (network code NZ) of short-period and broadband seismometers were also used. Station spacing varies, but is typically 25–50 km across North Island. For this thesis, eight short-period and five broadband stations were used (Figure 2.4a); these are listed in Appendix A.

GeoNet is Aotearoa New Zealand’s national geological hazards monitoring system, and is optimised for rapid earthquake detection and magnitude estimation rather than for producing high-resolution research catalogues with low magnitudes of completeness. Earthquakes are detected and picked automatically using a short-term-average/long-term-average (STA/LTA) amplitude algorithm. Hypocentres are determined using the IASP91 velocity model (*Kennett and Engdahl, 1991*), which is not ideal for a subduction zone setting. The GeoNet catalogue

does not report hypocentral uncertainties and, where locations are poorly resolved, depths are fixed at 5, 12, 25, or 33 km. Although the catalogue reliably captures earthquake occurrences, hypocentres and origin times can be poorly constrained ([Warren-Smith et al., 2024](#)), which limits its suitability for analyses that depend strongly on accurate hypocentre information, such as local earthquake tomography.

2.3.2 Instrument response

Examples of the instrument response are shown in Figure 2.2 for temporary station BS60 sampling at 200 Hz (December 2017 to February 2018) and 100 Hz (February 2018 to October 2018), and for a short-period and broadband permanent seismometer.

Examples of power spectral density plots are shown for each instrument type and each sample rate used (Figures 2.5 and 2.6). These plots were used to assess background noise levels in the 1–14 Hz band used for earthquake detection and to confirm that the data quality was suitable for analysis relative to the high- and low-noise models of [Peterson \(1993\)](#).

2.3.3 Additional seismograph networks

The Hikurangi Ocean Bottom Investigation of Tremor and Slow Slip (HOBITSS) experiment (Figure 2.4) was a temporary deployment of ocean-bottom seismometers (OBS) and pressure gauges that operated from May 2014 to June 2015 ([Wallace et al., 2014](#)). The HOBITSS network was located updip of the Gisborne deployment, thereby offering complementary coverage of seismicity. In this thesis, we use the earthquake catalogue derived from HOBITSS ([Yarce et al., 2019](#)) to compare earthquake distributions (Chapter 3), to interpret the combined earthquake catalogues (Chapter 4), and to provide additional travel-time data for local earthquake tomography (Chapter 5).

The HOBITSS network comprised 15 OBS, deployed with 6–22 km spacing. Although three instruments were installed seaward of the trench, only those on the offshore outer forearc are used here, because of the earthquake selection criteria applied in this study. All OBS sampled at 100 Hz. Data from eight instruments are used in this thesis; the remainder either did not record sufficient data or lay outside the study region.

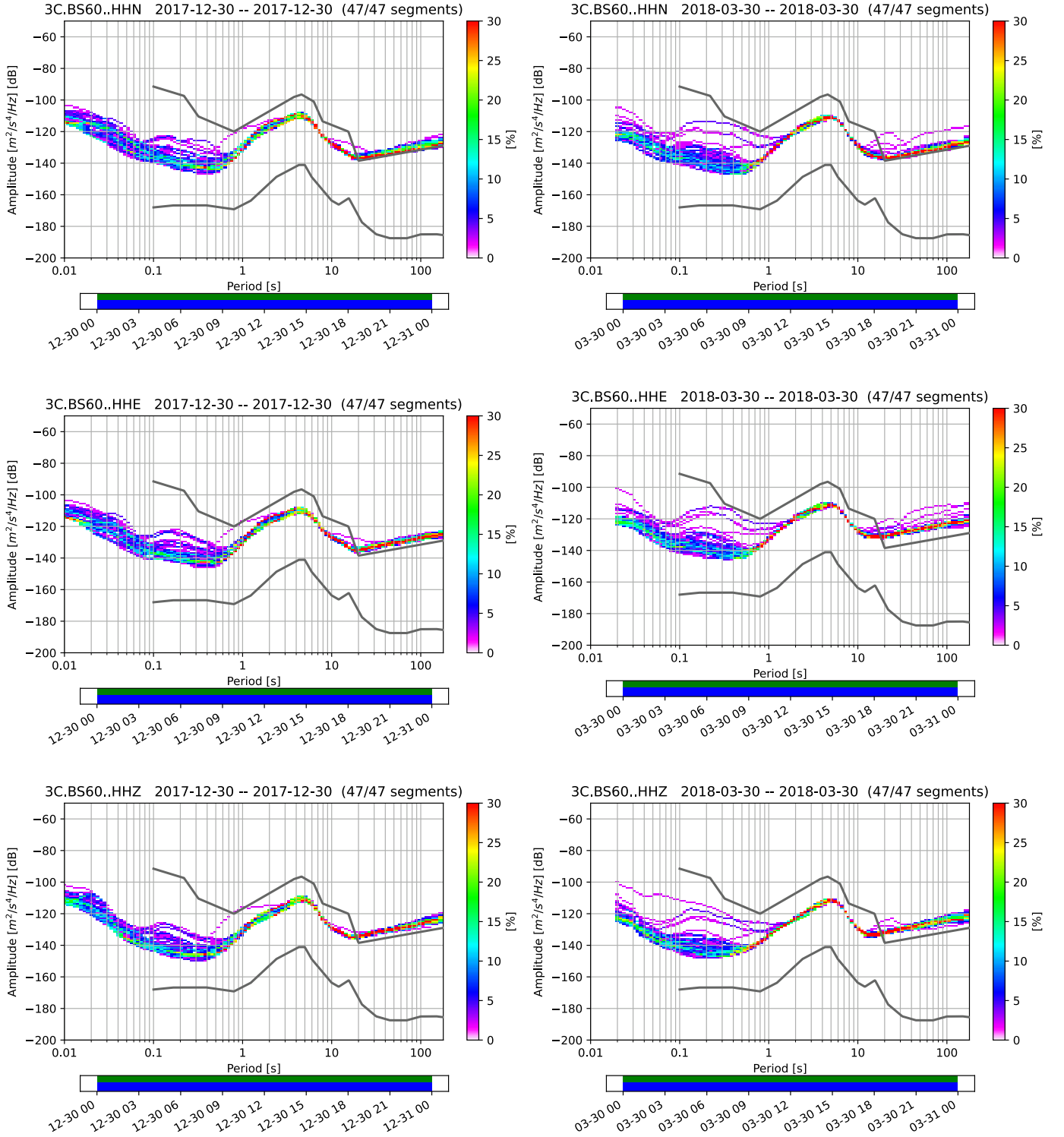


Figure 2.5: Probabilistic power spectral density (PSD) distributions showing typical background seismic noise levels for station BS60. Left column shows 24-hr window during 200 Hz sample rate period. Right column shows 24-hr window during 100 Hz sample rate period. PSD plots are plotted following the method of *McNamara (2004)*, for frequencies ranging from 0.005–100 Hz. Dark grey lines show the typical low- and high-noise model for broadband stations from *Peterson (1993)*. Top row shows north component; middle row shows east component; bottom row shows vertical component. Blue line shows observation period; green line represents data continuity (i.e., stations recorded continuously during this 24-hr window).

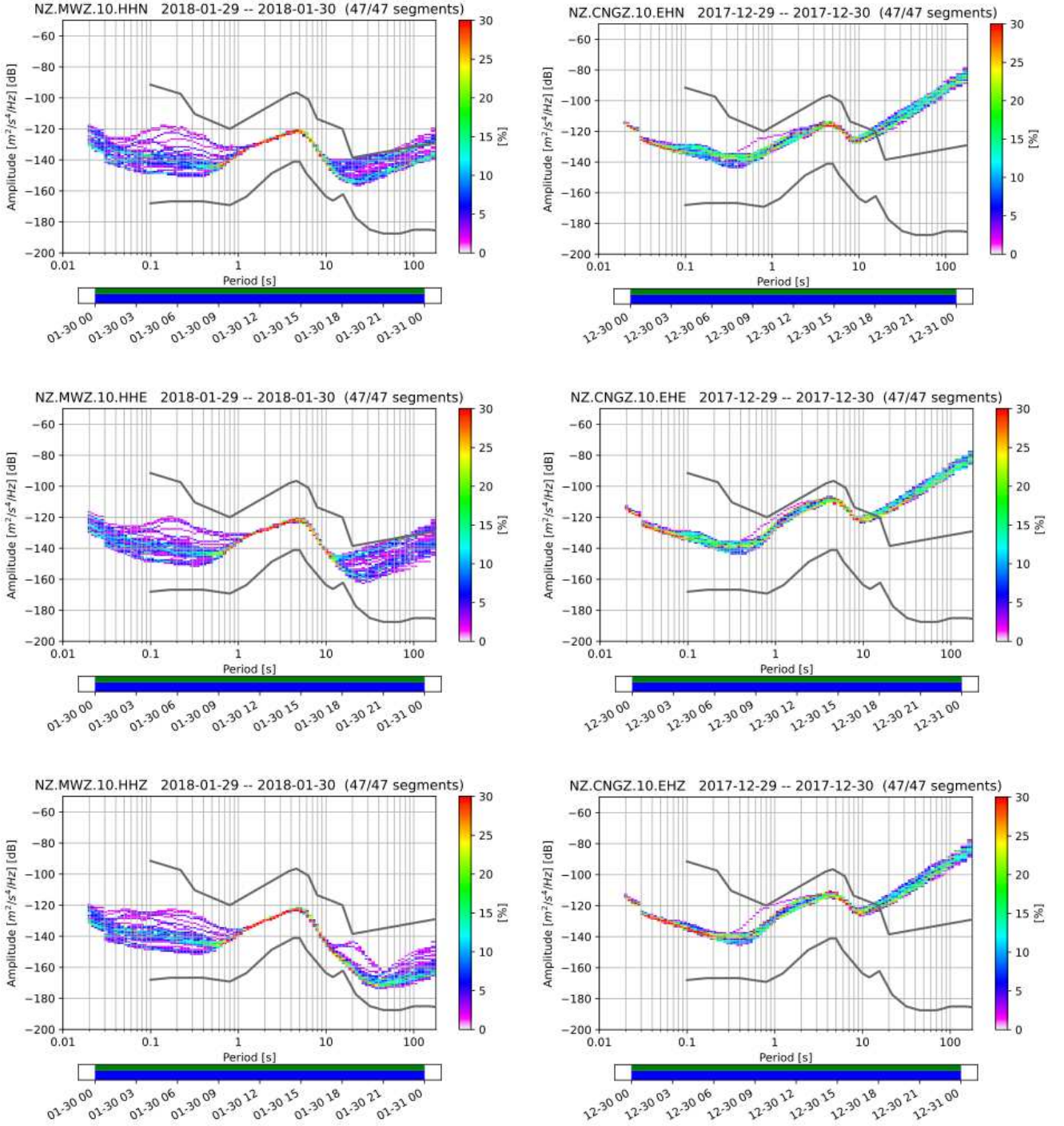


Figure 2.6: Probabilistic power spectral density (PSD) distributions showing typical background seismic noise levels for stations from the permanent network. Left column shows 24-hr window from a broadband station (MWZ). Right column shows 24-hr window from a short-period station (CNGZ). PSD plots are plotted following the method of *McNamara (2004)*, for frequencies ranging from 0.005–50 Hz. Dark grey lines show the typical low and high-noise level for broadband stations from *Peterson (1993)*. Top row shows north component; middle row shows east component; bottom row shows vertical component. Blue line shows observation period; green line represents data continuity (i.e., stations recorded continuously during this 24-hr window).

2.3.4 Geodetic instrumentation

SSEs are detected geodetically, typically using global navigation satellite systems (GNSS), though more recently with seafloor pressure gauge data (e.g., [Woods et al., 2024](#)). Since October 2002, GeoNet has operated a permanent network of continuously operating Global Navigation Satellite System stations (Figure 2.4; [Gale et al., 2015](#)), with the first SSE detected in October 2002 ([Douglas et al., 2005](#)). Since then, numerous SSE catalogues for the northern Hikurangi margin have been published (e.g., [Wallace, 2020](#); [Perez-Silva et al., 2025](#); [Michel et al., 2025](#)). Geodetic daily displacement data were used in this thesis to constrain SSEs that occurred during the deployment of the Gisborne network.

2.4 Data availability

Raw waveform data from permanent seismograph and GNSS stations are available from GeoNet, accessed at <https://www.geonet.org.nz/data/tools/FDSN> (last accessed and downloaded June 2025; [GNS Science, 2019, 2021](#)). Waveforms and metadata are available through the facilities of the EarthScope Consortium. These services are funded through the National Science Foundation’s Seismological Facility for the Advancement of Geoscience (SAGE) Award under Cooperative Agreement EAR-1724509. Data from the temporary seismograph network are hosted at https://doi.org/10.7914/SN/3C_2017 ([Bell et al., 2017](#)).

CHAPTER 3

A new earthquake catalogue, focal mechanisms, and 1-D velocity model for the northern Hikurangi margin

Some of the contents of this chapter are published in JGR: Solid Earth.

Woodward, A., I.D. Bastow, R. Bell, L. Wallace, K. Jacobs, S. Henrys, et al. (2026). Seismological characterization of Northern Hikurangi Margin Slow Slip Regions Associated with Normal Faults, Seamounts, and Seeps. *Journal of Geophysical Research: Solid Earth*, **131**, e2025JB032916, [doi:10.1029/2025JB032916](https://doi.org/10.1029/2025JB032916).

The datasets associated with this publication are available through Zenodo: [Woodward et al. \(2025\)](https://doi.org/10.5281/ZENODO.15058318). Local Earthquake and Focal Mechanism Catalogue for the Northern Hikurangi Margin December 2017–October 2018 [Dataset], [doi:10.5281/ZENODO.15058318](https://doi.org/10.5281/ZENODO.15058318).

3.1 Introduction

The shallow part of the northern Hikurangi margin, where the most frequent SSEs occur, has been constrained with dense, offshore networks (e.g., [Yarce et al., 2019](#); [Mochizuki et al., 2021](#)). The permanent network is too sparse to provide comparable constraints for the onshore slow slip region. We use a temporary land-based network of 49 seismograph stations, deployed at 2 km spacing from December 2017 to October 2018 at the surface directly above the downdip portion of the northern Hikurangi slow slip source region, to derive a new earthquake catalogue, a 1-D velocity model, and focal mechanisms for the deeper onshore forearc.

The network was deployed directly above the deeper portion of the plate interface (>15 km

depth), downdip of the largest and most frequent northern Hikurangi SSEs. We construct a new high-resolution earthquake and focal mechanism catalogue, derive a 1-D velocity model for the study region, and perform time-dependent inversions of GNSS time series to characterise small SSEs that occurred during the deployment. We use these datasets to characterise the distribution and style of seismicity in the upper plate, subducting slab, and plate interface.

In Chapter 4, we use the catalogues derived in this chapter to investigate how seismicity rates and patterns vary during SSEs, including events that may rupture the deeper (15–30 km) portion of the interface in the transition between the East Coast and Kaimanawa SSEs, and use focal mechanism solutions to infer the architecture of fault and fracture networks and their role in facilitating fluid migration from the slab mantle to the plate interface, upper plate, and surface seeps. This allows us to assess how the rough subducting plate interface at northern Hikurangi influences forearc dewatering and the conditions required for SSE genesis.

3.1.1 Existing earthquake detection

Routine earthquake detection is conducted by GeoNet, the national hazards monitoring service for Aotearoa New Zealand, using the permanent network of seismometers. As discussed in Chapter 2, the GeoNet catalogue includes many earthquakes with fixed depths, limiting their use in subsequent seismological techniques such as local earthquake tomography. Furthermore, the national network has ≈ 50 km station spacing, limiting the detection of low magnitude earthquakes.

The longest record of seismicity in this region is published by GeoNet ([Petersen et al., 2011](#)), using the permanent national network of short-period and broadband seismograph stations (Figure 2.4). Earthquake detection, location, and magnitude calculation are routinely carried out and published in near real time by GeoNet. The GeoNet catalogue for Te Tai Rāwhiti/Gisborne, however, contains analyst-fixed depths that lack associated depth uncertainties; 20% of earthquakes lack S-wave picks, which are especially critical for improving depth certainty ([Warren-Smith et al., 2024](#)). GeoNet and the International Seismological Centre routinely determine focal mechanisms, but typically only for earthquakes with $M_W > 4.5$

([Lentas, 2018](#); [Lentas et al., 2019](#); [GNS Science, 2021](#)). This means there are only a handful of focal mechanisms determined during our study period. As the largest magnitude earthquakes typically occur on the plate interface, there are a lack of constraints on intraslab focal mechanisms, and consequently, rupture mechanisms of the intraslab earthquakes are poorly understood.

3.2 Deriving a new earthquake catalogue

We first manually inspect the continuous waveforms to identify earthquakes. Because this thesis and its associated publications are the first use of the temporary network for local earthquake analysis, it was important to manually inspect waveforms to understand possible issues with stations as detailed in Chapter 2. Manual inspection for earthquakes was useful for understanding the frequency of earthquakes and to notice swarms of earthquakes.

We then detect earthquakes using the EQTransformer picking code ([Mousavi et al., 2020](#)). Phases are then associated using GaMMA ([Zhu, 2022](#)). A subset of these earthquakes are manually inspected, and compared to a smaller catalogue of manually-derived earthquake picks and the GeoNet catalogue. Earthquake hypocentres and origin times are determined using NonLinLoc, with a range of 1-D and 3-D velocity models. We use the catalogue to derive a new 1-D velocity model for the region using VELEST ([Kissling et al., 1994](#)), centred on the Gisborne network. We then calculate earthquake magnitudes and determine focal mechanisms using the FOCMEC software ([Snoke, 2003](#)). Our final catalogue contains twice as many earthquakes as the GeoNet catalogue, with additional constraints on hypocentre uncertainty, and a magnitude of completeness improved from 2.0 to 1.7.

3.2.1 Manual earthquake picking

Continuous seismogram data were Butterworth bandpass filtered with corner frequencies of 1 and 15 Hz. The vertical components of the continuous data were viewed in 30-minute windows with 5-minute overlaps in groups of eight stations. Where an earthquake arrival was observed, all three components of the waveform data were extracted 30 seconds before the arrival and 250 seconds after. The three components were then viewed together in groups of

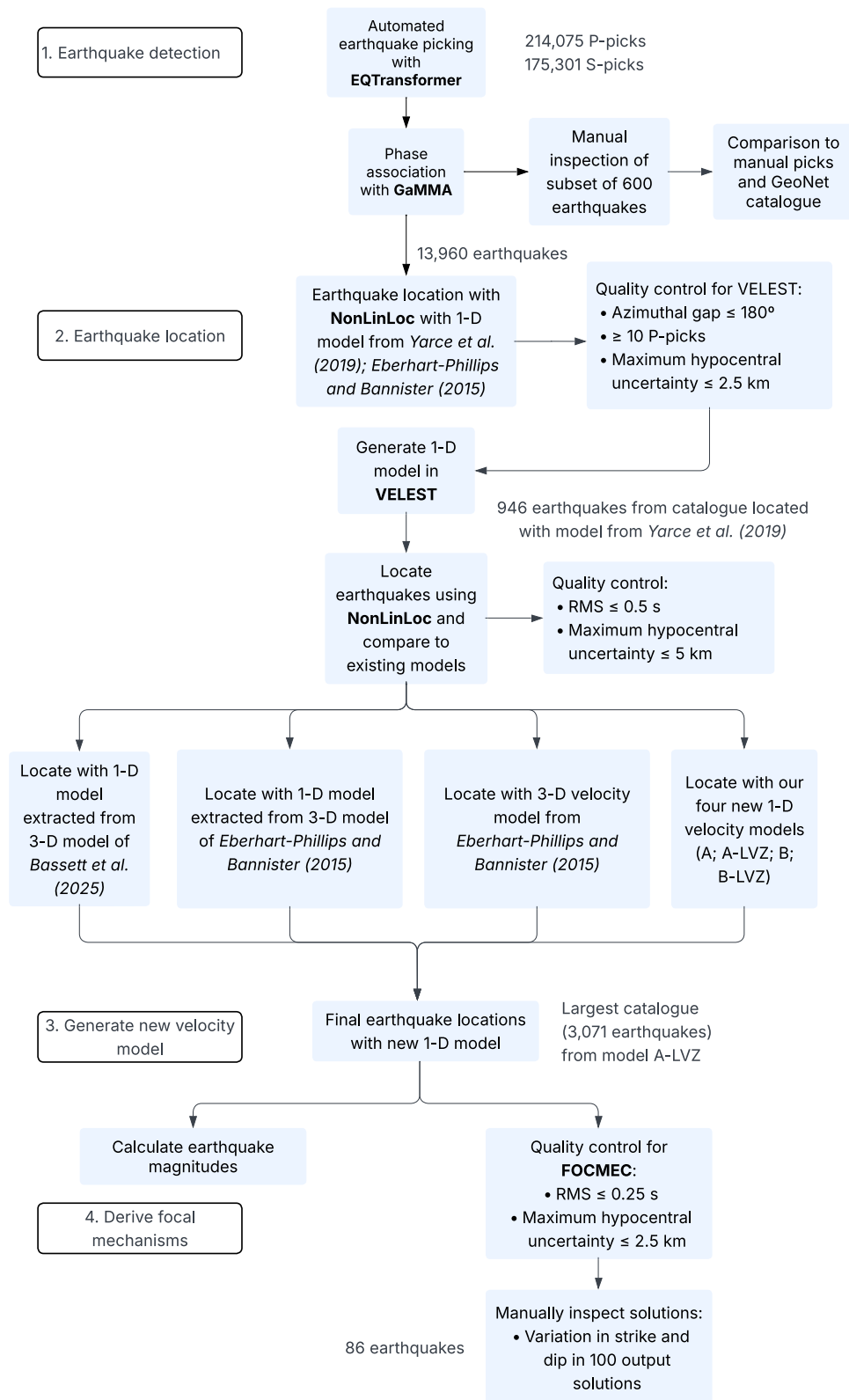


Figure 3.1: Workflow in this chapter.

three stations. P-waves were picked on the vertical components; S-waves were picked on the horizontal components.

The arrival times of P- and S-phase picks are assigned quality ratings, corresponding to uncertainties in arrival time of 0.05 s, 0.1 s, or 0.5 s. Quality ratings are subjective, but helpful for general understanding of data quality. The pick uncertainty is incorporated into the location error calculation in the location inversion procedure. Uncertainty in pick time increases with noisier data or inconsistent time window and amplitude scaling when visualising waveforms ([Diehl and Kissling, 2007](#)). S-wave arrival times typically have greater uncertainty than P-wave arrivals because S-wave arrivals often occur in the coda of previously arriving phases.

Most earthquakes are classified as having an uncertainty of 0.1 s. Impulsive arrival times with excellent signal-to-noise ratios are assigned a quality rating corresponding to an uncertainty of 0.05 s. For poor quality picks, the assigned quality rating corresponds to an uncertainty of 0.5 s. This enables the inclusion of low-quality picks, which are often critical for constraining earthquake depths, whilst incorporating the larger error in pick time in subsequent calculations. The uncertainties in arrival times (0.05 s, 0.1 s, and 0.5 s) are incorporated in the location error calculation.

3.2.2 Automated earthquake detection

Human analysts systematically pick arrival times earlier than automated picks and when multiple people are picking arrival times for the same earthquake catalogue, there can be inconsistencies in interpretation of arrival time ([Diehl and Kissling, 2007](#); [Ross et al., 2018](#); [Münchmeyer et al., 2022](#)). Automated detection methods are therefore useful for creating consistent catalogues.

Automated earthquake pickers have been compared extensively ([Münchmeyer et al., 2022](#)). We deploy an automated phase picker through SeisBench, which allows for the use of multiple pickers and datasets through one platform ([Woollam et al., 2022](#)). When selecting pickers, it is important to consider the training data, the picking algorithm, and the spatial constraints of the seismograph network. There is no substitute, however, for manually inspecting waveforms to ensure that detections are indeed earthquakes (discussed further in Section 3.6) and that

the pick times are aligned with where a human analyst might reasonably pick the arrival times of P-wave and S-wave phases. The purpose of this study is to use seismicity to provide tectonic and geologic constraints. Therefore, we use parameters that result in high-quality hypocentre determinations, rather than simply the largest catalogue possible. We therefore impose strict criteria (e.g., minimum number of picks or maximum travel-time residuals) and explore the impact of velocity models on seismicity distributions to ensure that our final interpretations are robust.

3.2.3 Earthquake detection and phase picking

EQTransformer is a deep learning-based earthquake detection and phase picking tool, chosen due to being open source, well-documented, and well-suited to the spatial scale of our study region ([Mousavi et al., 2020](#)). It is based on a convolutional neural network with a hierarchical attention mechanism. The EQTransformer model is trained on the Stanford EArthquake Dataset (STEAD), which is a global dataset of local earthquake signals and noise ([Mousavi et al., 2019](#)). STEAD includes 1.2 million, 60-second waveforms, of which $\approx 100,000$ are noise and the remaining are earthquake signals that sample a range of tectonic settings. The earthquakes have a global distribution, including some located in Aotearoa New Zealand, though this is not a requirement for successful use of the model. [Mousavi et al. \(2020\)](#) demonstrated that EQTransformer works well even in regions that are not represented in the training data. The remaining waveforms are of local earthquakes at distances < 350 km from the seismometer at which they were recorded. The waveforms are manually identified and labelled. EQTransformer, given its use of local earthquake training data, is well-suited to local earthquake studies – particularly at station distances of less than 150 km from earthquakes ([Mousavi et al., 2020](#)) – and is therefore well-suited to our localised study. EQTransformer generates probabilities of earthquake detections, P-wave arrivals, and S-wave arrivals as a function of time when applied to continuous waveform data. We implement EQTransformer through SeisBench ([Woollam et al., 2022](#)).

We set probability thresholds of 0.05 for detection of P- and S-wave phases and 0.1 for detection of earthquakes following [Merry \(2022\)](#), with 55 s overlaps of the continuous wave-

forms ([Pita-Slim et al., 2023](#)), with the aim of detecting as many earthquakes as possible. We subsequently apply quality control criteria. We benchmarked EQTransformer-detected earthquakes against the manually-picked and GeoNet catalogues ([Petersen et al., 2011](#)), then manually inspected the earthquakes that EQTransformer failed to detect. Missed detections were invariably from poor signal-to-noise ratio windows, or from earthquakes that were only observable on more distant permanent stations outside our study area. We manually reviewed a subset of EQTransformer-detected events to quantify the proportion of false detections and inaccurate picks. After analysing 600 automatically picked and associated events, all 600 were deemed real earthquakes. These 600 earthquakes accounted for ~5% of the initial catalogue. Earthquakes were inspected during times when different sampling rates were used and after each maintenance period.

3.2.4 Phase association with GaMMA

EQTransformer identifies potential earthquakes on a trace-by-trace basis. Once the initial picks and detections catalogue has been generated, each pick must be associated with an event, thus creating a catalogue of detected events with multiple phase arrivals for each event. We use the Gaussian Mixture Model Associator (GaMMA) to associate phases ([Zhu et al., 2022](#)). GaMMA provides associated phase picks for each event with a preliminary event time and location.

GaMMA uses a Density-Based Spatial Clustering of Applications with Noise (DBSCAN) algorithm to associate multiple phases as one event, using hyperbolic move-out times of arrival times across the network with a uniform velocity model. We test the effects of different GaMMA parameters, as outlined in Table [3.1](#).

For events with large numbers of picks on stations across the network, which are separated by up to 180 km, GaMMA has a tendency to separate picks that are likely from a single earthquake into two earthquakes, depending on the parameters chosen. Relaxing the parameters to allow greater differences in arrival times between different stations runs the risk of merging multiple earthquakes that truly occur in quick succession into a single earthquake. Earthquakes occurring in quick succession may be earthquake swarms.

Earthquake swarms lack a formal definition, though they are generally accepted as above-

Table 3.1: GaMMA phase association parameters used in this study.

Parameter	Value	Comments
DBSCAN epsilon	10.0 s	Maximum distance between any two samples where they can still be considered in the neighbourhood of each other i.e., the maximum time between two picks where they can be considered neighbours.
DBSCAN minimum samples	3	Number of samples that must be in the neighbourhood of a single sample, for that single sample to be considered a core point.
Oversampling factor	10	Initial number of clusters is given by the (number of picks / number of stations) $\times$ oversampling factor
Min. picks per event	8	
P-wave picks per event	≥ 4	
S-wave picks per event	≥ 2	
Maximum σ_{11}	2.0 s	Maximum phase time residual.
V_P at 14–24 km depth	5.973 km/s	From Yarce et al. (2019) velocity model.
V_S at 14–24 km depth	3.388 km/s	From Yarce et al. (2019) velocity model.

Table 3.2: GaMMA phase association parameter testing.

Run	DBSCAN EPS	Covar. prior	Min. picks	Min. P	Min. S	Max. σ_{11}	No. events
1	15	20, 20	8	4	2	2	13 960
2	10	20, 20	8	4	2	2	12 226
3	10	20, 20	6	2	1	2	21 044
4	25	20, 20	8	4	2	2	14 715
5	20	20, 20	8	4	2	2	14 559
6	15	20, 20	8	4	2	5	14 559
7	5	20, 20	8	4	2	5	6 497
8	15	20, 20	8	4	2	3	14 818
9	15	20, 20	8	4	2	4	15 304
10	25	20, 20	8	4	2	5	16 364
11	15	5, 5	8	4	2	2	15 385
12	15	10, 10	8	4	2	2	14 770
13	15	15, 15	8	4	2	2	14 319
14	5	20, 20	8	4	2	2	6 497

normal rates of seismicity without a mainshock ([Nishikawa et al., 2021](#)). There are numerous observations of earthquake swarms at the Hikurangi margin (e.g., [Bartlow et al., 2014](#); [Nishikawa et al., 2021](#)) and at other subduction zones (e.g., [Nishikawa et al., 2023](#)), associated with the slow slip cycle, thus we want to retain earthquake swarms should they exist. In order to retain potential earthquake swarms, while limiting the possibility of erroneously splitting single earthquakes into multiple events, we initially require that the maximum difference in arrival times for picks to be considered neighbours is 15 s based on the likely range of travel-times across the study region for one earthquake. Subsequent quality control measures then retain only high quality hypocentres with low RMS misfit (<0.5 s) and azimuthal gap (270°). Earthquakes that have been erroneously split into multiple events were removed by this processing stage. The remaining earthquakes with origin times in quick succession were manually inspected.

Table 3.2 show parameters used in GaMMA and the number of earthquakes in the resultant catalogue. Parameters for run 1 are chosen for the final catalogue based on manual inspection of the phase picks during swarms. Run 1 parameters produce the highest number of earthquakes that survive the post-association, quality control phase and retains earthquake swarms, discussed further in Chapter 4. After retaining only high-quality hypocentres, 54 earthquakes are observed within 30 s of the previous earthquake, with depth differences less than 10 km and latitude/longitude differences of less than 0.1° . These events are manually inspected and are identified as distinct events, associated with likely swarms of earthquakes, discussed further in Chapter 4. We observe a recurrence of seismicity in 2018 on 3 March, 4 May, 2 August, and 3 August. These earthquakes are manually verified and show an unusually large number (>8) of earthquakes in under 30 minutes, suggesting earthquake swarms.

3.2.5 Earthquake location with NonLinLoc

Earthquake locations are determined using observed phase arrival times and predicted travel times, for an assumed elastic wavespeed model, in order to recover the hypocentre and origin time. Commonly, a probability density function is determined, which quantifies the misfit between observed and predicted arrival times and their associated uncertainties over a search

volume. Constraining earthquake hypocentres is an inverse problem, whereby the location and origin time of an earthquake is determined using the arrival times of different seismic phases at different seismic stations picked from seismograms. The choice of velocity model, which is used to determine ray paths and subsequent travel times, is therefore critical.

During location, earthquakes are treated as point sources; however, in reality, earthquakes can rupture faults tens to hundreds of kilometres long. The earthquake hypocentre is therefore defined as the position (x, y, z) from which seismic energy initially radiates. The P-wave velocity always exceeds the rupture velocity, and can therefore be used to determine from where the energy first radiated, regardless of earthquake magnitude, which is a function of rupture length or duration.

At n seismic stations, located at $\mathbf{x}_i = (x_i, y_i, z_i)$, the arrival time of a seismic phase, d'_i , is picked. The arrival time is dependent on the origin time, t , and the travel time between the hypocentre and the station, $T(\mathbf{x}, \mathbf{x}_i)$. The arrival time is therefore expressed as:

$$d'_i = T(\mathbf{x}, \mathbf{x}_i) + t. \quad (3.1)$$

A velocity model is assumed, such as those computed from ak135 or iasp91 travel-time tables. Using the assumed velocity model, and a model vector, $\mathbf{m} = (x, y, z, t)$, arrival times are computed:

$$\mathbf{d} = A(\mathbf{m}), \quad (3.2)$$

where $\mathbf{d}$ represents the data vector, A represents the arrival time, and $\mathbf{m}$ represents the model vector.

For the observed arrival times, we seek the model that best fits them. Let the starting model be denoted by $\mathbf{m}^\circ$, which is iteratively updated such that the predicted data lie closer to the observed data. This results in

$$m_j = m_j^\circ + \Delta m_j, \quad (3.3)$$

where the starting model, $\mathbf{m}_j^\circ$ is updated by Δm_j , to give an updated model vector, m_j .

The problem is non-linear, but it can be linearised using a Taylor series expansion:

$$d_i \approx d_i^\circ + \sum_j \left. \frac{\delta d_i}{\delta m_j} \right|_{m^\circ} \Delta m_j, \quad (3.4)$$

where d_i is the arrival time for a given model vector, d_i° is the starting model prediction, the sum is the Taylor series expansion about the starting model, $\mathbf{m}^\circ$, and Δm_j is the change in the starting model. We aim to minimise the difference between the observed and predicted data, represented by

$$\Delta d_i^\circ = d_i' - d_i^\circ. \quad (3.5)$$

The partial derivatives can be represented by G_{ij} , such that the difference between observed and predicted arrival times can be simplified to

$$\Delta \mathbf{d} = G \Delta \mathbf{m}. \quad (3.6)$$

This vector-matrix equation represents a series of simultaneous linear equations.

NonLinLoc is a non-linear location software package used for calculating travel-time grids and constraining earthquake hypocentres using a probabilistic grid-search method. Non-linear methods, such as NonLinLoc, are useful because the misfit function can be calculated over the entire parameter space, rather than relying on linear processes that iterate from a first guess, risking falling into a local minima. NonLinLoc generates a travel-time grid, for points within a 3-D space, such that a travel time is calculated between every possible source location and receiver. The travel time is calculated based on an input 1-D velocity model. Varying the 1-D velocity model depths and wavespeeds changes the calculated travel times, thus adjusting the hypocentre. It is therefore important to consider the abilities and limitations of different velocity models.

Calculating travel times from a laterally homogeneous or 1-D velocity model, where velocity

varies only with depth, is computationally less expensive and requires less storage than a 3-D model. Furthermore, where 3-D models exist (e.g., [Eberhart-Phillips et al., 2020](#)), the resolution of such models is comparable to a 1-D velocity model, given the highly localised nature of this study. A higher resolution velocity model is derived in Chapter 5, and earthquakes are relocated with this model. Regardless of dimensions, all velocity models result in interpolation or smoothing compared to the true Earth. It is therefore more robust to consider multiple velocity models, to test the sensitivity of the earthquake hypocentres to errors in the velocity models. Here we locate events using models from [Yarce et al. \(2019\)](#) and iasp91, which is used in the GeoNet catalogue.

The velocity models are used to compute travel-time grids for the first arrival signal at a receiver from a given point in the grid. This reduces the earthquake location problem to a much simpler lookup table. NonLinLoc uses the Eikonal method of [Podvin and Lecomte \(1991\)](#) to determine the travel time for one node in the grid to every other point in the model, including at the surface at station locations. The method aims to minimise the difference between the observed and predicted travel times. A probability density function is used:

$$pdf(\mathbf{x}) = \left[\sum_{a,b} \frac{1}{\sqrt{\sigma_a^2 + \sigma_b^2}} \exp \left(- \frac{[T_a^o - T_b^o] - [TT_a^c(\mathbf{x}) - TT_b^c(\mathbf{x})]^2}{\sigma_a^2 + \sigma_b^2} \right) \right]^N, \quad (3.7)$$

where $T_{a,b}^o$ represents a pair of observed arrival times for the same event, $TT_{a,b}^c$ represents calculated travel-times for observations a and b , and N is the total number of observations. σ_a and σ_b are the errors in pick time and travel-time. The probability density function is a function of differential travel time and therefore only depends on longitude, latitude, and depth, not on absolute travel times.

To locate events, an Oct-Tree importance-sampling method is used. Oct-Tree algorithms recursively determine which cell in search volume has the highest probability, then subdivides that cell into a further eight cells and repeats the probability calculation ([Lomax et al., 2009](#)). Initially, a coarse grid is used, which recursively gets finer towards the highest probability regions. Oct-Tree algorithms can map multiple peaks in probability density function, preventing the location getting stuck in local minima. NonLinLoc parameters used are outlined in Table

3.3.

Table 3.3: NonLinLoc parameters used in this study.

Parameter and Description	Value/Setting
Transverse Mercator projection	Centred on 178.00°E, -38.42°S
V_P/V_S (to determine V_S model from V_P model)	1.730
Number of initial X cells	10
Number of initial Y cells	10
Number of initial Z cells	4
Number of Oct-Tree samples	200,000
Station distribution weighting	Yes
Minimum node size	0.1 km
Search volume	$175 \times 200 \times 200$ km

3.3 A new 1-D velocity model centred on the Gisborne network

To improve earthquake locations, we derive a 1-D velocity model centred on the Gisborne network. The GeoNet earthquake catalogue is derived from earthquakes located with the IASP-91 velocity model, or a coarse 3-D model. We derive a 1-D velocity model suited to our highly localised study region. We then use this velocity model in Chapter 5 as the starting model for a staggered inversion for a 3-D V_P and V_P/V_S model, as recommended by [Kissling et al. \(1994\)](#).

VELEST simultaneously inverts for hypocentres and 1-D velocity structure using an iterative least-squares matrix inversion that aims to minimize the RMS misfit between observed and calculated travel times ([Kissling et al., 1994](#)). VELEST provides an effective mean velocity model for the region sampled by the raypaths. Regions with denser ray coverage are therefore better represented in the output velocity model. VELEST was developed to create 1-D velocity models for earthquake location and to provide initial reference models for seismic tomography inversion schemes, using earthquake arrival time data.

Initial hypocentres were determined using a 1-D velocity model derived from the 3-D velocity model of [Yarce et al. \(2021\)](#) at the co-ordinates closest to the temporary network using Non-LinLoc software ([Lomax et al., 2000, 2001, 2023](#)). Observed arrival times can be inverted to

obtain a 1-D velocity model using the VELEST programme ([Kissling et al., 1994](#)). We randomly generated 200 starting velocity models between the bounds of existing velocity models for the region (e.g., [Eberhart-Phillips and Fry, 2017](#); [Yarce et al., 2019, 2021](#)). Particularly at shallow depths, velocity structure from previous studies varies significantly, possibly due to variations in onshore and offshore coverage. For example, at a depth of 0 km, previously published velocity models have a V_P between 1.5 and 4.3 km/s ([Eberhart-Phillips and Fry, 2017](#); [Bassett et al., 2025](#)).

The uppermost layer was 2 km thick, from sea level to 2 km elevation. Layers were 1 km thick from 0–12 km depth, 2 km thick from 12–32 km depth, and 4 km thick from 32–50 km depth. Thicker layers were used at depths 32–50 km because these deeper layers were sampled by fewer earthquake raypaths. The starting velocity for the shallowest layer (sea level to 2 km elevation) was a randomly selected V_P value between 1.3 and 5.3 km/s, based on the range of previously published values (Figure 3.2). From this initial value, velocities could only increase with depth. We imposed a 20% chance that velocity could increase with respect to the layer above, thereby introducing variability in layer thickness between starting models. The 20% chance was chosen so that the output 200 starting models evenly sampled the bounds of the starting model, shown in grey in Figure 3.2. A lower percentage prevented the starting models from including the higher velocity values within each range for a given depth, while a higher percentage resulted in many of the velocity models reaching the highest velocity values at a relatively shallow depth.

VELEST parametrisation

We inverted P-wave travel times for earthquakes with azimuthal gaps $\leq 180^\circ$, ≥ 10 P-phase picks, a maximum hypocentral uncertainty ≤ 2.5 km, and an RMS residual of ≤ 0.4 s, located within the bounds of the seismograph stations used in the study. These criteria were met by 956 earthquakes out of the 3,041 earthquakes. We first implemented VELEST in single-event mode, where velocity models were held effectively fixed through high damping and inverted only for hypocentres. This limits bias towards the model used to locate the earthquakes initially. Earthquakes that had an updated RMS residual > 1.0 s or azimuthal gap $> 180^\circ$ after single

event mode were removed from subsequent analysis; 10 such earthquakes were removed, leaving 946 suitable earthquakes. We then implemented an iterative inversion for hypocentres, velocity model, and static station corrections.

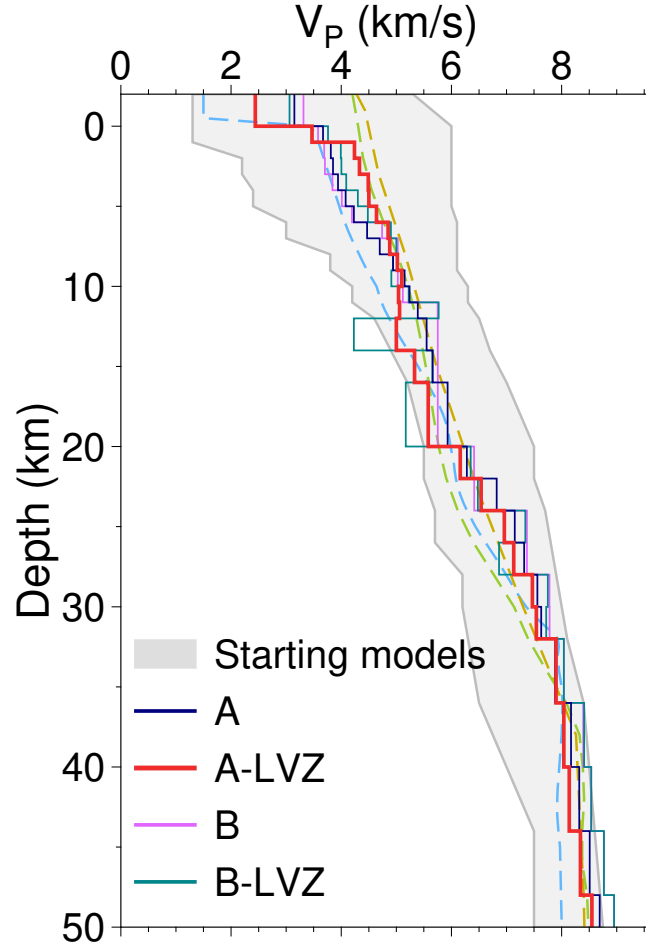


Figure 3.2: P-wave velocity models from this study. Gray shaded region: range of starting models used in VELEST. Yellow dashed line: velocity model extracted from *Eberhart-Phillips and Bannister (2015)*. Blue dashed line: velocity model extracted from *Bassett et al. (2025)*. Green dashed line: velocity model extracted from *Yarce et al. (2021)*. Red line (A-LVZ): final P-wave velocity model used in this study.

We inverted the 200 starting models and tested origin time, hypocentre, velocity model, and static station correction damping parameters. We initially inverted for hypocentres only by fixing the (highly damped) velocity model. This allowed each starting velocity model to reach a minimum before the hypocentres and velocities could vary jointly. We then ran VELEST in simultaneous inversion mode, varying damping parameters for static station corrections, velocities, origin times, and hypocentres. Optimal damping parameters were chosen by inspecting the trade-off between RMS travel-time misfit versus damping term. We used a velocity damping term of 1.0, and station static correction, hypocentre, and origin time

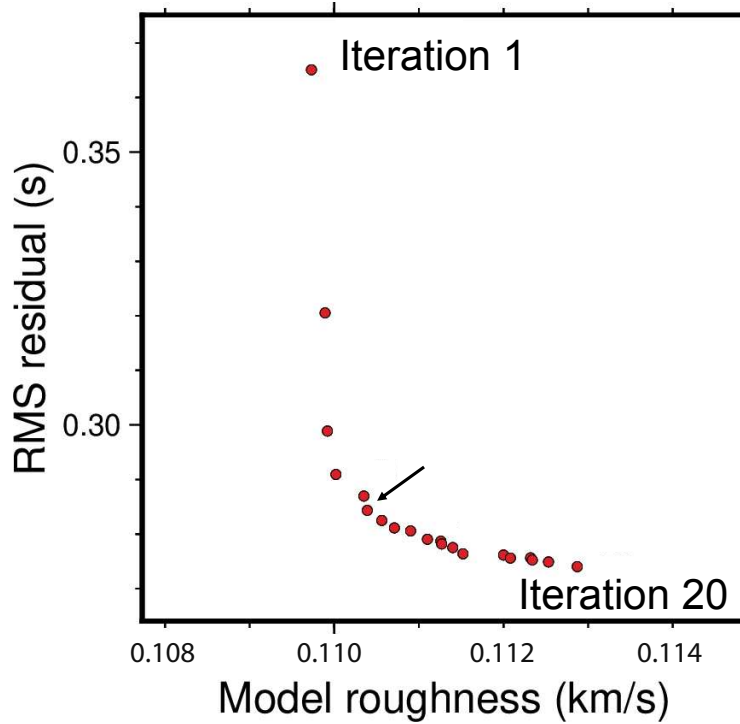


Figure 3.3: Example trade-off curve of RMS residual (s) versus model roughness (mean velocity contrast between successive layers) for a starting model used in VELEST. 20 iterations were run and in this example the fifth iteration was chosen (arrow).

damping of 0.1. We allowed each of the 200 velocity models to run up to 20 iterations, and calculated the model roughness at each iteration. The mean RMS travel-time misfit for the hypocentres used in the inversion was 0.29 s. After inverting the 200 starting velocity models, we selected output models with a mean RMS misfit of the hypocentres < 0.29 s. We then selected the iteration of the knee of the curve as the final output velocity model for each of the 200 starting models that show a decrease in RMS misfit (Figure 3.3).

We generated four 1-D velocity models (Figure 3.2), defined as model A: the output mean velocity model from 200 inversions; model A-LVZ: the output mean velocity model from 200 inversions, with low velocity zones allowed at every iteration; model B: the output mean velocity model from 200 inversions, which is then used as the starting model for one more iteration; model B-LVZ: the output mean velocity model from 200 inversions, which is then used as the starting model for one more iteration with low velocity zones allowed.

We located all 13,960 earthquakes using the four generated 1-D P-wave velocity models, using NonLinLoc, with $V_P/V_S = 1.73$, based on Wadati plot analysis of the new catalogue (Figure

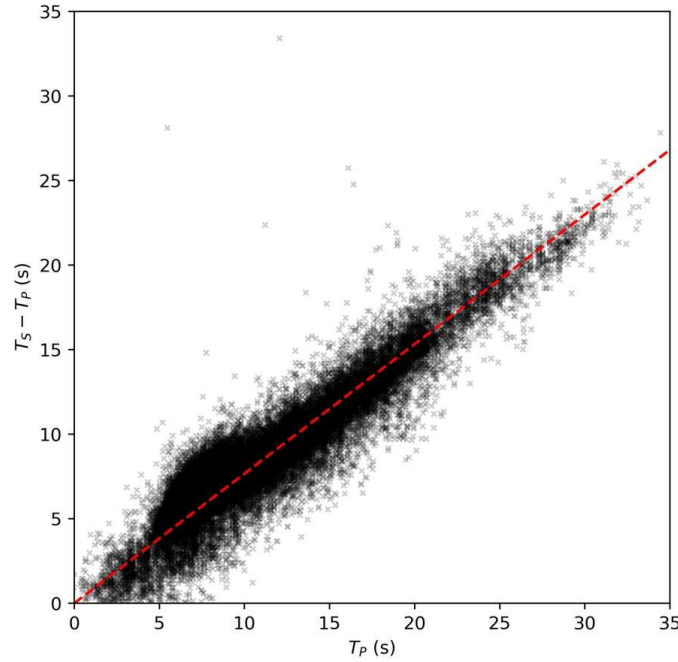


Figure 3.4: Wadati diagram of differential travel-times from the catalogue. Red line: least-squares linear regression to determine the gradient of the differential S-P travel times against the P-wave travel times.

3.4; *Wadati and Oki, 1933*). The A-LVZ model results in the greatest number of well-located earthquakes (RMS misfit < 0.5 s, maximum hypocentral error < 5 km, and lowest median misfit), when compared with our other 1-D velocity models and previously published 1-D and 3-D velocity models. A low velocity zone is observed at 10–14 km depth, consistent with the depth of the interpreted fluid-rich region at the plate interface (*Warren-Smith et al., 2019; Arai et al., 2020; Bangs et al., 2023*).

3.4 Relative relocation with hypoDD

Relative earthquake relocations are routinely used to improve the precision of earthquake locations. While relative relocation routines do not improve absolute earthquake locations, the relative locations of earthquakes can delineate structures such as faults, along which bands of seismicity may occur. We see numerous earthquakes beneath the Gisborne network due to its high station density.

We implement the double-difference algorithm of *Waldhauser (2000)* using the hypoDD

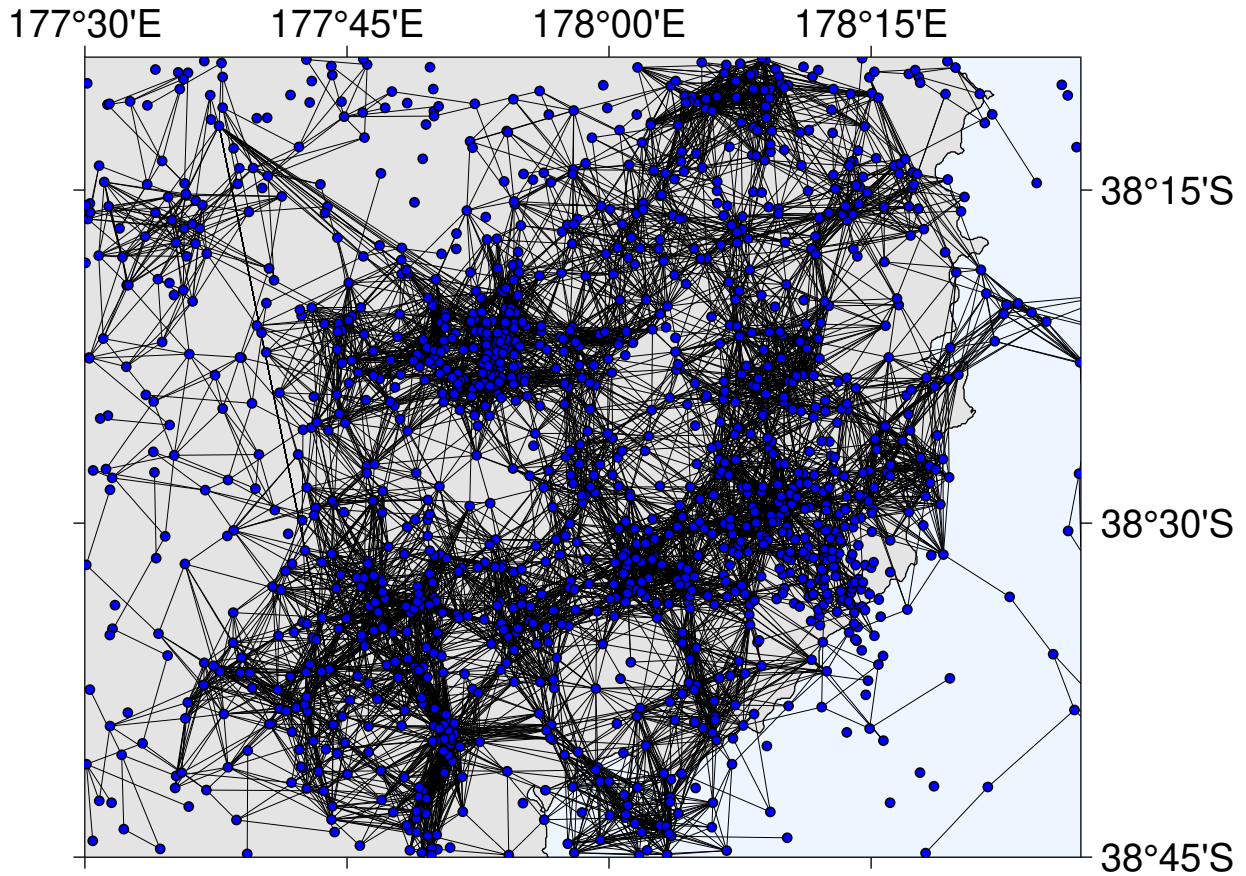


Figure 3.5: Blue circles: earthquakes. Black lines connect earthquake pairs of differential travel-times used in hypoDD.

software (version 1.3). If the hypocentral distance between two earthquakes is small relative to the distance between the pair of earthquakes and a seismometer, the difference in travel time between each earthquake and seismometer can be used to characterise the spatial offset between the two earthquakes. This improvement in relative earthquake locations can reveal fault structures. The localised nature of this study makes it well-suited to relative relocation methods, where there are dense clusters of earthquakes separated by distances of hundreds of metres to a few kilometres, recorded on seismometers up to 150 km away.

Parametrisation of hypoDD requires relatively uniform seismometer and earthquake distributions; thus, we focus relocation efforts on the Gisborne region, centred on the dense Gisborne network. Our focus is on improving imaging of faulting structures near Gisborne, and directly below the seismograph network where we investigate the plate interface architecture. We include only earthquakes with hypocentral uncertainty < 5 km and RMS misfit < 0.5 s, located near Gisborne. This results in 1,002 earthquakes with 36,597 P-picks and 31,617 S-picks.

We first calculate differential travel times for earthquake pairs (Figure 3.5). Weightings are derived from NonLinLoc outputs, as a function of travel time residual calculated during the initial location with our newly derived 1-D velocity model. The double difference residual is defined as:

$$dr_k^{ij} = \frac{\delta t_k^i}{\delta \mathbf{m}} \Delta \mathbf{m}^i - \frac{\delta t_k^j}{\delta \mathbf{m}} \Delta \mathbf{m}^j, \quad (3.8)$$

where

$$\Delta \mathbf{m}^{ij} = (\Delta dx^{ij}, \Delta dy^{ij}, \Delta dz^{ij}, \Delta d\tau^{ij}) \quad (3.9)$$

and i , and j are the two earthquakes, and k is the seismometer. The double difference residual is minimised using the weighted conjugate gradient least-squares method ([Paige and Saunders, 1982](#)). The vector difference between earthquake pairs is iteratively adjusted, and locations are updated after each iteration. If the vector adjustment at each iteration becomes unstable or too large, the adjustment is limited; this is controlled by the damping parameter. We tested damping parameters of 60, 70, 80, 90, and 100, within the recommended damping range for similar-sized problems ([Waldhauser, 2000](#)). The effect of damping is characterised by the condition number, CND, which is the ratio of the largest to smallest eigenvalue of the system matrix, i.e., a representation of the stability of the inversion. We select a damping parameter of 90, as this results in a CND between the recommended bounds (40–80) ([Waldhauser, 2000](#)) (Figure 3.6).

For any hypocentral pair, there are only 8 non-zero components of the design matrix. An initial solution is found using the initial locations and quality ratings from NonLinLoc. The locations, residuals, and partial derivatives are updated. The weights are recalculated using the misfit of data from the previous iteration. The data are iterated again, with the updated weightings. We perform 20 iterations and plot the RMS misfit, and mean change in origin time and hypocentre between each iteration. After iteration 6, subsequent iterations do not result in meaningful variation in hypocentre, origin time, or RMS misfit.

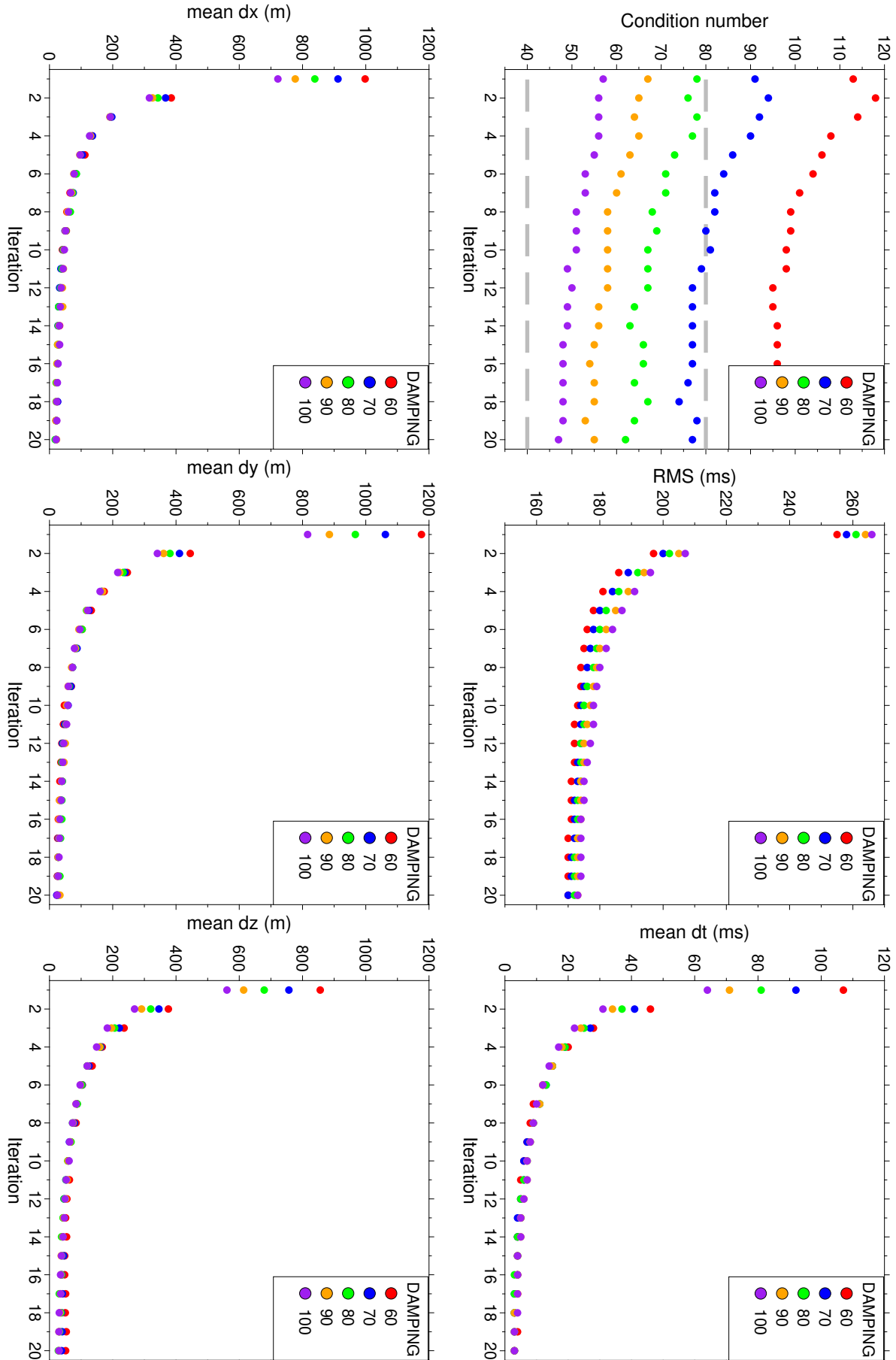


Figure 3.6: HypoDD damping parameter and iteration testing. Condition number: grey dashed lines denote ideal condition number value.

We calculate travel-time pairs for earthquakes separated by < 5 km, with a maximum cluster centroid-receiver distance < 150 km. Least-squares methods of relative relocations improve the relative locations, but do not provide uncertainty estimates for absolute locations.

Table 3.4: Parameters for earthquake clustering and relative travel time calculation

Parameter	Description	Value
MAXDIST	Maximum distance between earthquake pair and station	150 km
MAXSEP	Maximum hypocentral separation between earthquakes in a pair	5 km
MAXNGH	Maximum number of neighbours per event	20
MINLNK	Minimum number of links to define a neighbour	8
MINOBS	Minimum number of links per pair saved	10
MAXOBS	Maximum number of links per pair saved	20
DIST	Maximum distance between cluster centroid and seismometer	150 km
OBSCT	Minimum number of links in cluster	8
V_P/V_S	V_P/V_S used to determine S-wave model	1.73

Figure 3.7 shows the original and relocated earthquakes in map view and cross-section, centred on the temporary seismograph network. The relative relocations show tighter clustering of earthquake hypocentres, though do not form lineations that can be interpreted as faults. Previous studies have suggested that the plate interface consists of a fracture network (*Chalumeau et al., 2024*), which may be particularly complex in regions of subducted topography (*Wang and Bilek, 2011*). Offshore seismic reflection data shows colocated normal and thrust-type faults (*Barnes et al., 2020*). The lack of structure in the hypoDD results may be due to the presence of many small fractures, rather than resolvable km-scale faults.

3.5 Local magnitude calculation

Local magnitudes are determined by simulating a Wood-Anderson-type seismograph on a modern digital seismograph. The maximum amplitude for a given earthquake is compared with the reference amplitude for zero magnitude at the same epicentral distance.

Local magnitude, M_L , is defined as the maximum trace amplitude, A , recorded on a Wood-Anderson seismograph, relative to an earthquake of zero magnitude at the same epicentral distance, Δ from the seismograph. M_L is calculated using:

$$M_L = \log A(\Delta) - \log A_0(\Delta), \quad (3.10)$$

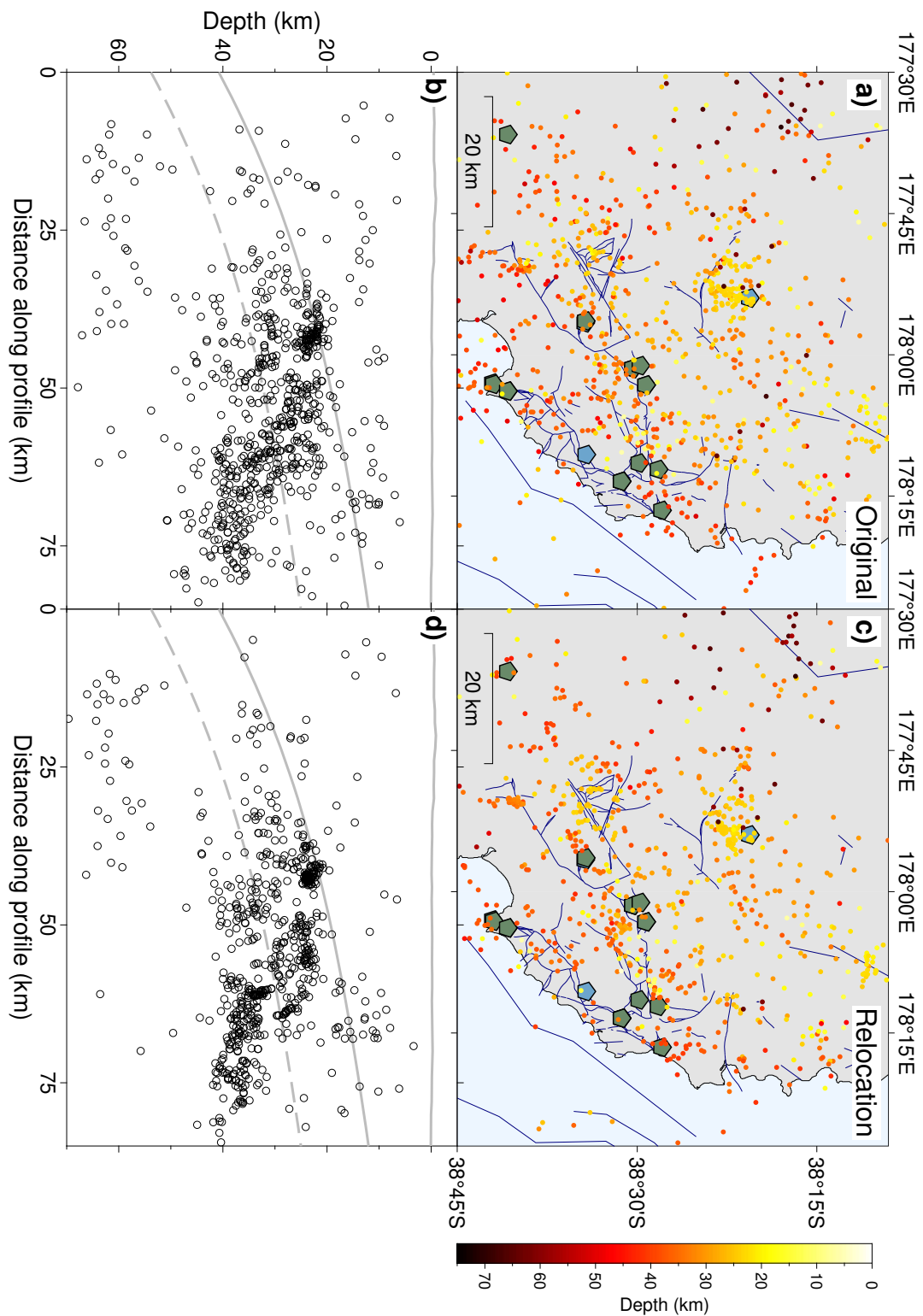


Figure 3.7: **a)** Map view of original earthquake locations from NonlinLoc. Pentagons: mud volcanoes (green) and seeps (blue). **b)** Cross-section view of original earthquake locations from NonlinLoc. Cross-section shown from north-west to south-east corner of map in panel a). **c)** Map view of relocated earthquakes from hypDD. **d)** Cross-section view of relocated earthquakes from hypDD.

where A is the maximum trace amplitude of the earthquake, and A_0 is the maximum trace amplitude of the reference event of zero magnitude at the same epicentral distance, Δ . The reference amplitude is region-specific to account for local attenuation. Calibrating local magnitude scales for the effects of local geology, such as attenuation through the mantle or fluid-rich regions, is critical for quantitatively analysing spatiotemporal seismicity distributions and magnitudes. A well-calibrated local magnitude scale is important for correctly calculating the cumulative frequency-magnitude distribution and associated b -value for the region. By accounting for local attenuation, the local magnitude scale improves the accuracy of M_L compared with broader regional magnitude scales (e.g., [Rhoades et al., 2020](#)). Here we use the magnitude scale derived by [Rhoades et al. \(2020\)](#):

$$M_L = \log A(\Delta) - 0.51 - 0.79 \times 10^{-3} \Delta + 1.67 \log(\Delta) + \delta H_{40} + S_i, \quad (3.11)$$

where S_i is the station correction term, and

$$\delta H_{40} = \begin{cases} 2.38 \times 10^{-3} H_{40}, & \text{for } h \geq 40 \\ 0, & \text{for } h < 40, \end{cases}$$

where h is earthquake depth and $H_{40} = h - 40$ ([Rhoades et al., 2020](#)).

The δH_{40} term corrects for uppermost mantle attenuation. Earthquakes at depths ≥ 40 km in Aotearoa New Zealand are typically located in the subducting slab of the Hikurangi margin, which is cold relative to ambient mantle temperatures ([Rhoades et al., 2020](#)). Compared to the background mantle, subducting slabs typically have higher wavespeeds and lower attenuation.

3.5.1 Station magnitude corrections

Station magnitude corrections account for uppermost variations in geologic structure directly beneath seismograph stations. For the permanent seismograph stations, we use the station correction values derived in [Rhoades et al. \(2020\)](#).

From the new catalogue, we calculate the maximum half peak-to-peak amplitudes (A_{W-A}).

To simulate a Wood-Anderson response, we correct using gain and normalisation frequency specific to each instrument. For each station-earthquake pair, we determine the maximum absolute peak-to-peak amplitude from the horizontal components of the seismograph. We then halve the peak-to-peak amplitude, to determine the maximum zero-to-peak amplitude.

The distance correction term, $\log(A_0)$, can be defined as:

$$-\log(A_0) = n \log\left(\frac{r}{17}\right) + K(r - 17) + 2, \quad (3.12)$$

where r is the hypocentral distance (km), n is a geometrical spreading parameter, and K is an S-wave attenuation parameter. The observed amplitude, A_{ijk} is thus defined as

$$\log(A_{ijk}) + 2 = -n \log\left(\frac{r_{ij}}{17}\right) - K(r_{ij} - 17) + M_{Li} - C_{jk}, \quad (3.13)$$

where i is the event index, j is the station index, and k is the component index (north-south or east-west). We invert for n , K , M_L , and C . For each station, two station correction terms are determined – one for each horizontal component. We find, however, that inverting for n and K is very sensitive to the earthquakes used. Using subsets of the earthquake catalogue results in values of n and K that vary significantly. This is likely due to the uneven spatial distribution of earthquakes and temporal clustering of hypocentres at particular locations, discussed further in Chapter 4.

We therefore instead use the existing local magnitude scale from [Rhoades et al. \(2020\)](#) and calibrate earthquake magnitudes and station corrections relative to this scale and the corrections of the permanent network with jack-knifing of the earthquake catalogue. For the temporary seismograph network, we derive station corrections by calibration with the permanent network and existing local magnitude scale.

We solve for station corrections using the local magnitude scale of [Rhoades et al. \(2020\)](#):

$$-\log(A)_{calc} = -\log(A) + M_L, \quad (3.14)$$

from which we calculate the misfit relative to the predicted attenuation curve, defined as:

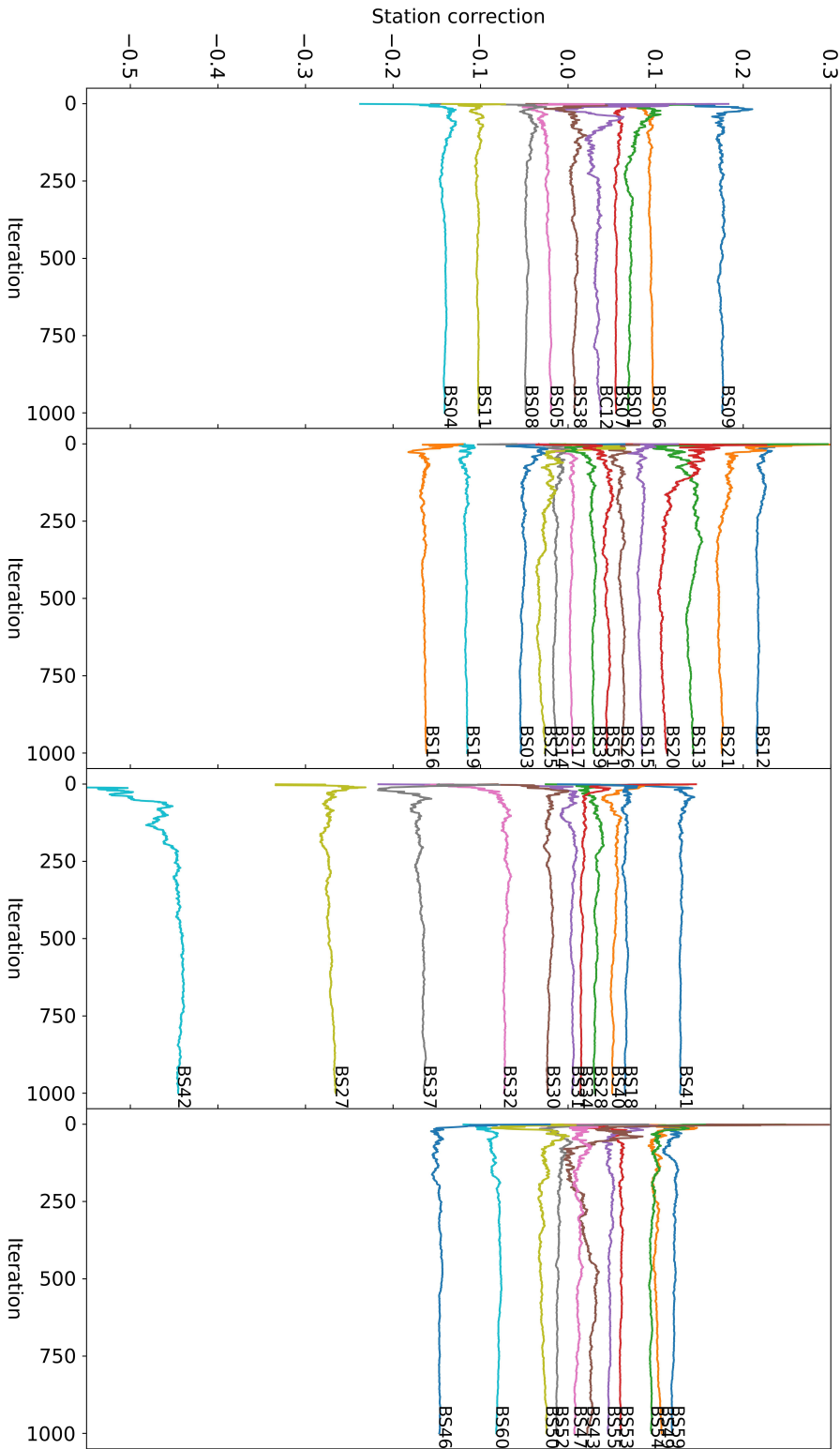
$$-\log(A)_{pred} = -0.51 + 7.9 \times 10^{-4} \Delta + 1.67 \log(\Delta). \quad (3.15)$$

We first calculate the misfit using all earthquake-station pairs in the final catalogue, yielding 52,962 amplitude measurements. Amplitude measurements are obtained by determining the turning points of the waveform 20 seconds before and 30 seconds after the S-phase arrival. The maximum peak-to-peak measurement halved between positive and negative turning points is used as our amplitude measurement. There is strong variation in the number of earthquakes at each depth; the majority of earthquakes are located in the upper 40 km i.e., in the overriding plate and crust of the subducting slab. We bin the data into 10 km depth intervals, and sample subsets of the data to test the dependency of the station corrections on depth distribution of earthquakes. The data are jack-knifed, generating 100 datasets each containing between 200 and 12,000 earthquakes, depending on the bin width and number of earthquakes within each bin. We find that regardless of the number of earthquakes within each bin, the resultant station corrections are most significantly sensitive to the ray coverage. When earthquakes at depths > 40 km are included in the misfit calculation, the mean station correction increases compared with when only earthquakes at depths < 40 km. [Rhoades et al. \(2020\)](#) demonstrated that there is a strong depth-dependency in local magnitude calculation due to mantle attenuation, and therefore incorporated a depth-dependent term.

We recalculate the local magnitude scale parameters and jack-knifing with only earthquakes at depths < 40 km, and find that the calculated station corrections are in line with those previously published for the permanent network stations ([Rhoades et al., 2020](#)). The data are binned into 5 km depth intervals, and from each bin an equal number of earthquakes are randomly sampled, to create 1000 subsets of data each containing 1000 earthquakes. The station correction for each subset of 1000 earthquakes is calculated, thus generating 1000 station correction values for each station. We then calculate the cumulative mean station correction, as more data are sampled, for up to 1000 iterations.

We then plot the zero-meanded station corrections, to pseudo-normalise them. Typical station magnitude uncertainties are 0.2 (e.g., [Hutton and Boore \(1987\)](#), [Merry et al. in prep.](#)). We

Figure 3.8: Station magnitude corrections with Jack-knifing. Station data are separated into four panels for ease of viewing.



see that after iteration 700, all stations have flattened curves, and we take the 700th iteration station corrections into subsequent local magnitude calculations.

The geology of this region is characterised by Miocene medium- to fine-grained sandstone and mudstone in the south-east portion of the network and sandy, shelly limestone and calcareous sandstone in the north-west ([Mazengarb and Speden, 2000](#)). Separating these two lithologies is a band of poorly bedded mudstone in which the Waimata Valley mud volcanoes are located. While the differences in lithology are unlikely to be seismologically detectable, the stations located in the Waimata Valley region (BS09, BS20, and BS21; Figure 3.8) have larger station magnitude corrections and travel-time residuals, likely due to attenuation of the seismic signal through the mud conduits that feed the mud volcanoes, discussed further in Section 3.10.2 ([Leighton, 2020](#)).

3.6 Refining the earthquake catalogue

Despite the quality control criteria, EQTransformer and subsequent processing resulted in 102 events in the final earthquake catalogue that were actually misidentified active-source shots. From January to February 2018, a 3-D seismic reflection survey operated offshore Gisborne (Figure 3.9). These shots have known origin times and locations, as reported in [Bell et al. \(2017\)](#). Shots are easily ignored in manual picking, due to their regular 60-second spacing and increasing or decreasing signal-to-noise ratio through time as the ship sailed closer or further respectively from the seismograph network ([Bell et al., 2017](#)).

We required all events in the catalogue to have at least 2 S-wave phase picks; it is therefore surprising that S-wave phases were picked for active-source shots that do not create shear waves. Manual inspection of these events shows that they do indeed have small secondary arrivals. We suggest that the secondary arrivals are due to reverberations of the airgun signal in shallow sediments or the water column or are converted phases.

Active source shots were generally only observable on the seismograms when the ship approached the shoreline (Figure 3.9), therefore, shots in the earthquake catalogue were clustered in time. These were identifiable due to their clustered event locations and origin times spaced every 60 seconds. We extracted suspicious events from the catalogue based on

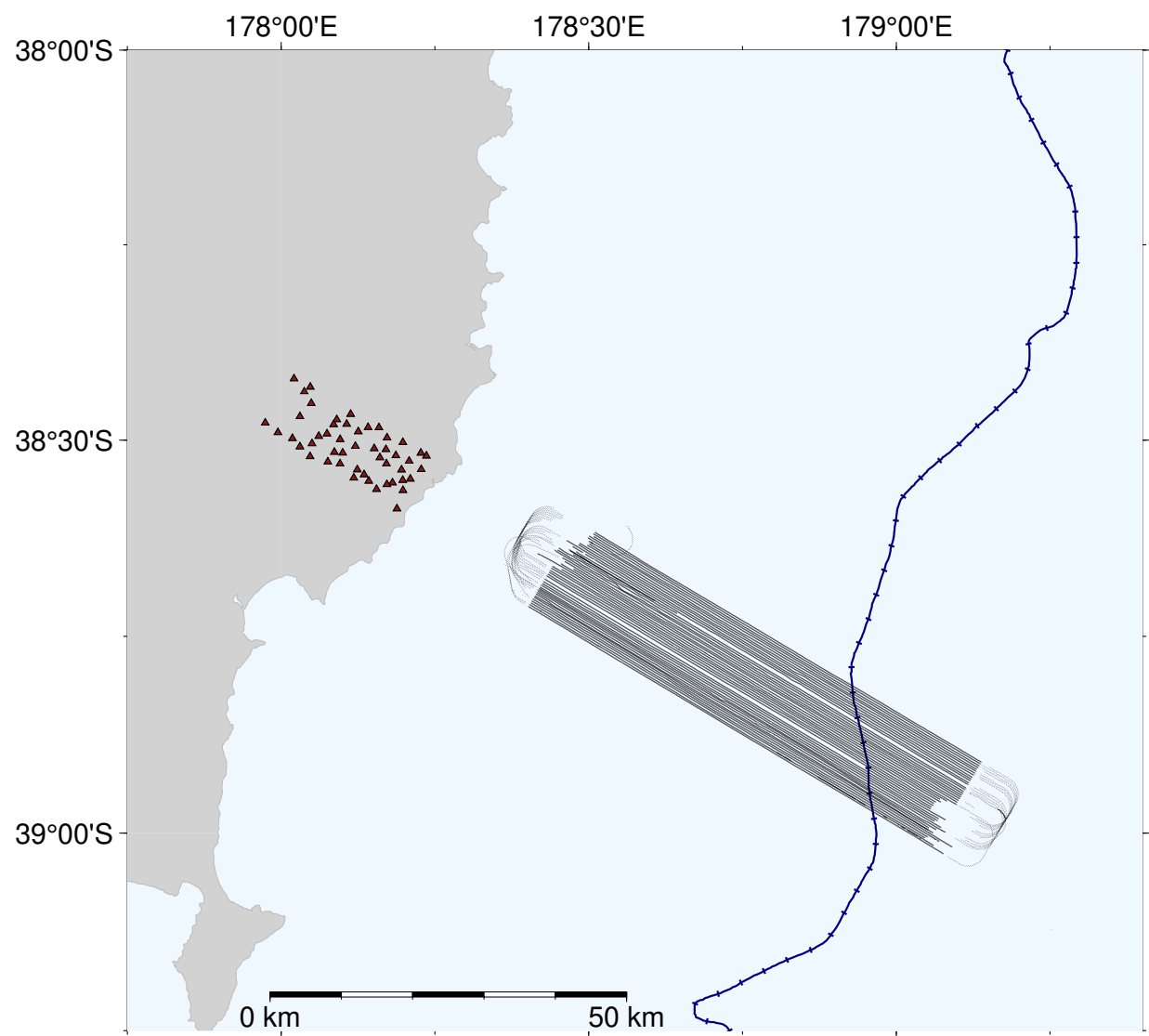


Figure 3.9: Active-source shot locations (black dots) during 3-D seismic reflection survey January-February 2018 (*Bell et al., 2017*). Red triangles: temporary seismograph network.

these criteria, resulting in a list of 155 events. These were manually inspected; 102 were categorised as shots and removed from the catalogue.

The NonLinLoc locations for the active-source shots were onshore, clustered close to the shore-line in the upper crust, directly beneath the south-east portion of the temporary seismograph network. The poor locations, i.e., the inability of NonLinLoc to recover the true airgun source location, are likely due to the water column and its effects on travel time and, therefore, event location, as well as the inclusion of false ‘S-wave phases’. Well-constrained hypocentres require S-phase picks ([Warren-Smith et al., 2024](#)). The shots have known source locations offshore; the spurious locations are likely due to the 1-D and 3-D velocity models used which have poor offshore coverage, thus poorly constraining the region through which the airgun signal travels ([Eberhart-Phillips and Bannister, 2015](#); [Yarce et al., 2021](#)). Furthermore, the active source shots have a mean magnitude of -0.47 (though magnitude is a function of location), considerably smaller than the earthquake mean magnitude of 1.45. Collectively, these characteristics allow us to identify airgun shots and remove them from the catalogue.

3.7 Frequency-magnitude analysis

We analyse the frequency-magnitude relationship and b -value for our new catalogue and for the GeoNet catalogue for the same period. The frequency-magnitude relationship is described using:

$$\log N = a - bM, \quad (3.16)$$

where N is the number of earthquakes with magnitude greater than M , and a and b are constants ([Gutenberg and Richter, 1944](#)). Following [Wiemer \(2000\)](#), we calculate the magnitude of completeness (M_C), which represents the magnitude above which a power law fit can describe the frequency-magnitude distribution of the observed data. M_C is the minimum magnitude for which 90% of the data can be fit by a straight line. We calculate the misfit for possible M_C values, between M_L 0.5 and 2.0, with increments of 0.1 following the method of [Chamberlain et al. \(2018\)](#). The difference between a synthetic frequency-magnitude distribution (Equation

3.16) and the observed frequency-magnitude distribution is calculated.

The goodness of fit is calculated using the normalised difference (R) between the number of earthquakes observed in the magnitude bin (0.1) and the predicted number of earthquakes; the larger the R-value, the better the fit. We reach a maximum R-value for M_C of 1.7, with a value of 99.7%. We compare the M_C value calculated using the power-law fit method to the maximum curvature method (Wiemer, 2000), which calculates the gradient of the frequency-magnitude plot at discrete intervals. This method tends to underestimate M_C , and is sensitive to the bin width chosen. For example, a magnitude bin of 0.1 gives M_C 1.3, while a magnitude bin of 0.25 gives $M_C = 1.75$. We use M_C of 1.7 to calculate the b -value. This gives an a -value of 4.89 and b -value of 1.06 (Figure 3.10), where a b -value of 1 is the global average (Lay and Wallace, 1995). For the same observation period, using the same methods, the GeoNet catalogue gives $M_C = 2.0$.

3.8 Focal mechanism analysis

The FOCal MEchanism determinations (FOCMEC) code uses first-motion polarities to perform an incremental grid search for focal mechanisms that best fit the observed polarity distribution on a lower-hemisphere projection (Snoke, 2003). We attempted to determine focal mechanism solutions for all earthquakes with $M_L > 2.5$, as well as any earthquake (regardless of magnitude) where the hypocentral uncertainty was < 2.5 km and the RMS travel-time misfit was < 0.25 s. This resulted in 445 earthquakes.

We manually picked P-wave arrivals on unfiltered vertical component seismograms, and classified arrivals as upwards or downwards (compressional or dilatational) and emergent or impulsive. For each earthquake-station pair, azimuth and take-off angle were computed with our 1-D velocity model and the TauP toolkit (Crotwell et al., 1999), defining the observation position on the focal sphere. Emergent picks were labelled near-nodal and down-weighted. The grid search used 1° increments; we allowed at most one polarity error (a pick in the opposite-polarity quadrant) per acceptable solution and retained up to 100 solutions per event. This resulted in 146 focal mechanisms.

We then calculated the median strike, dip, and rake, and discarded mechanisms with strike or

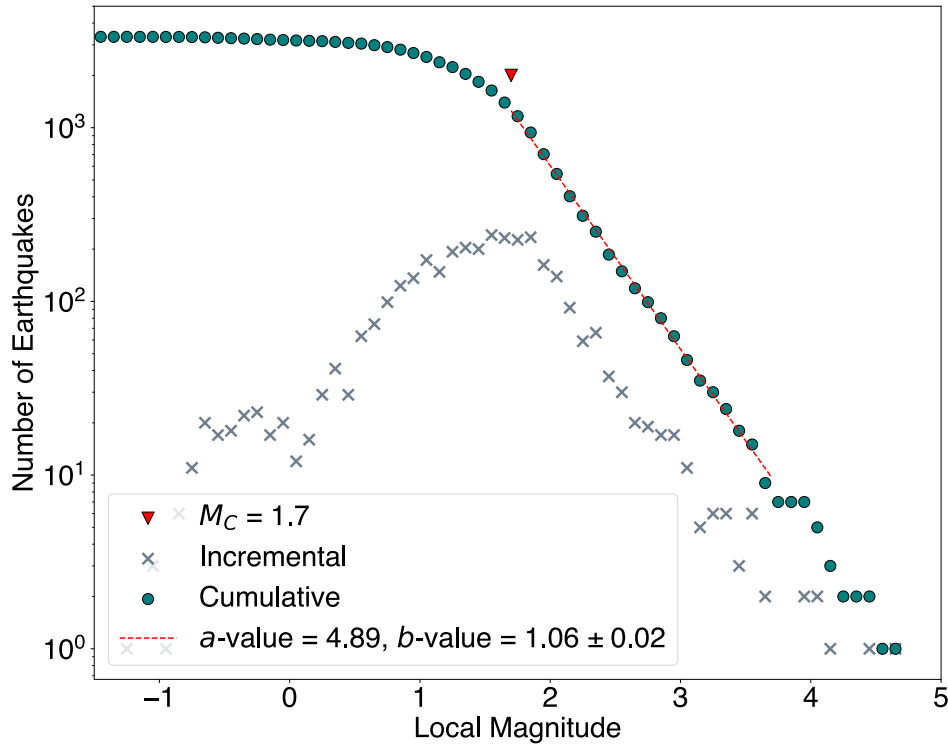


Figure 3.10: Frequency-magnitude distribution for the catalogue in this study. Circles: Gutenberg-Richter curve for no. of earthquakes with magnitude greater than ML. Crosses: incremental count for number of earthquakes of each magnitude in 0.1 increments. Red inverted triangle: magnitude of completeness (MC). b-value uncertainty is calculated following Aki (1965).

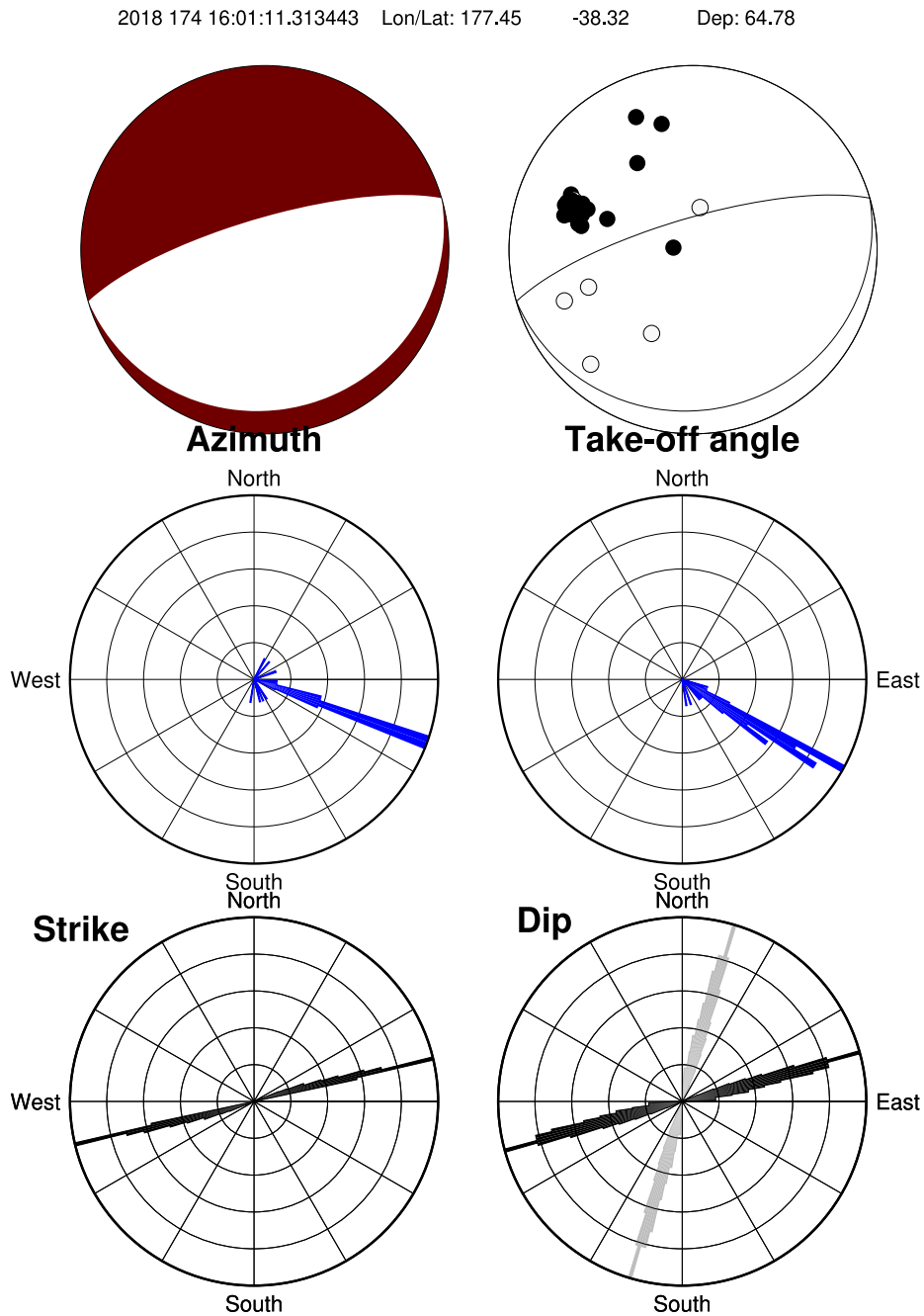


Figure 3.11: Example focal mechanism. **Top row:** Focal mechanism derived from polarity picks. Open circles: dilatational polarities. Black circles: compressional polarities. **Middle row:** Rose diagrams show azimuth of earthquake-station pair and take-off angles for measurements used to derive this focal mechanism. **Bottom row:** histograms of strike and dip solutions for focal planes for this earthquake.

dip standard deviation $>10^\circ$. For accepted mechanisms, mean standard deviations were 5° (strike), 2.7° (dip), and 0.3° (rake). After visual inspection, we retained 86 focal mechanisms, 44 of which display normal faulting. An example of measurements to derive a focal mechanism is shown in Figure 3.11.

3.9 Results

We initially detect and locate 13,858 earthquakes. Applying criteria of hypocentral uncertainty ≤ 5 km and RMS misfit between predicted and observed travel times ≤ 0.5 s yields a catalogue of 3,071 earthquakes with mean and median depth uncertainties of 2.58 km and 2.38 km, respectively (Figure 3.12). Earthquake magnitudes range between -0.84 and 4.40 (Figure 3.12). For the same observation period, the GeoNet catalogue reports 1,717 earthquakes with magnitudes between 0.4 and 4.6, but without hypocentre uncertainties.

We derive 86 focal mechanisms (Figure 3.15). Earthquakes for which focal mechanisms were derived are located primarily onshore and at similar depths to those observed in previous studies (e.g., [François-Holden et al., 2008](#); [Lentas, 2018](#); [Lentas et al., 2019](#); [Warren-Smith et al., 2019](#)). Normal faulting is observed in the upper plate and in the subducting slab. Thrust faults are observed, particularly near the plate interface, though we also observe upper plate and subducting slab thrust events.

Earthquake distributions are consistent between 1-D and 3-D velocity models. The new 1-D velocity model, A-IVZ, yields the largest number of high-quality hypocentres (i.e., hypocentral uncertainty ≤ 5 km and RMS travel-time misfit ≤ 0.5 s), so we use these earthquakes in subsequent analysis.

Figure 3.13 shows the catalogue derived in this thesis compared to the GeoNet catalogue for the same observation period. Our catalogue improves the detection and location of earthquakes, particularly $M_L < 1.0$. GeoNet also uses fixed-depth earthquakes, resulting in unrepresentative earthquake locations and inaccurate magnitude determinations. In particular, we improve coverage of upper plate and deeper subducting slab earthquakes.

The upper plate is in a state of extension along the east coast ([Henrys et al., 2025](#)), while

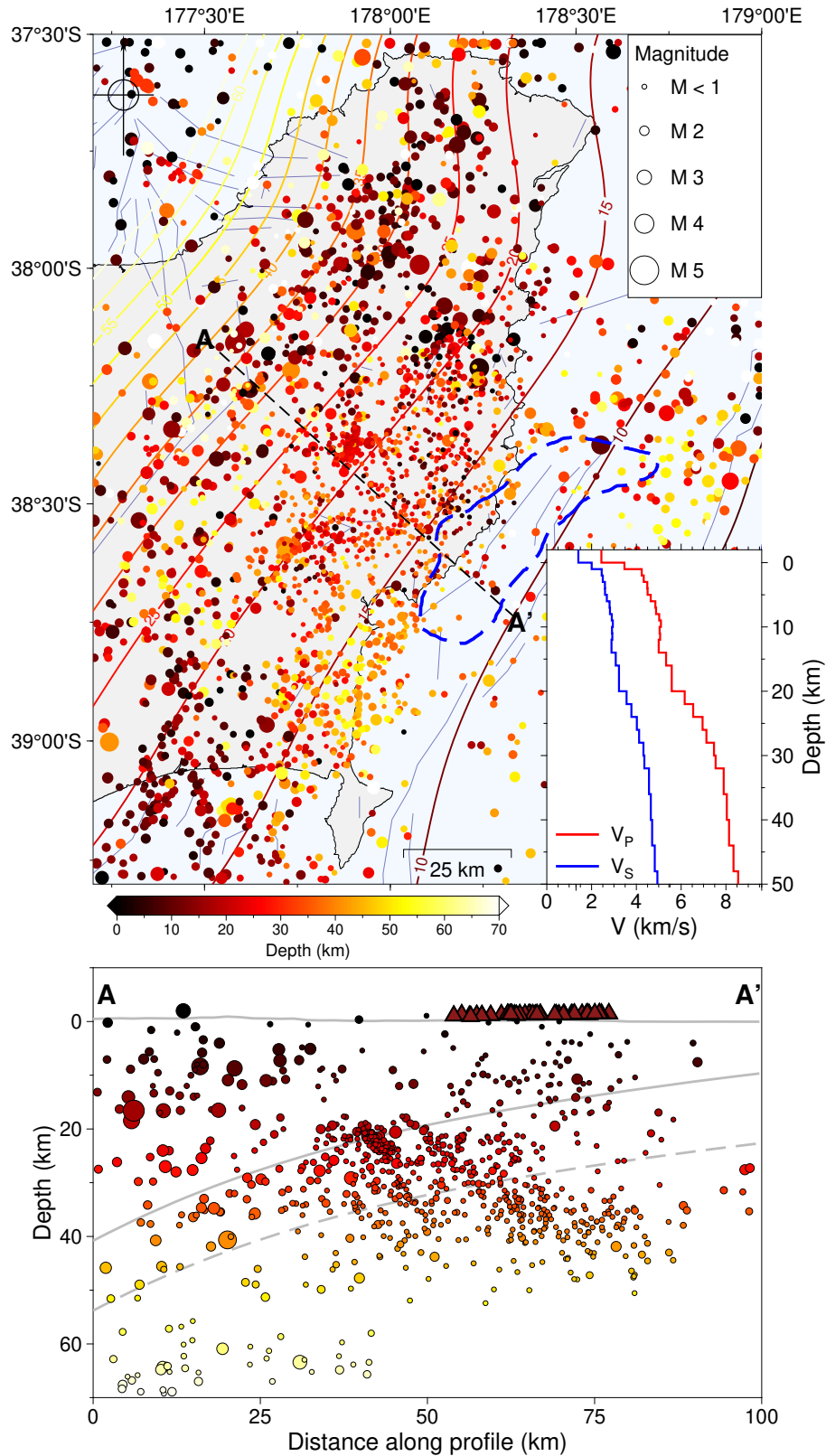


Figure 3.12: Map view: Final earthquake catalogue. Earthquakes with $M_L < 1.0$ are plotted as a constant size. Navy lines: faults. Blue dashed line: seismicity gap (Yarce *et al.*, 2019; Mochizuki *et al.*, 2021). Black dashed line: cross-section shown in lower panel (AA'). Contours: plate interface depth (Williams *et al.*, 2013). Inset: velocity model used to locate earthquakes. **Cross-section view:** Earthquakes through line AA' (25 km either side of line). Triangles: Gisborne network seismometers.

further inland, upper plate seismicity is observed in association with the Taupō Volcanic Zone. Upper plate seismicity recorded in this study is $M_L < 2.0$, below the detection capability of routine analysis in Aotearoa New Zealand ([Delahaye et al., 2009](#)). The addition of the high-density seismograph network enables the detection of upper plate seismicity.

3.10 Discussion

3.10.1 Implications of velocity model on earthquake locations

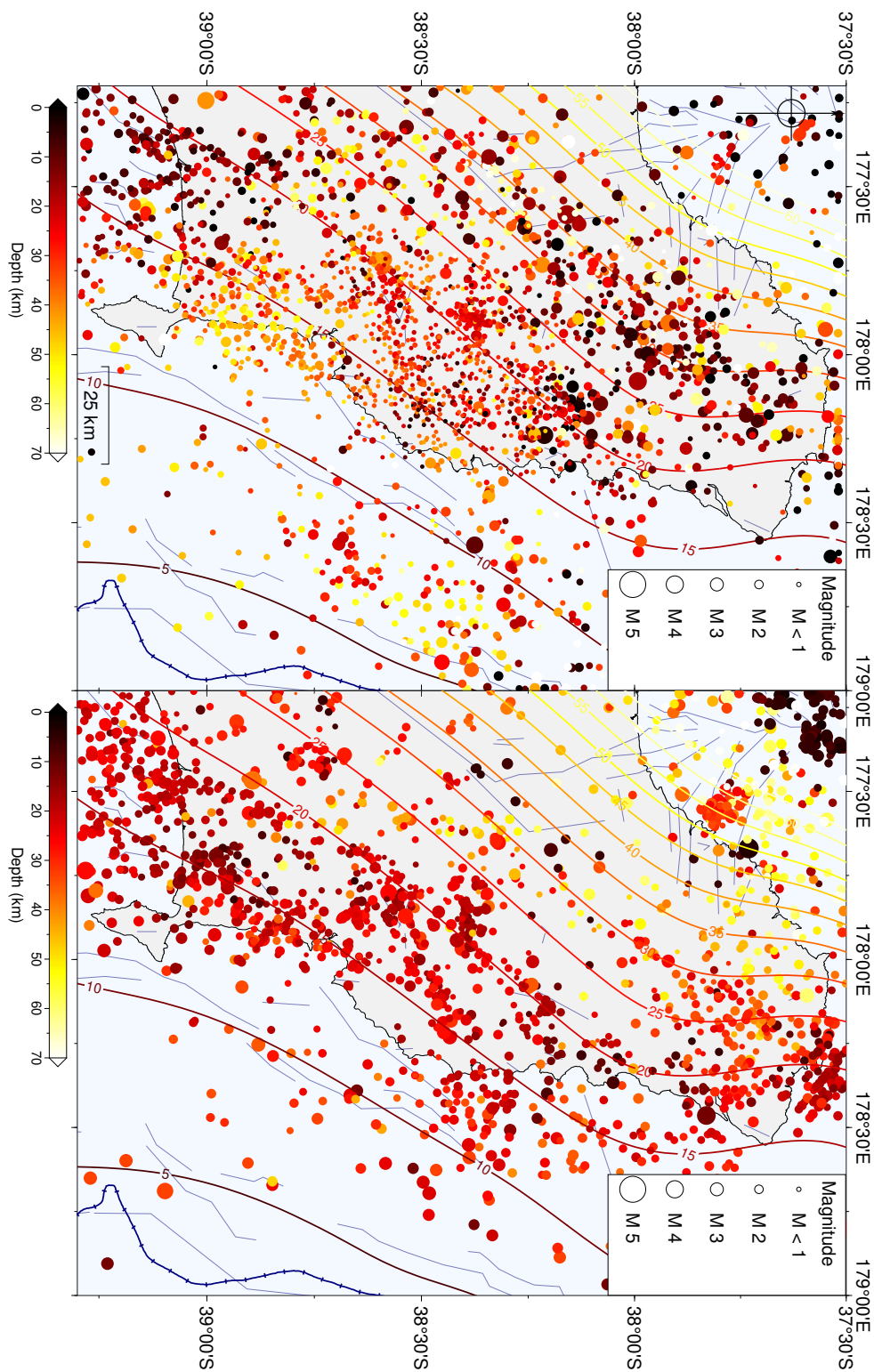
We located the earthquake catalogue using our newly derived 1-D velocity model, the 1-D velocity model of [Yarce et al. \(2021\)](#), a 1-D extraction from the 3-D New Zealand velocity model of [Eberhart-Phillips and Bannister \(2015\)](#), a 1-D extraction from the 3-D velocity model of [Bassett et al. \(2025\)](#), and with a full 3-D velocity model inversion of [Eberhart-Phillips and Bannister \(2015\)](#).

The earthquake distributions are consistent between 1-D and 3-D velocity models, but the use of our new 1-D model results in a smaller mean hypocentral error (Figure 3.16). The new 1-D velocity model, A-LVZ, yields the largest number of high-quality hypocentres (i.e., hypocentral uncertainty ≤ 5 km and RMS travel-time misfit ≤ 0.5 s), thus we use these earthquakes in subsequent analysis. We also located the earthquakes with a 3-D velocity model ([Eberhart-Phillips and Bannister, 2015](#)), which provides consistent seismicity distributions when compared with locations with 1-D velocity models. Furthermore, the same spatiotemporal clusters of seismicity are observed across a wide range of 1-D and 3-D velocity models, suggesting their observation is robust.

3.10.2 Station travel-time variation

Station travel-times are highly sensitive to the presence of fluids. Calculation of station V_P/V_S (based on travel-time measurements) provides insight into how the shallowest structure beneath each station affects travel-times, thus providing an indication of the presence of fluids. All arriving ray paths travel through the shallowest structure directly beneath the station. Therefore, a high V_P/V_S may suggest that a station is located on fluid-rich ground e.g., where

Figure 3.13: Left) Catalogue derived in this study. Right) Catalogue from GeoNet for the same observation period.



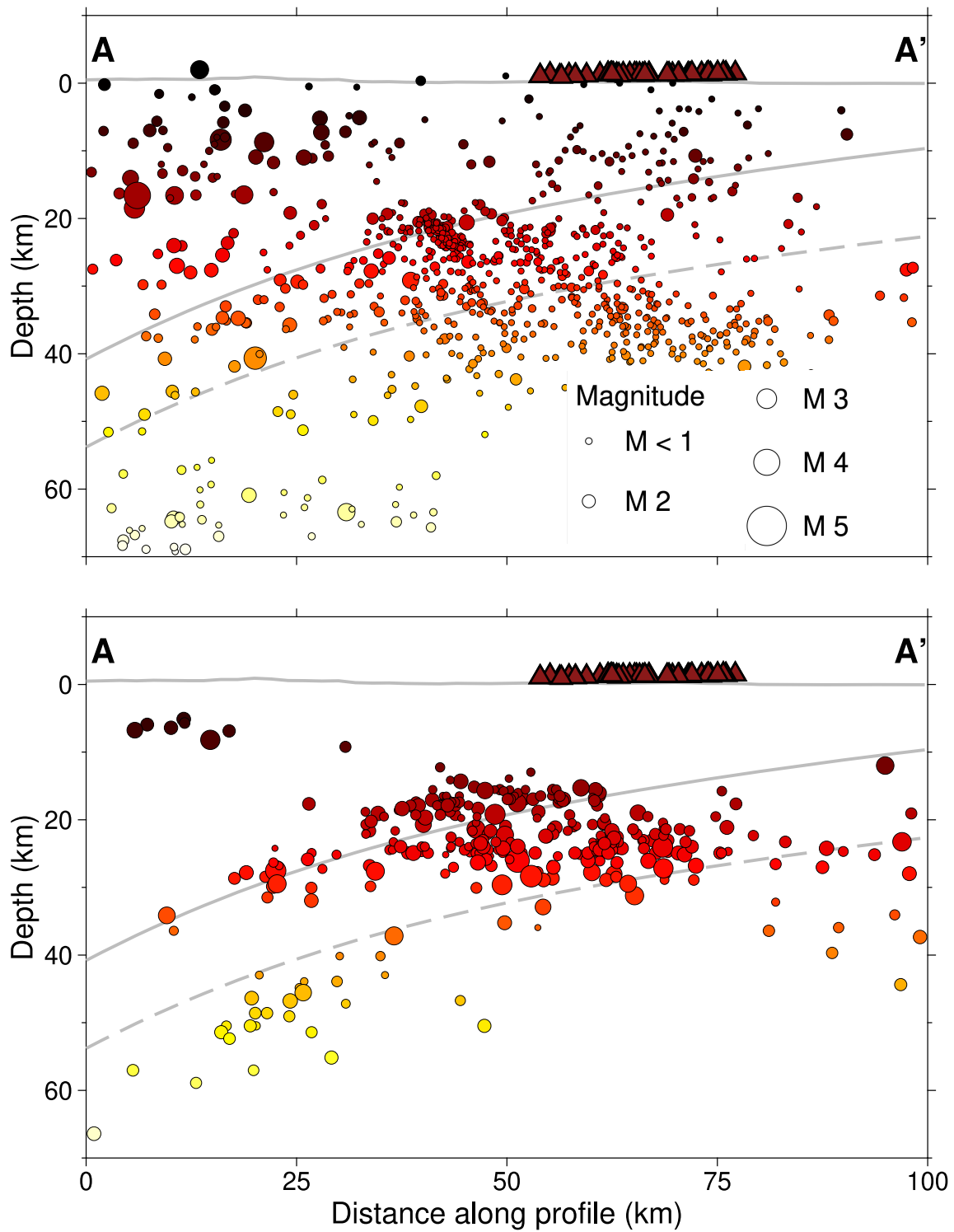


Figure 3.14: Upper) Catalogue derived in this study in cross-section. Cross-section orientation is the same as that in Figure 3.12. Lower) Catalogue from GeoNet for the same observation period.

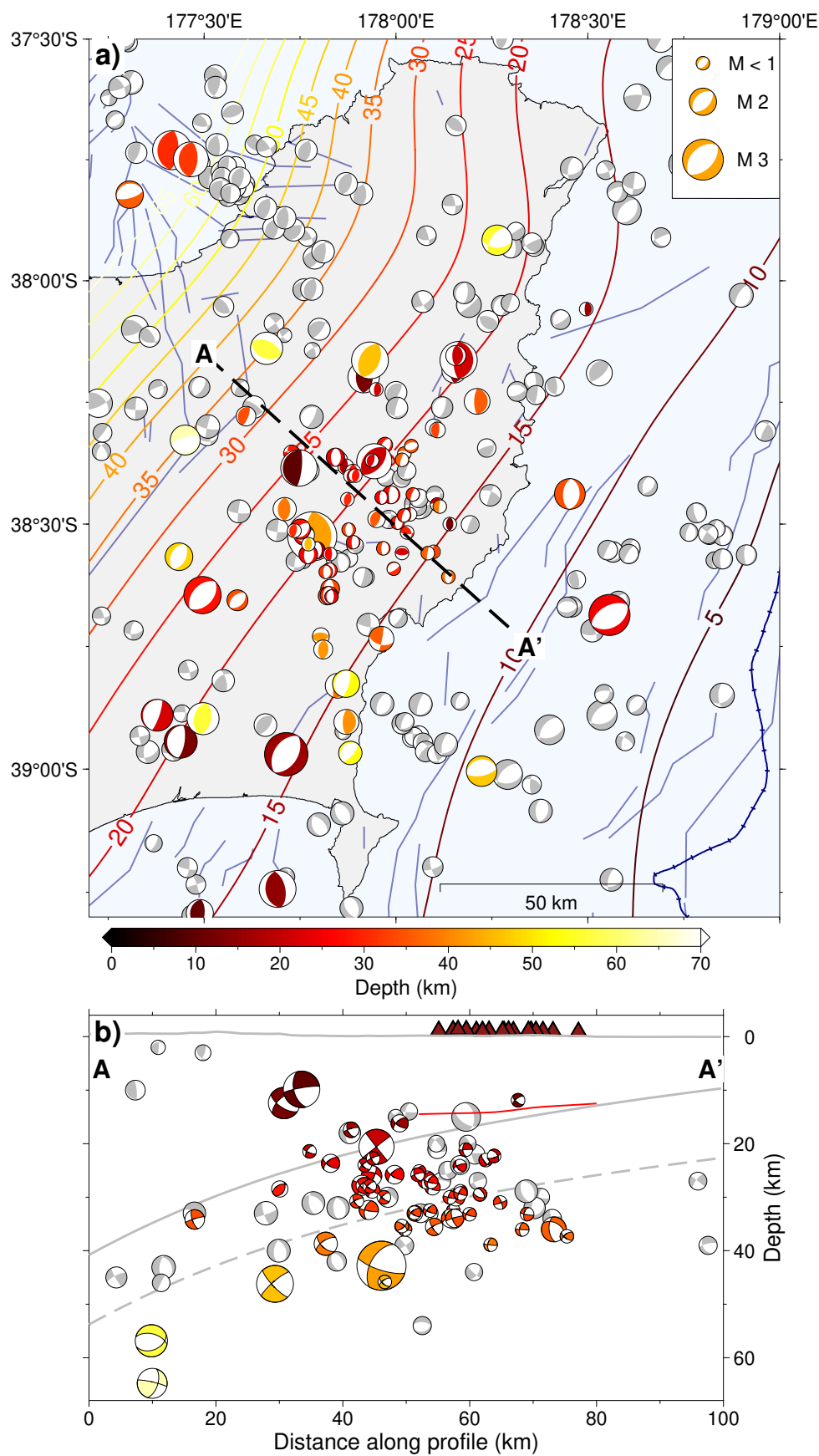


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Figure 3.15: **a)** Map view of focal mechanisms derived in this study. Grey focal mechanisms are from the ISC (*Lentas et al., 2019*). **b)** Cross-section through AA' of focal mechanisms derived in this study (width is ± 25 km about the cross-section line). Red triangles: Gisborne temporary seismograph network. Red line: plate interface from *Leah et al. (2022)*. Grey line: plate interface from *Williams et al. (2013)*. Dashed grey line: subducting slab Moho from *Williams et al. (2013)*.

there are mud volcanoes.

We calculate V_P/V_S for each station-event pair where

$$\frac{V_P}{V_S} = \frac{t_S}{t_P}, \quad (3.17)$$

where

$$t_S = T_S - T_0, \quad (3.18)$$

$$t_P = T_P - T_0, \quad (3.19)$$

and T_S and T_P are travel-times of the P and S wave respectively, and T_0 is the origin time of the earthquake. We calculate the average along-path V_P/V_S for each station-event pair, then calculate the median V_P/V_S for each station. Median V_P/V_S for each station is shown in Figure 3.17. Higher V_P/V_S may be linked to fractures and fluid-rich regions, discussed further in Chapter 4.

Station V_P/V_S values are correlated with regional geology. We observe higher V_P/V_S at stations closest to the 2018 mud volcano eruption sites (cyan stars, Figure 3.18; *Leighton, 2020*), indicative of higher fluid presence resulting in low V_S . The four stations closest to the mud volcanoes show particularly high V_P/V_S values, ranging 1.85–1.90, likely due to the mud conduits that supply the volcanoes (*Ridd, 1970*). Additionally, stations in the north-west of the Gisborne network show lower V_P/V_S values of 1.71–1.78, correlated with the sandstone bedrock (*Mazengarb and Speden, 2000*). Towards the coast, particularly where slumping and deep-seated landslides are observed (*Böttner et al., 2018*), mudstones and muds comprise the

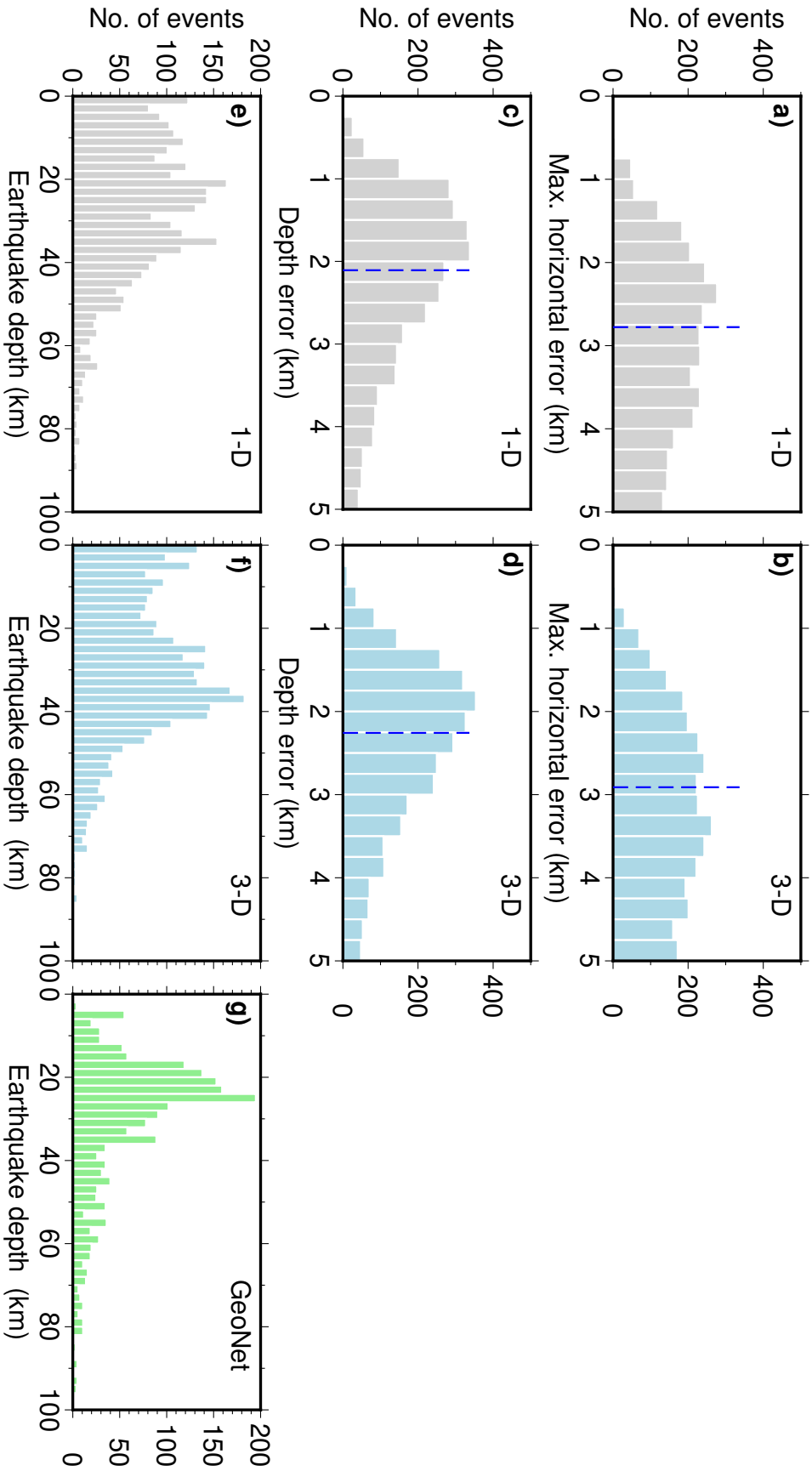


Figure 3.16: First column (grey): catalogue from this study located with our new 1-D velocity model. Middle column (blue): catalogue from this study located with the 3-D velocity model of *Eberhart-Phillips and Bannister (2015)*. Third column: GeoNet catalogue where available. **a)** horizontal error (km) distribution with 1-D model. **b)** horizontal error (km) distribution with 3-D model. **c)** depth error (km) distribution with 1-D model. **d)** depth error (km) distribution with 3-D model. **e)** earthquake depth distribution with 1-D model. **f)** earthquake depth distribution with 3-D model. **g)** GeoNet catalogue depth distribution. Note that GeoNet do not report hypocentre errors for their catalogue.

bedrock, resulting in higher V_P/V_S measurements.

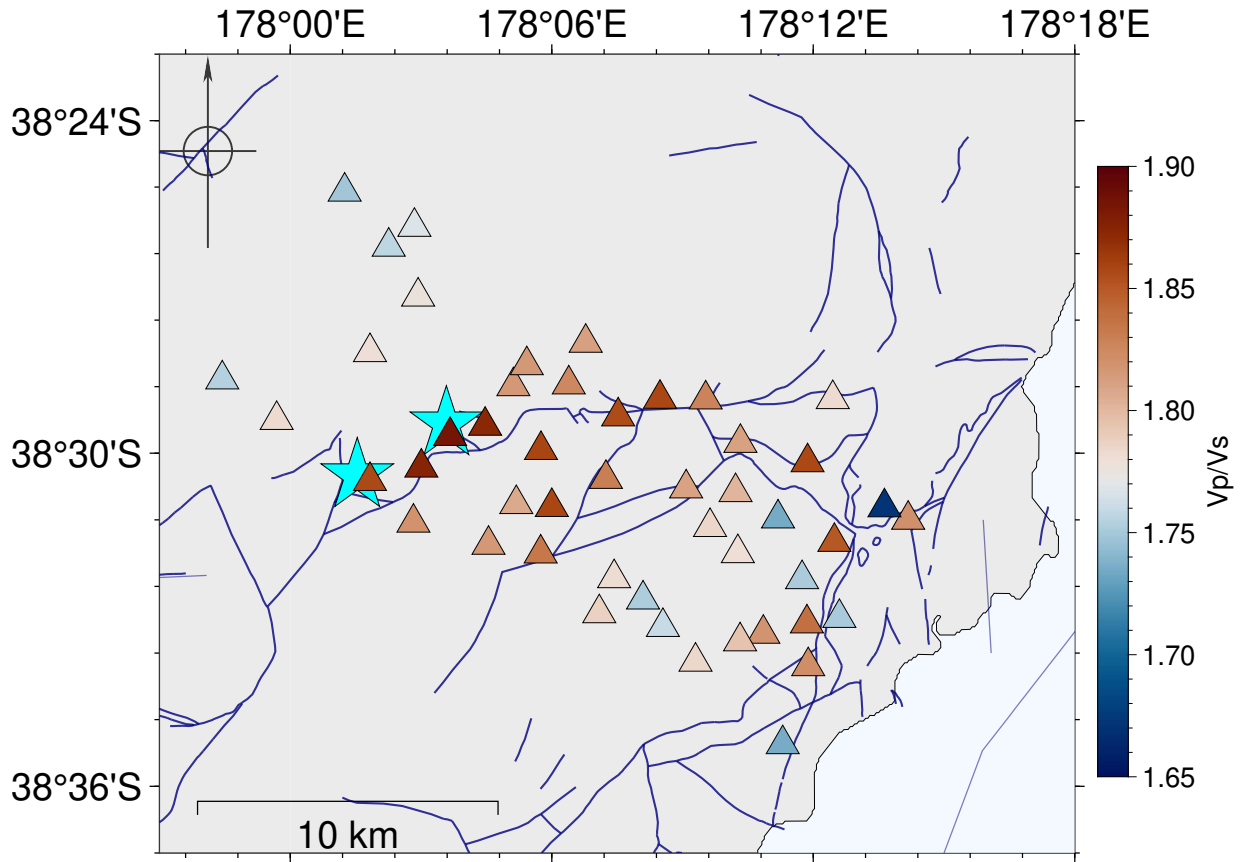


Figure 3.17: Median station V_P/V_S values calculated from earthquake-station pair measurements of P and S-P travel-times. Stars: mud volcanoes.

Median station V_P/V_S calculated from earthquakes at depths shallower than 20 km sample only the upper plate (Figure 3.18). The V_P/V_S values at the stations closest to the mud volcanoes are higher for earthquakes at depths greater than 20 km (Figure 3.19) than for earthquakes shallower than 20 km depths (Figure 3.18). *Ridd (1970)* demonstrated that these mud volcanoes are fed by mud diapirs at depth that possibly continue to the plate interface; this is supported by the variation in V_P/V_S . It is also possible, however, that the high V_P/V_S values here are associated with faults and fluids. These stations lie close to faults (Figure 3.18; *Mazengarb and Speden, 2000*) that may be fed by fluids at depth, thus increasing the V_P/V_S . To discern these mechanisms, future work considering seismic anisotropy and investigation of variation in delay times and fast directions is needed.

Figure 3.20 and 3.21 shows P- and S-wave travel-time residuals. Stations close to mud volcano eruption sites show particularly delayed S-wave travel-times, compared to the predicted travel-

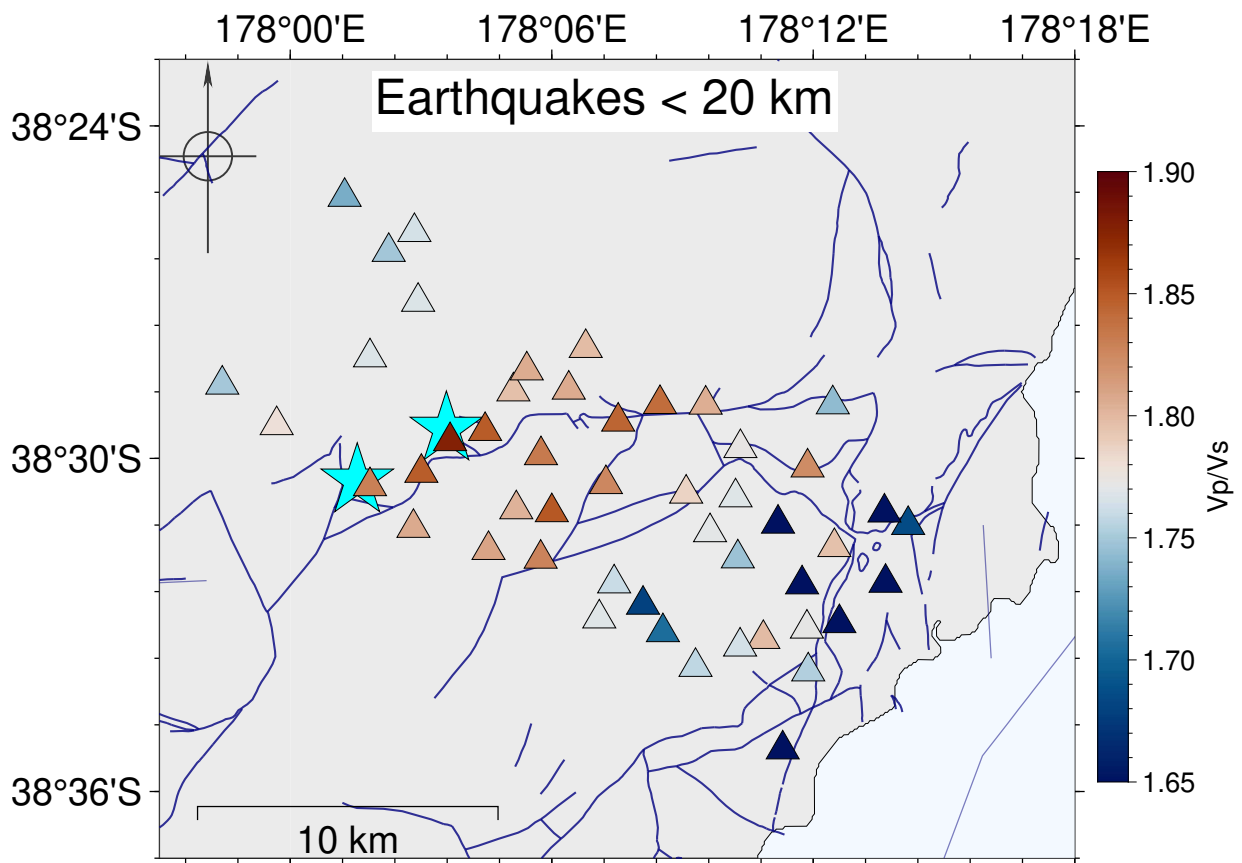


Figure 3.18: Median station V_P/V_S values calculated from earthquake-station pair measurements of P and S-P travel-times from earthquakes <20 km depth. Stars: mud volcanoes.

time from the local 1-D velocity model. The south-east portion of the temporary seismograph network overlies multiple faults measured at the surface ([Mazengarb and Speden, 2000](#); [Litchfield et al., 2014, 2022](#)). The upper plate is in a state of extension ([Henry et al., 2025](#)) and the plate interface is interpreted as a fluid-rich region, from which fluids migrate upwards through the upper plate (e.g., [Warren-Smith et al., 2019](#); [Bangs et al., 2023](#)). The presence of fluids in this region likely accounts for the delayed S-wave travel-times compared to the 1-D velocity model and the seismograph stations in the north-west portion of the temporary network.

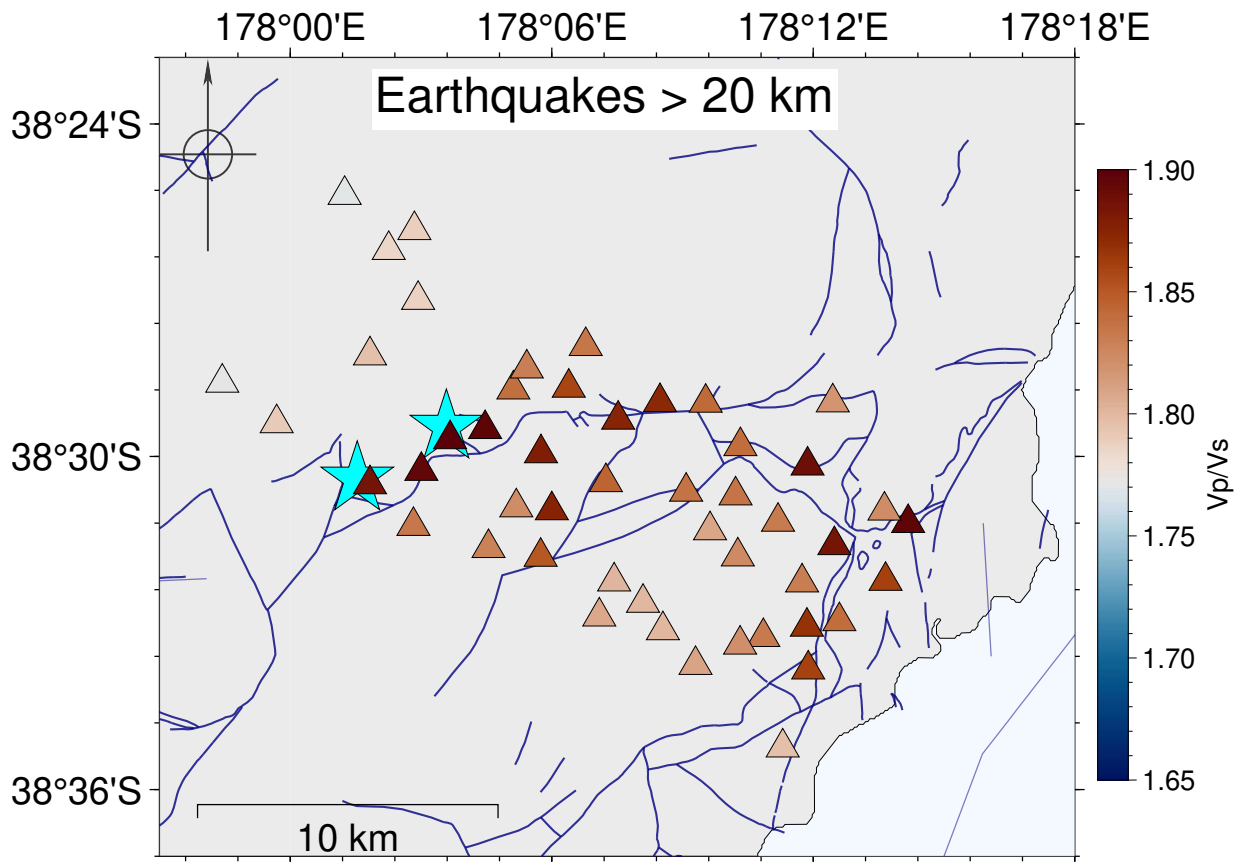


Figure 3.19: Median station V_P/V_S values calculated from earthquake-station pair measurements of P and S-P travel-times from earthquakes > 20 km depth. Stars: mud volcanoes.

3.10.3 Seismicity gap and plate coupling

The HOBITSS ocean-bottom seismograph network, deployed from May 2014 to June 2015, was designed to characterise seismicity in the offshore slow slip region. Seismicity observations in the HOBITSS catalogue continue to ~30 km onshore ([Yarce et al., 2019](#)). Our study to characterise the onshore portion, using the Gisborne network, results in a complementary

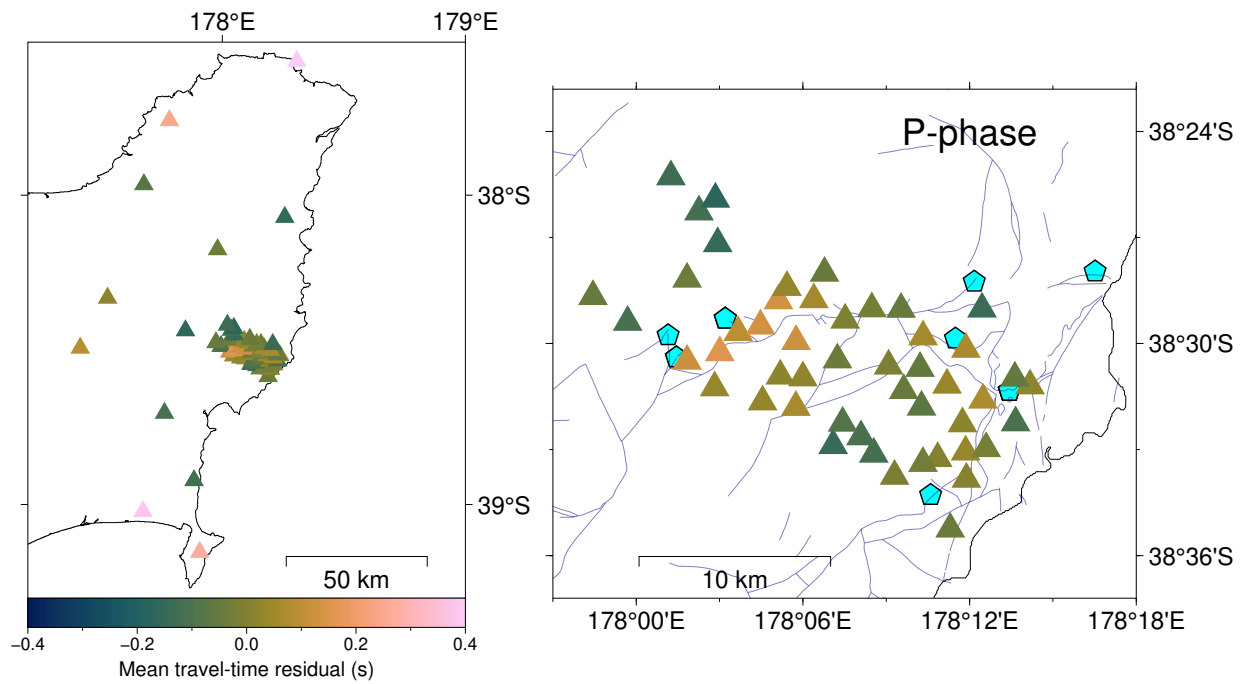


Figure 3.20: Mean P-wave travel-time residuals at each station. Pentagons: mud volcanoes or cold seeps.

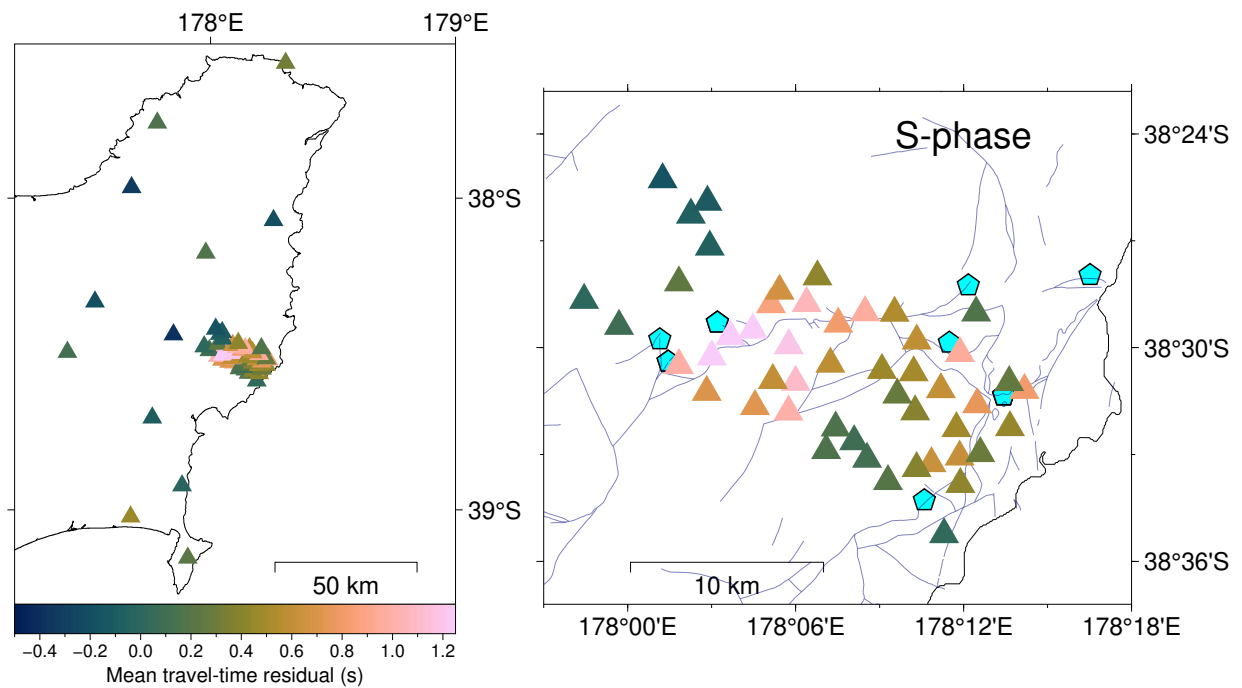


Figure 3.21: Mean S-wave travel-time residuals at each station. Pentagons: mud volcanoes or cold seeps.

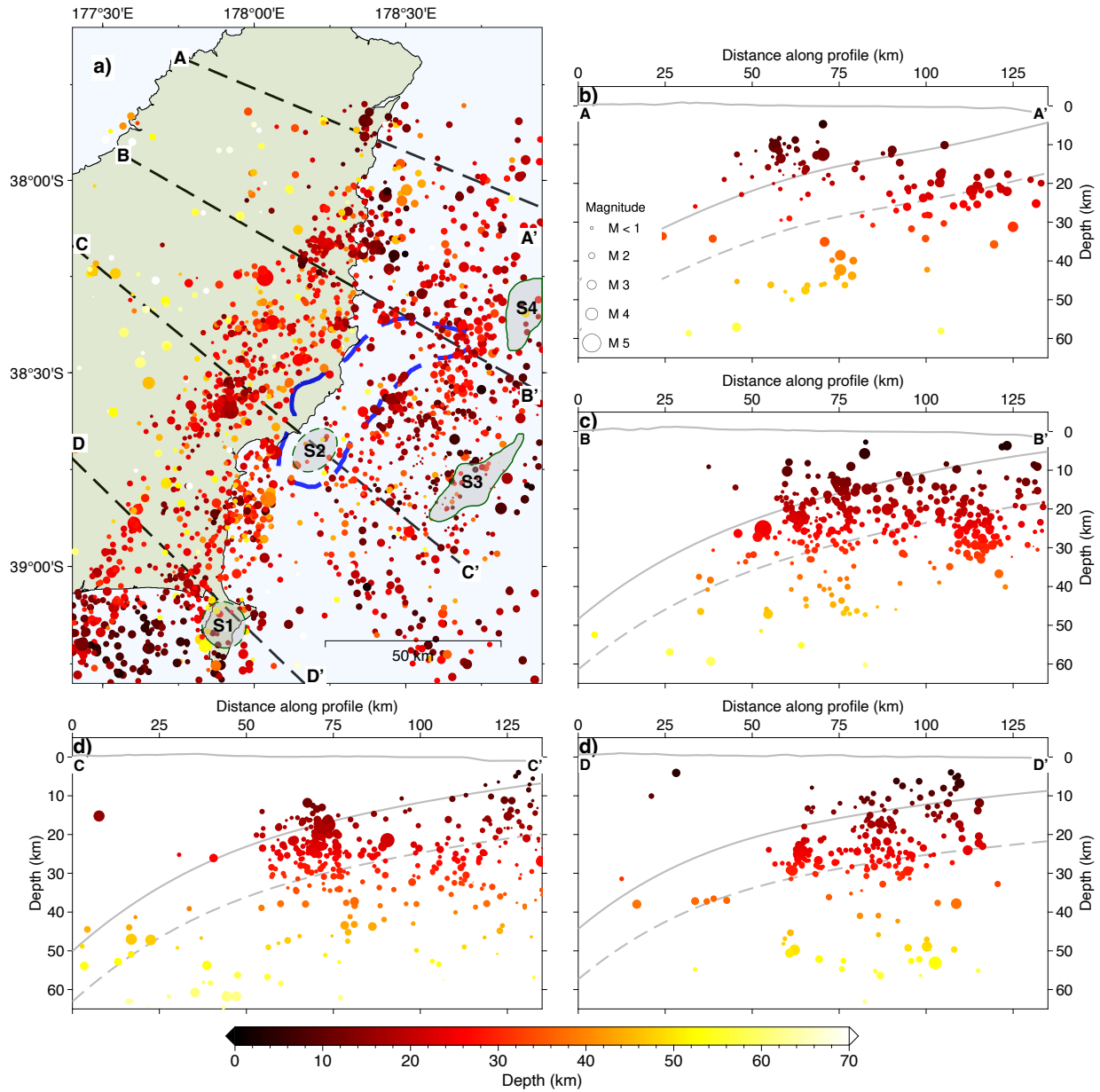


Figure 3.22: Earthquakes derived from the HOBITSS catalogue (Yarce *et al.*, 2019), and cross-sections AA' – DD' from north to south. Blue dashed line: seismicity gap from Yarce *et al.* (2019). Green dashed ellipses (S1 and S2): purported seamounts. Green solid lines (S3 and S4): seamounts (Bell *et al.*, 2010; Barker *et al.*, 2018). Gray line: plate interface (Williams *et al.*, 2013).

earthquake dataset with detailed coverage of onshore and offshore seismicity.

Seismicity is distributed through the upper plate, the plate interface, and the subducting slab. We observe a seismicity gap consistent with that reported by [Yarce et al. \(2019\)](#) and [Mochizuki et al. \(2021\)](#) (Figure 3.22). [Yarce et al. \(2019\)](#) identified a similar gap in the 2014 catalogue, even with the addition of offshore OBS stations that significantly improved coverage near the gap. Although we do not use OBS stations here and our offshore coverage is reduced, we still observe the same gap, and seismicity resumes further offshore beyond it. [Yarce et al. \(2019\)](#) suggested that the gap may reflect (1) interseismic locking (i.e., a temporal gap), (2) heterogeneous fluid distribution and permeability, or (3) a combination of these processes. The persistence of the seismic gap, observed in the HOBITSS catalogue from May 2014 to June 2015 and in this study from December 2017 to October 2018, suggests that it reflects interseismic locking, and we favour this first hypothesis. The inter-SSE plate coupling model, based on cGPS data from 2002–2010 shows that this segment of the margin (near Gisborne) shows slightly deeper plate coupling compared to the neighbouring segments (Figure 1.4; [Wallace, 2020](#)). [Yarce et al. \(2019\)](#) attribute the lack of seismicity in this region to plate locking. We observe a similar gap in seismicity, in agreement with [Yarce et al. \(2019\)](#). The combined earthquake catalogues show that the gap in seismicity continues further south-west than the shape defined in [Yarce et al. \(2019\)](#), suggesting that the locked patch may be larger. If we assumed that the entire locked patch could slip by 1 m in an earthquake, with an area of 50×25 km, this could result in a M_w 6.98 earthquake. Given its proximity to the coastline, and the city of Gisborne, this may pose a significant hazard.

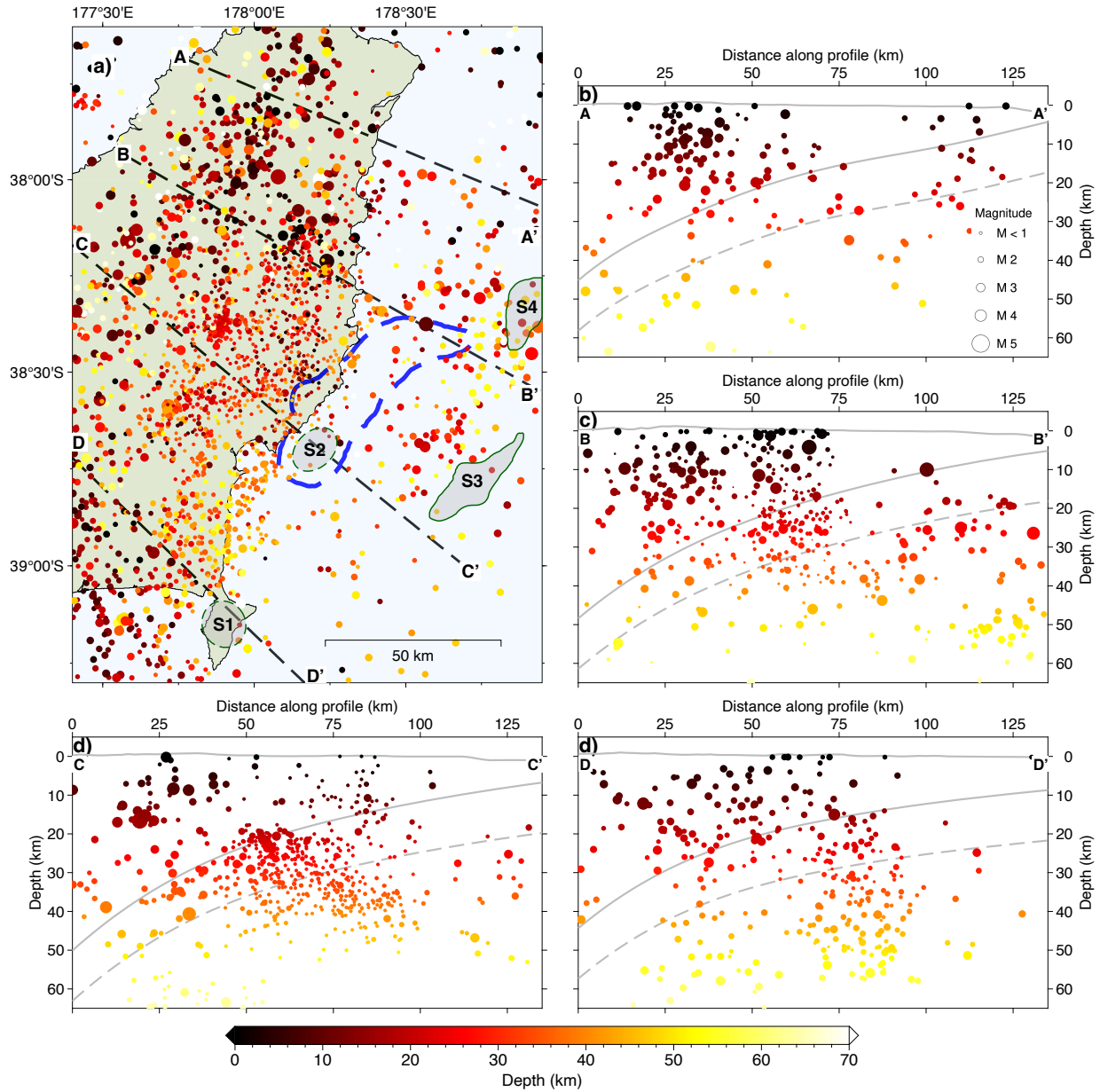


Figure 3.23: Earthquakes derived in this study in map view, and cross-sections AA' – DD' from north to south. Blue dashed line: seismicity gap from *Yarce et al. (2019)*. Green dashed ellipses (S1 and S2): purported seamounts. Green solid lines (S3 and S4): seamounts (*Bell et al., 2010; Barker et al., 2018*). Gray line: plate interface (*Williams et al., 2013*).

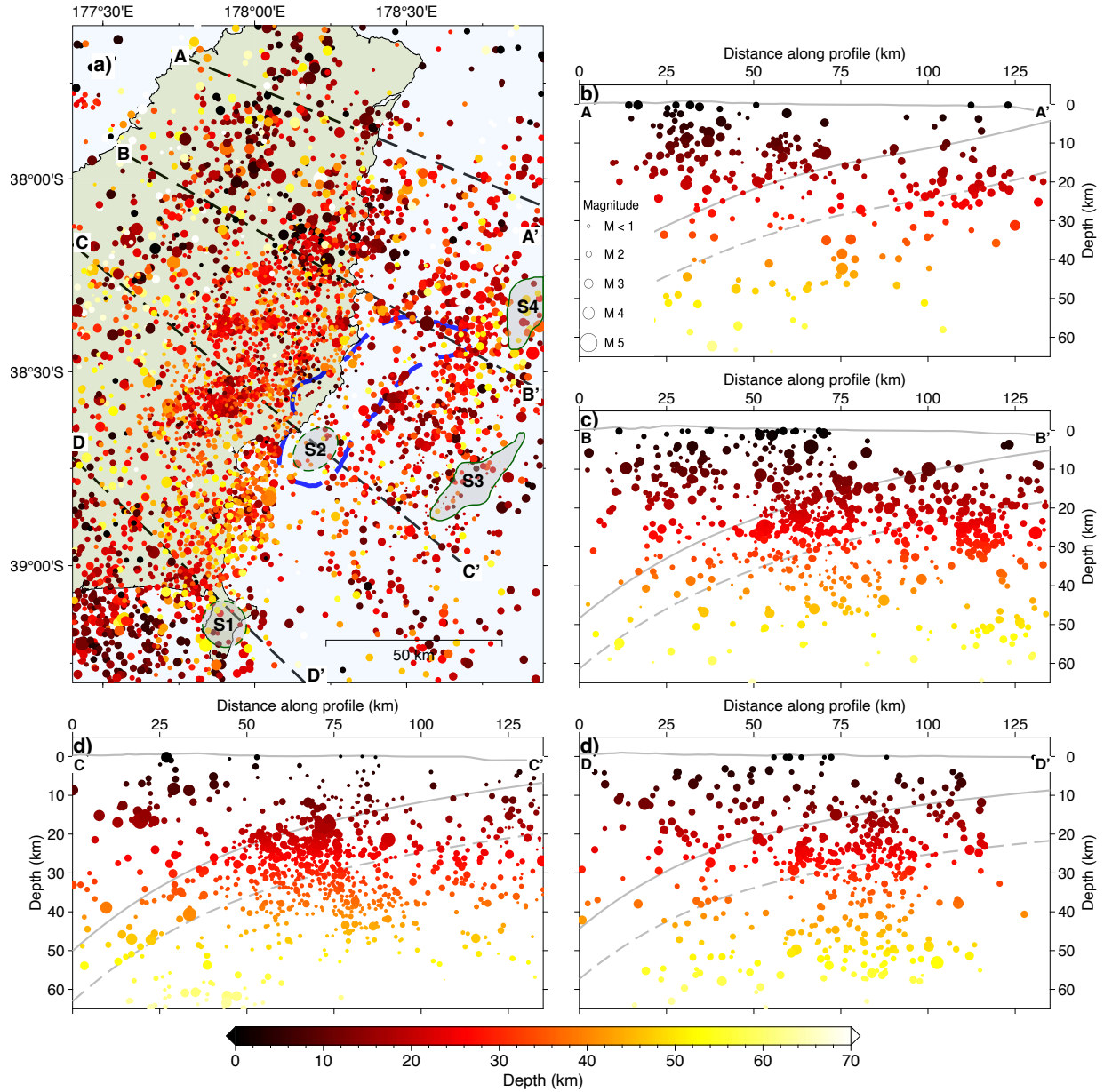


Figure 3.24: Earthquakes from the combined catalogues of this study and the HOBITSS catalogues in map view, and cross-sections AA' – DD' from north to south. Blue dashed line: seismicity gap from *Yarce et al. (2019)*. Green dashed ellipses (S1 and S2): purported seamounts. Green solid lines (S3 and S4): seamounts (*Bell et al., 2010; Barker et al., 2018*). Gray line: plate interface (*Williams et al., 2013*).

3.11 Conclusions

We present a high-resolution seismological study of the northern Hikurangi margin, Aotearoa New Zealand, using data from a dense temporary seismograph network that operated from December 2017 to October 2018 to detect local earthquakes. Earthquakes are observed in the upper plate, subducting slab and along the plate interface. We locate these earthquakes with multiple velocity models, including existing 1-D and 3-D velocity models and our newly-derived 1-D model. Initially, we detect over 13,000 earthquakes, which we then filter to retain only high-quality hypocentres with a maximum hypocentral uncertainty of 5 km and mean RMS travel-time misfit <0.5 s. Our newly derived 1-D velocity model, which includes a low velocity zone at 10–14 km depth, results in the largest catalogue of earthquakes that meet these criteria. The 1-D model is centred on Te Tai Rāwhiti/Gisborne, where the plate interface is at the same approximate depth as the low velocity zone. This gives a final catalogue of 3,071 earthquakes. Using this final catalogue, we calculate a completeness magnitude of 1.7, compared with 2.0 for the contemporaneous GeoNet catalogue.

We picked first motion polarities for 445 earthquakes and determined the fault plane solutions using the incremental grid search software FOCMEC. We retained 86 focal mechanisms, where the standard deviation of the strike, rake, and dip of the fault plane solutions was $<10^\circ$. Of these, 46 displayed normal faulting mechanisms.

In Chapter 4, we use the earthquake and focal mechanism catalogue to investigate the seismological characteristics of the margin. In particular, we consider the relationship between seismicity, subducted seamounts, seeps, and SSEs. These processes and phenomena affect or are affected by fluid distributions.

CHAPTER 4

The relationship between slow slip events and seismicity: the role of normal faults, seamounts, and seeps

The contents of this chapter are published [Woodward et al. \(2026a\)](#):

Woodward, A., I.D. Bastow, R. Bell, L. Wallace, K. Jacobs, S. Henrys, et al. (2026). Seismological characterization of Northern Hikurangi Margin Slow Slip Regions Associated with Normal Faults, Seamounts, and Seeps. *Journal of Geophysical Research: Solid Earth*, 131, doi:[10.1029/2025JB032916](#).

Acknowledgements

The time-dependent slow slip models were provided by Laura Wallace during a research visit to GEOMAR-Helmholtz Centre for Ocean Research Kiel. For methodological detail, please refer to Appendix [D](#).

4.1 Introduction

In this chapter, we analyse the earthquake catalogue, focal mechanisms derived in Chapter [3](#) in relation to SSEs, subducting seamounts, and surface observations of seeps.

Fluids play a critical role in SSE cycles; SSE occurrence has been linked to temporal variations in fluid pressures in the subducting slab and plate interface shear zone ([Audet et al., 2009](#)). Observations and modelling suggest that fluid pressure fluctuations can generate overpressures that modulate SSE recurrence by reducing the effective normal stress required for initiation on the plate interface (e.g., [Audet et al., 2009](#); [Kodaira et al., 2004](#); [Liu and Rice, 2007](#); [Warren-Smith et al., 2019](#)). Some studies propose that fluids in SSE source regions are periodically

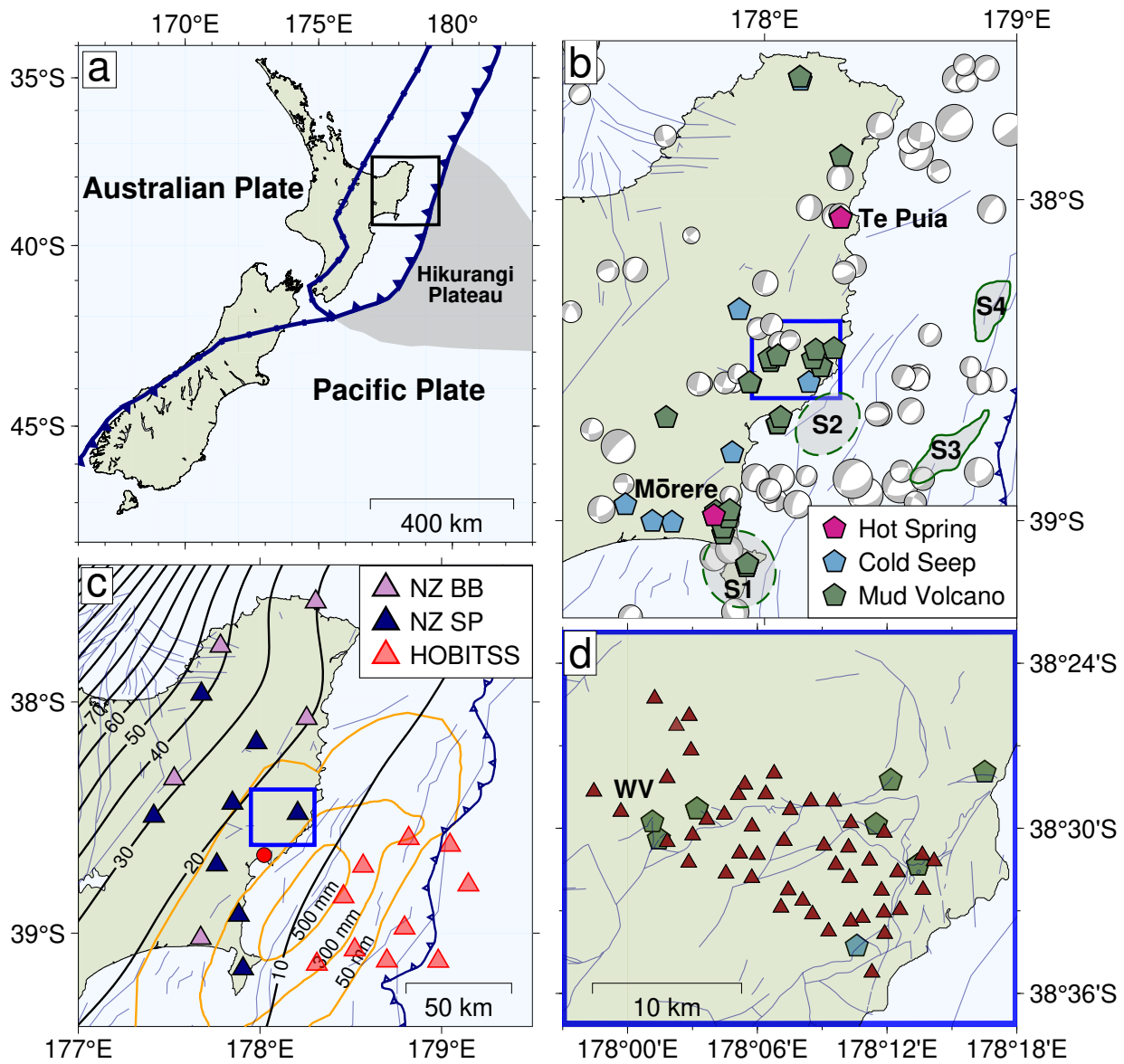


Figure 4.1: **a)** Tectonic map of Aotearoa New Zealand. Black box: study region as shown in (c). Blue lines: plate boundaries (Bird, 2003). **b)** Normal focal mechanisms from the ISC from 1965 to the present day, ranging from M_w 3.4 to 6.7 (Lentas, 2018; Lentas et al., 2019). S1 & S2: suggested seamounts (Chow et al., 2020; Bangs et al., 2023). S3 & S4: seamounts inferred from seismic reflection and magnetic data (Bell et al., 2010; Barker et al., 2018). S3: Papāku seamount. Orange line: 50 mm cumulative slip contour. Blue box: region shown in (d). Hot spring, seep, mud volcano data from Reyes et al. (2022). **c)** Seismograph network map of Raukūmara Peninsula. NZ BB: permanent broadband seismograph stations. NZ SP: permanent short period seismograph stations. HOBITSS: ocean-bottom seismometers used in the HOBITSS study (Todd et al., 2018). Navy lines: faults (Seebeck et al., 2022). Red circle: Gisborne. Orange contours: cumulative 50, 300, and 500 mm slow slip displacement contour 2002–2012 (Wallace, 2020). Black contours: plate interface depth (Williams et al., 2013). **d)** Dark red triangles: temporary seismograph network (NZ3D-FWI 3C). Blue lines: faults (Mazengarb and Speden, 2000). WV: Waimata Valley.

drained into upper-plate fracture networks via fault-valving processes (e.g., [Gosselin et al., 2020](#); [Ozawa et al., 2024](#); [Warren-Smith et al., 2019](#)), though the mechanisms and timescales of such drainage remain debated (e.g., [Nakajima and Uchida, 2018](#); [Ozawa et al., 2024](#)). Subducting seamounts generate complex fracture networks as the upper plate deforms to accommodate them ([Arai et al., 2017](#); [Arnulf et al., 2021](#); [Bangs et al., 2023](#)), enhancing fluid pressures at and above the plate interface (e.g., [Chesley et al., 2021](#); [Ellis et al., 2015](#)) and increasing stress downdip of the head of the seamount ([Sun et al., 2020](#)). Whether seamounts primarily act as seismic asperities that promote earthquake rupture (e.g., [Scholz and Small, 1997](#)) or as barriers that favour aseismic slip ([Mochizuki et al., 2008](#); [Wang and Bilek, 2011](#)) remains unresolved. SSE occurrence is likely controlled by a combination of fluid distribution, frictional and structural properties, and temperature conditions, influenced by seamount subduction and sediment distribution ([Wallace, 2020](#)). High-quality earthquake catalogues with low completeness magnitudes in SSE-prone regions are crucial for constraining how plate roughness and fluid pathways relate to seismicity and slow slip.

4.1.1 Tectonic background

The upper plate at the northern Hikurangi margin is in a state of extension ([Behboudi et al., 2023](#); [Henry et al., 2025](#); [Townend et al., 2012](#)), creating fractures that permit upward fluid migration. The subducting Pacific Plate has relatively thin sediment cover (≤ 1 km; [Barnes et al., 2020](#); [Ghisetti et al., 2016](#)) and a rough plate interface where seamounts protrude above the sediment ([Barnes et al., 2020](#); [Pedley et al., 2010](#)). The Hikurangi Plateau (Figure 4.1) forms the portion of the Pacific Plate that subducts at the Hikurangi Trough ([Mortimer and Parkinson, 1996](#)); it is considerably thicker than the 4–9 km thick oceanic crust subducting at the northeastern end of the Hikurangi-Tonga-Kermadec subduction system ([Davy et al., 2008](#); [Mochizuki et al., 2019](#)). The plateau is an ~11 km-thick large igneous province of tholeiitic basalt ([Bassett et al., 2023](#)), formed 98–118 Ma, with numerous 52–99 Ma alkali basalt seamounts protruding from its surface (Figure 4.1b; e.g., [Hoernle et al., 2010](#)). Subduction of the relatively buoyant plateau has caused uplift of the east coast of the North Island ([Litchfield et al., 2007](#)), resulting in a subaerial forearc.

Subducting seamounts enhance plate roughness, giving rise to variable stress and fluid pressure conditions and heterogeneous lithologies and frictional properties that have been linked with SSE recurrence ([Barnes et al., 2010](#); [Shaddox and Schwartz, 2019](#)). They also damage the upper plate, producing enhanced fracture networks that provide pathways for fluid migration ([Arai et al., 2017](#); [Bangs et al., 2023](#)). Several studies have characterised fluid pressure variations through the SSE cycle, observing an accumulation and release of fluid pressure before and during SSEs ([Warren-Smith et al., 2019](#); [Zal et al., 2020](#)). However, imaging seamounts at >10–15 km depth is beyond the limit of high-resolution active-source seismic reflection surveys.

In Chapter 3, we derived an earthquake catalogue of 3,071 earthquakes. We detected 86 focal mechanisms, of which 46 display normal faulting type mechanisms. In this chapter, we combine our catalogue with existing earthquake catalogues and focal mechanisms and consider the role of normal faulting in the shallow slow slip region, which hosts numerous SSEs and seeps. We then consider the relationship between SSEs and microseismicity that occurs during them.

4.2 Normal faults: links with SSEs, seamounts, and seeps

4.2.1 Normal faults in the subducting slab

Our focal mechanisms show numerous normal faulting earthquakes (depth uncertainty ≤ 2.5 km) penetrating the slab mantle at depths ≥ 15 km below the plate interface (Figure 4.2). Similar events have long been observed at the northern Hikurangi margin, including the 2007 M_w 6.6 Gisborne earthquake, which ruptured at ~ 40 km depth and triggered a SSE and mud volcano eruptions near Waimata Valley/Gisborne (Figure 4.3, [François-Holden et al., 2008](#)). Other historical M_w 5.5–6.5 normal-faulting earthquakes also occurred ≤ 40 km beneath the interface ([Bannister et al., 1989](#); [Anderson et al., 1990](#); [Webb and Anderson, 1998](#); [Doser and Webb, 2003](#); [Ekström et al., 2012](#)). Our catalogue samples comparable depths but smaller magnitudes, consistent with its shorter observation period.

Normal faulting in subducting slabs is commonly attributed to slab bending, although the

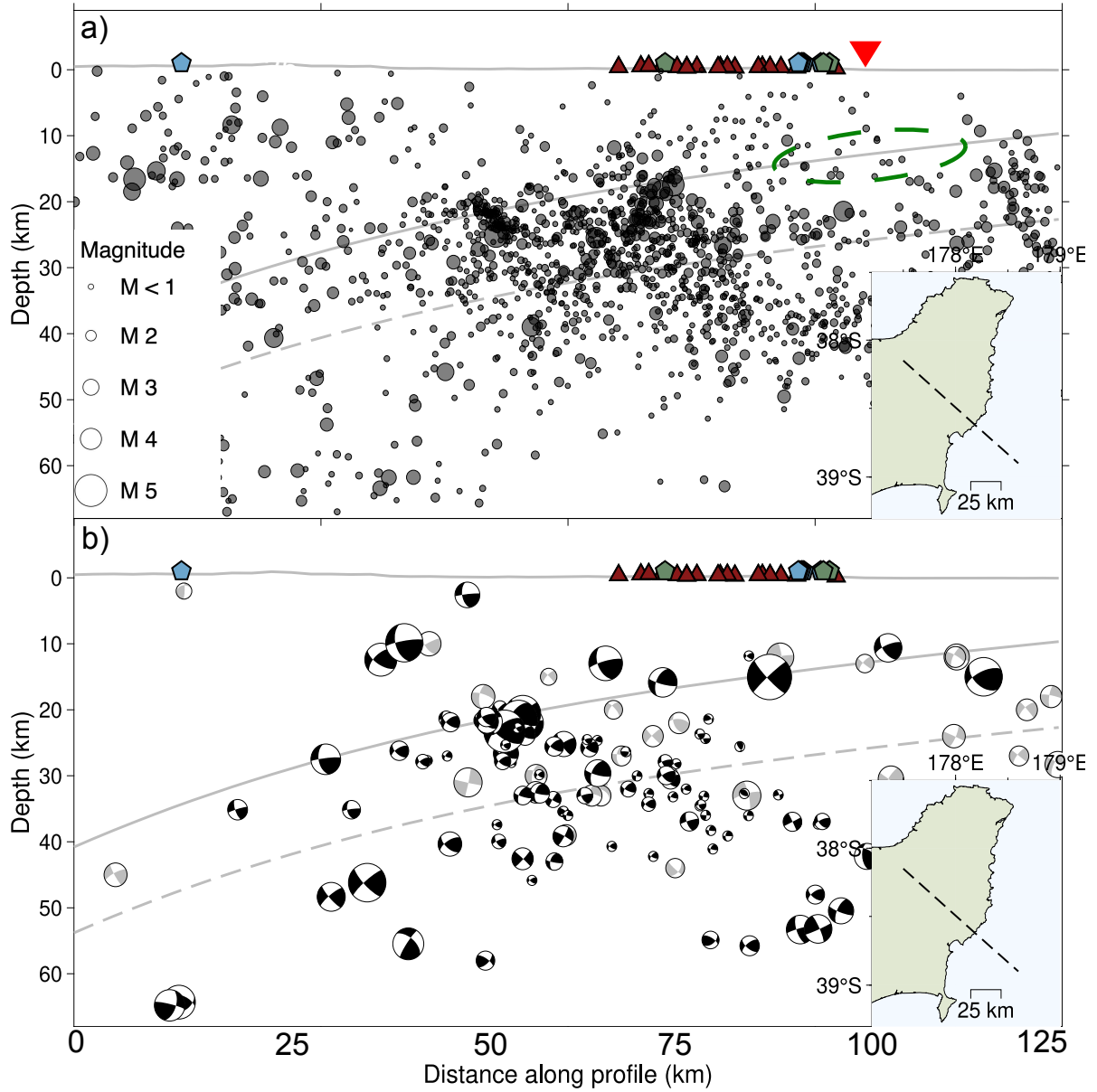


Figure 4.2: a) Circles: earthquakes from the combined catalogues of this study and the HOBITSS catalogue. (Yarce *et al.*, 2019). Dashed green ellipse: seamount proposed by Leah *et al.* (2022), as plotted in Figure 4.1. Inverted red triangle: coastline. b) Focal mechanisms: normal faults from previous studies (Lentas, 2018; Lentas *et al.*, 2019). Cross-section shows all normal faults (± 100 km) in catalogue projected relative to the slab at orientation shown on inset map. For cross-section at same orientation and width as used in Chapter 3, refer to Figure 4.3 Black focal mechanisms are from this study. Note that focal mechanisms from this study are plotted larger relative to the magnitude of historic events for visualisation purposes. Gray line: topography and plate interface. Dashed gray line: slab Moho (Yarce *et al.*, 2021; Bassett *et al.*, 2025).

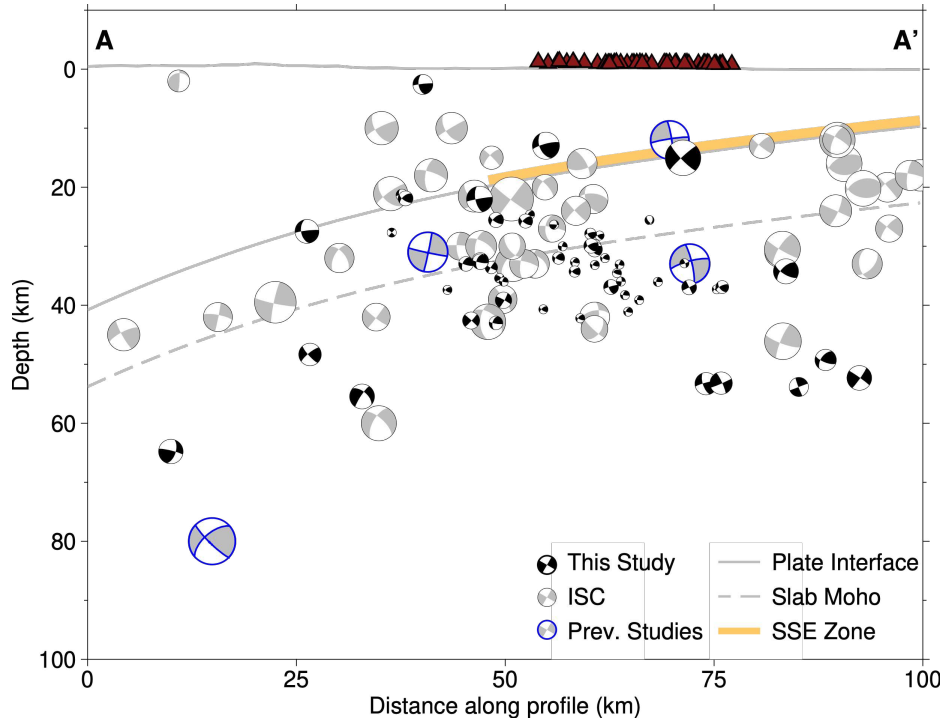


Figure 4.3: Cross section through study region showing newly derived and previously reported normal fault focal mechanisms. Gray lines: topography, plate interface, and slab Moho. Cross section width is ± 25 km. For profile orientation, please refer to Figure 3.15.

New Zealand Community Fault Model ([Seebeck et al., 2022](#)) shows fewer outer-rise faults than in many margins. Bending may be amplified ~ 100 km landward of the trench where slab dip steepens ([Barker et al., 2009](#); [Williams et al., 2013](#)). Uplift of buoyant subducting topography can also promote normal faulting (e.g., at the Ryukyu trench: [Arai et al., 2017](#)). Receiver function and seismic reflection data identify a subducted seamount beneath our study area (Figure 4.1, panel b), labelled S2; [Leah et al., 2022](#); [Bangs et al., 2023](#)), suggesting that localised uplift increases plate roughness and facilitates normal faulting.

Figure 4.4 shows a seismic reflection profile from the NZ3D offshore air-gun seismic dataset (for processing methodology, see [Bangs et al., 2023](#)), with the Pāpaku seamount at ~ 8 km depth. Downdip of the seamount, basement reflectors are offset with a normal sense of motion, suggesting the presence of a normal fault in the slab. Similarly steeply-dipping normal faults have been observed from seismicity and seismic reflection data at the Ryukyu trench, Japan, where it is suggested that localised uplift of subducted seamounts results in ≤ 1 km sub-vertical (70 – 80° dip) displacement downdip of the seamount ([Arai et al., 2017](#)). Several seafloor

seeps ([Watson et al., 2020](#)) overlie the interpreted normal fault and normal-faulting focal mechanisms in the slab (Figure 4.2/4.4). The co-location of normal faulting, seeps, and abrupt slab-dip changes indicates that both slab bending and seamount-related uplift drive faulting and enhance permeability.

Seismic reflection data suggests a lack of normal faulting in the outer rise ([Gase et al., 2023](#)), that could provide evidence for slab bending, though this is likely due to the very shallow slab dip in the offshore portion. The relatively buoyant Hikurangi Plateau may act to inhibit slab bending. There are however, historical observations of outer rise normal faulting ([Bannister et al., 1989](#); [Doser and Webb, 2003](#); [François-Holden et al., 2008](#)). At ~125 km landwards of the trench, the slab dip increases ([Williams et al., 2013](#)). We suggest that slab bending and localised uplift of topography are not mutually exclusive ideas; both may play a role in enabling fluid transfer from the slab lithosphere to the surface.

The east coast of the Raukūmara Peninsula (Figure 4.1c) hosts numerous cold springs, hot springs, and mud volcanoes ([Ridd, 1970](#); [Reyes et al., 2010, 2022](#); [Leighton et al., 2022](#)), reflecting upward fluid migration from the slab. Fluids are derived from a combination of subducting seawater, clay dehydration, serpentinisation, and mantle gas with higher proportions of dehydration and serpentinisation in SSE regions ([Reyes et al., 2022](#)); relative proportions of these fluid sources vary along-strike and with distance from the trench ([Ellis et al., 2015](#)). At the Te Puia and Mōrere hot springs (Figure 4.1c), thermal water samples show a higher mantle-derived helium gas signature ([Giggenbach et al., 1993](#); [Reyes et al., 2022](#)), suggestive of fluid pathways from the slab mantle to the surface ([Reyes et al., 2022](#)).

Isotopic evidence for mantle-derived fluids are observed at Te Puia and Mōrere hot springs (Figure 4.1), where temperatures of 71°C and 50°C respectively suggest more direct fluid conduits from the slab lithosphere ([Reyes et al., 2022](#)), consistent with deep fluid release inferred from regional tomography ([Eberhart-Phillips and Bannister, 2015](#); [Eberhart-Phillips and Fry, 2017](#)). Localised uplift from subducted seamounts likely damages the upper plate and further enhances slab-to-surface fluid pathways. Our observations of linear bands of seismicity show fluid pathways from the slab mantle to the plate interface. We also observe seismicity in the upper plate, particularly beneath surface seeps in Waimata Valley (Figure 4.1).

[Warren-Smith et al. \(2019\)](#); [Zal et al. \(2020\)](#) demonstrated that temporal variations in fluid pressure are likely associated with fluid release from the subducting slab and plate interface into the upper plate during SSEs. Small magnitude earthquakes beneath mud volcanoes and seeps are likely enabled by low effective stress from high pore fluid pressures. The damaged upper plate has a complex network of pathways, evidenced by geochemical data ([Reyes et al., 2010, 2022](#)), widespread seismicity, and varied focal mechanisms (Figure 4.2).

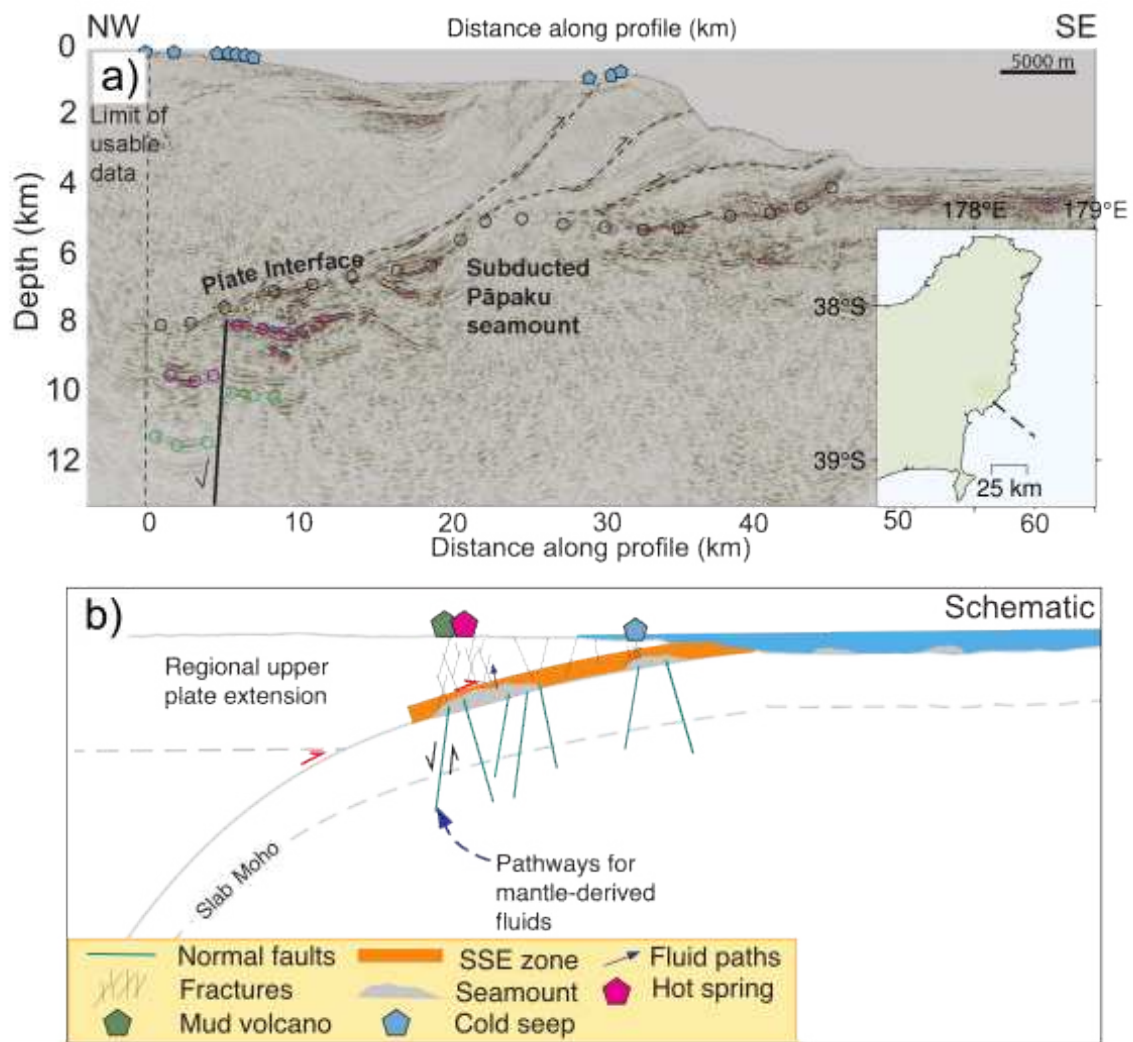


Figure 4.4: **a)** Seismic reflection profile from offshore dataset ([Bell et al., 2019](#)). Black dashed lines: thrust faults. Black solid line: interpreted normal fault. Open black circles: plate interface. Open purple and green circles: example basement reflector offsets caused by the normal fault. **b)** Schematic showing relationships between SSE zone, seamounts, fault and fracture networks, and seeps. Note the different scales in each panel.

Te Puia and Mōrere hot springs

Te Puia springs are located on an active landslide, where Late Cretaceous sediments are sliding eastwards at 55 mm/yr ([Mazengarb and Speden, 2000](#)). Discharge temperatures range from 54–69°C, with discharge rates of 8.7 L/s ([Pohatu et al., 2010](#)). At Te Puia and Mōrere hot springs (Figure 4.1c), thermal waters show elevated mantle-derived helium signatures ([Giggenbach et al., 1993](#); [Reyes et al., 2022](#)), with inferred mantle fractions of 24% and 41%, respectively ([Giggenbach et al., 1993](#)). Regional magnetic data suggest the presence of buried ophiolite bodies or subducted seamounts ([Hunt and Glover, 1995](#)). Subducting seamounts may facilitate sediment dewatering, and the ascending waters are modified by interaction with overlying rock and by mixing with surface meteoric waters ([Hunt and Glover, 1995](#); [Reyes et al., 2022](#)). Although evidence for hypothesised fluid pathways from the slab mantle to the surface is lacking ([Reyes et al., 2022](#)), this interpretation is consistent with regional-scale seismic tomographic models that suggest fluids are released at depth beneath the northern Hikurangi margin ([Eberhart-Phillips and Bannister, 2015](#); [Eberhart-Phillips and Fry, 2017](#)). Mantle-derived fluids appear to be especially abundant at Te Puia and Mōrere, where spring temperatures of 71°C and 50°C have been taken to suggest relatively direct fluid conduits from the slab lithosphere ([Reyes et al., 2022](#)). Localised uplift of subducted topography may damage the upper plate and enhance slab-to-surface fluid pathways (e.g., [Arai et al., 2017](#)). Small-magnitude earthquakes beneath mud volcanoes and seeps, likely enabled by reduced effective stress during fluid migration, are consistent with this interpretation. Records of use by local iwi and Hapū communities also indicate that these hot springs have remained active over at least several hundred years ([Pohatu et al., 2010](#)).

Assuming a geothermal gradient of $11 \pm 2^\circ\text{C}/\text{km}$ along the interface ([Reyes et al., 2022](#)), surface fluid expressions above a 13–21 km-deep slab correspond to interface temperatures of 143–231°C, spanning the onset of serpentinitisation ($\sim 185^\circ\text{C}$). Clay dehydration begins at >10 km depth ([Heise et al., 2013](#); [Ellis et al., 2015](#)). The observed seismicity is therefore consistent with fluid release by serpentinitisation, clay dehydration, mantle degassing, and expulsion of subducting seawater, especially where seamount subduction promotes normal faulting. Mōrere hot springs and Waimata Valley mud volcanoes are co-located with purported

subducted seamounts ([Leah et al., 2022](#); [Chow et al., 2022a](#); [Bangs et al., 2023](#)). The Ruatōria debris avalanche (discussed further in Chapter 5) may be due to the subduction of a seamount beneath the region of the Te Puia hot spring ([Collot et al., 2001](#)), though a seamount beneath the Te Puia hot spring has not been directly imaged.

Waimata Valley mud volcanoes

In December 2018, a mud volcano erupted in Waimata Valley (Figure 4.5), releasing 16,900 m³ of mud ≤ 50 m into the air ([Leighton et al., 2022](#)). Though the Waimata Valley mud volcanoes are cooler than the hot springs, they are particularly active relative to other east coast seeps and mud volcanoes. A normal fault that cuts across the eruption site separates the Tolaga and Mangatu formation mudstones ([Leighton, 2020](#)). A high fluid flux through conduits from the slab likely exists here too, demonstrated by the widespread seismicity beneath Waimata Valley (Figure 4.1) and mud diapirism ([Ridd, 1970](#); [Darby et al., 2000](#)). The Waimata Valley mud volcanoes also erupted in 1911 and 1930, demonstrating sustained fluid conduits on at least decadal timescales.

Prior to the eruption, during the deployment of the Gisborne network, local residents reported doming of the site that would become the eruption vent ([Leighton, 2020](#); [Leighton et al., 2022](#)). The eruption occurred in December 2018, resulting in the formation of a lake to the north-west (Figure 4.5), an arcuate vent, and mudflows. At the site, surface ruptures were reported following the Te Araroa earthquake in September 2016, suggesting regional links between earthquake occurrence and mud volcano activity.

4.3 Slow slip events and seismicity

4.3.1 Slow slip events from GNSS daily displacement data

Figure 4.6 shows three-component GNSS station data and 4.7 shows daily east-component GNSS time series. Displacement during the SSEs is most clearly observed on the east-component, which is aligned with the direction of convergence between the Australian and Pacific Plates, thus demonstrating their tectonic origin. We analysed all GNSS stations along

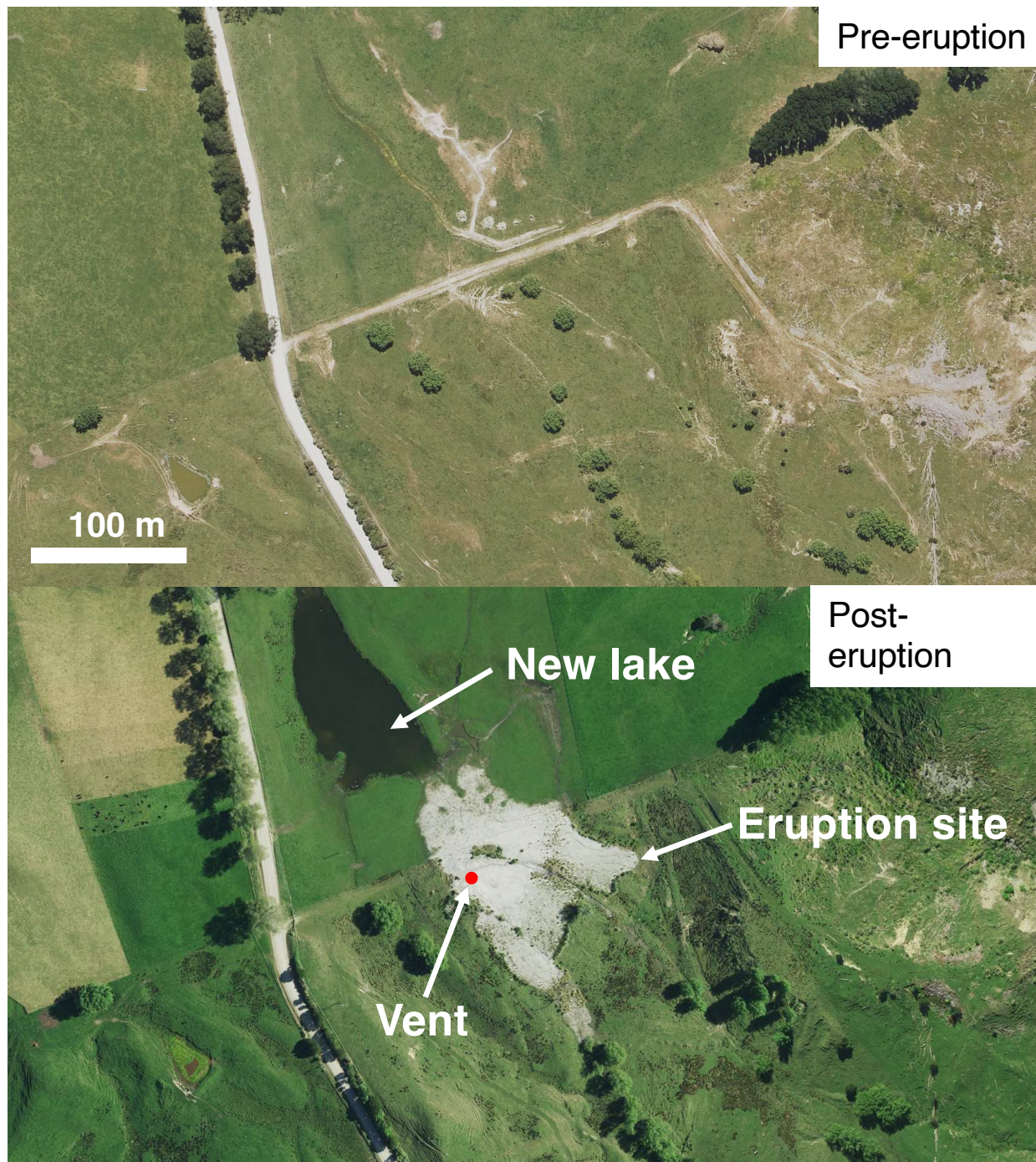


Figure 4.5: Satellite image of December 2018 mud volcano eruption site. **Upper panel)** pre-eruption. Orthophotography was captured by Gisborne District Council during the 2017–2019 flying season with 0.3 m resolution. **Lower panel)** Post eruption, showing the vent site (red circle), eruption site, and newly formed lake to the north-west. Orthophotography was captured by the Gisborne District Council during the 2024–2025 flying season with 0.1 m resolution. Data are publicly available through the Toitū Te Whenua Land Information New Zealand service.

the Raukūmara Peninsula. We calculated the mean displacement between the start and end of ten day sliding windows in one-day increments, to define the longer-term displacement trend at each station. We then identified turning points in these smoothed time series, defining a SSE as an eastwards-propagating excursion superimposed on the broader westwards trend (e.g., [Woods et al., 2024](#)), and calculated the displacement during this period for each station. The first SSE is observable on seven stations; the second SSE is observable on five stations. These SSEs are also recorded in the longer-term catalogue of [Perez-Silva et al. \(2025\)](#). Following [Perez-Silva et al. \(2025\)](#) we plot the earliest start and latest end times for each SSE. The stations on which these signals are observed are spatially clustered for each SSE, allowing us to place tentative constraints on their slip area.

Increased seismicity during SSEs is widely documented at the northern Hikurangi margin (e.g., [Delahaye et al., 2009](#); [Bartlow et al., 2014](#); [Jacobs et al., 2016](#); [Todd and Schwartz, 2016](#); [Yarce et al., 2019](#)), and long-term catalogs show that some slow slip patches consistently coincide with elevated seismicity ([Hughes et al., 2021](#)). Most prior work has examined the larger SSEs that recur every 1–2 years (e.g., the 2014 SSE that recorded >200 mm of slip; [Yarce et al., 2019](#); [Todd et al., 2018](#)). However, uncertainties remain regarding the timing and nature of SSE-seismicity interactions ([Chen et al., 2018](#); [Dascher-Cousineau and Bürgmann, 2024](#)), partly because the SSE catalogue spans only ~20 years. More frequent, smaller SSEs (e.g., [Perez-Silva et al., 2025](#)) therefore offer valuable opportunities to evaluate these relationships, particularly when captured by dense temporary networks. During SSE-1 and SSE-2, slow slip rates peak mid-way through each event, and earthquake rates (10-day windows, 1-day increments) reach maxima contemporaneously (Figure 4.9).

Although an increase in seismicity is observed in May 2018, no associated slow slip or spatial clustering is detected (Figure D.4), so our analysis focuses on SSE-1 and SSE-2. The increased rate in seismicity in May is widespread over a ~50 km region, while the increased seismicity during SSEs is more tightly clustered. The May increase in seismicity is in the plate interface and upper plate, and may be associated with upper plate slumping processes (e.g., [Mazengarb and Speden, 2000](#); [Litchfield et al., 2022](#)).

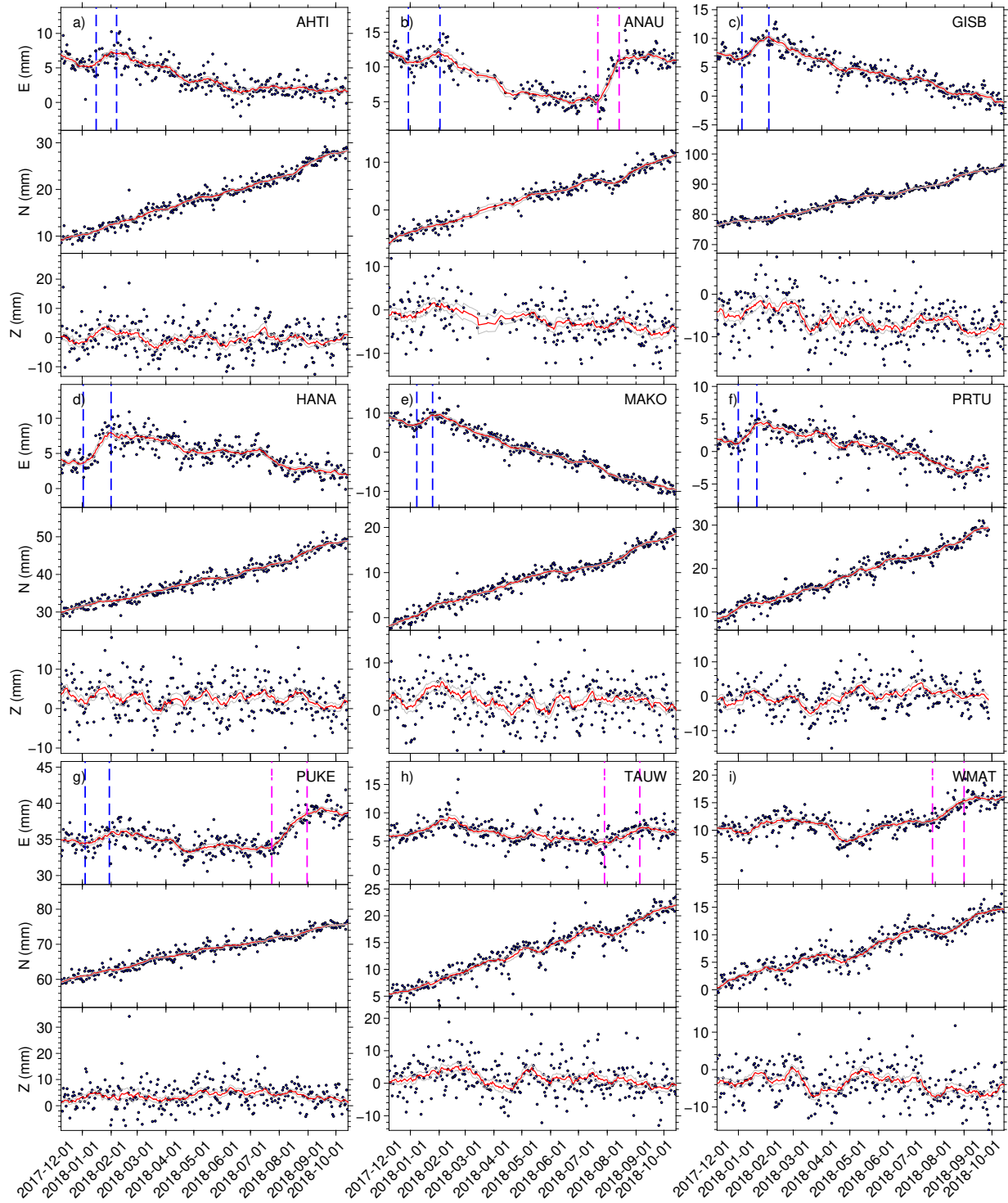


Figure 4.6: Three-component GNSS data. For station locations, refer to Figure 4.7. Red lines: ten-day averaged displacements.

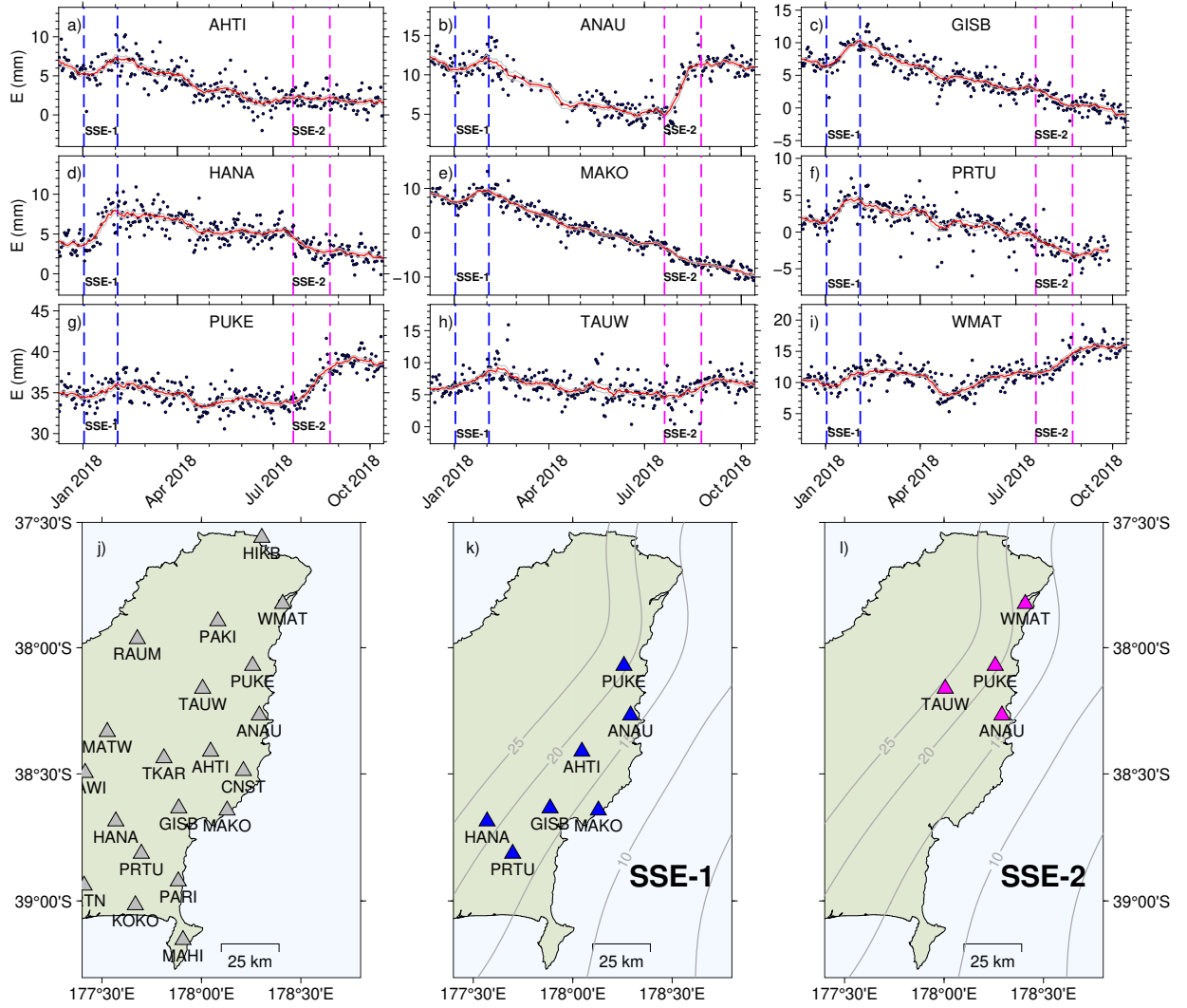


Figure 4.7: a-i) East-component GNSS station positions. Dashed blue lines: interpreted SSE in January 2018, referred to as SSE-1. Dashed magenta lines: interpreted SSE in July/August 2018, referred to as SSE-2. Note the variable y-axes, as here we report GNSS positions. j) Gray triangles: GNSS stations used in this study. k) Blue triangles: GNSS stations picked in SSE-1. l) Magenta triangles: GNSS stations picked in SSE-2.

4.3.2 Time-dependent slip models

The SSE durations observed from the GNSS displacement data provide preliminary bounds and parameters for the inversion of GNSS data for time-dependent slip models. For methodological detail, refer to Appendix D.

Time-dependent slip models provide information on how slip distribution evolves through the SSE cycle. Here we analyse the time-dependent slip models in ten-day increments, during SSEs of ~ 30 days, to investigate how the slip distribution and magnitude evolves and how this relates to the seismicity distributions. We calculate the summed moment release in ten-day intervals on the slip interface, by calculating the sum of the slip area multiplied by displacement for each cell in the model.

During the deployment of the temporary seismograph network, two SSEs were recorded (Figure 4.8). The first, referred to henceforth as SSE-1, had a magnitude of M_w 6.22, with a duration of 42 days. SSE-1 was deeper than most SSEs previously observed in this region, at plate interface depths of $\sim 10\text{--}30$ km, with the majority of slip onshore. The second (SSE-2) had a magnitude of M_w 6.40, with a duration of 53 days. The slip distribution of SSE-2 was similar to previous SSEs recorded in this region, with the majority of slip occurring offshore at a plate interface depth ≤ 20 km. SSE-3 had a magnitude of M_w 6.13 with a duration of 81 days. The majority of slip occurred at a plate interface depth ≤ 15 km, consistent with the majority of previous shallow, offshore SSEs. SSE-3 began during the deployment of the temporary network but continued slipping after the seismometer recovery, so we primarily focus on SSE-1 and SSE-2.

4.3.3 Characterising SSE properties

We calculate the stress drop for each SSE (Brune, 1970), by first calculating r , the source radius, using

$$r = \sqrt{\frac{M_0}{\mu\pi D}}, \quad (4.1)$$

where M_0 is the seismic moment, μ is the shear modulus (assumed to be 30 GPa), and D is the

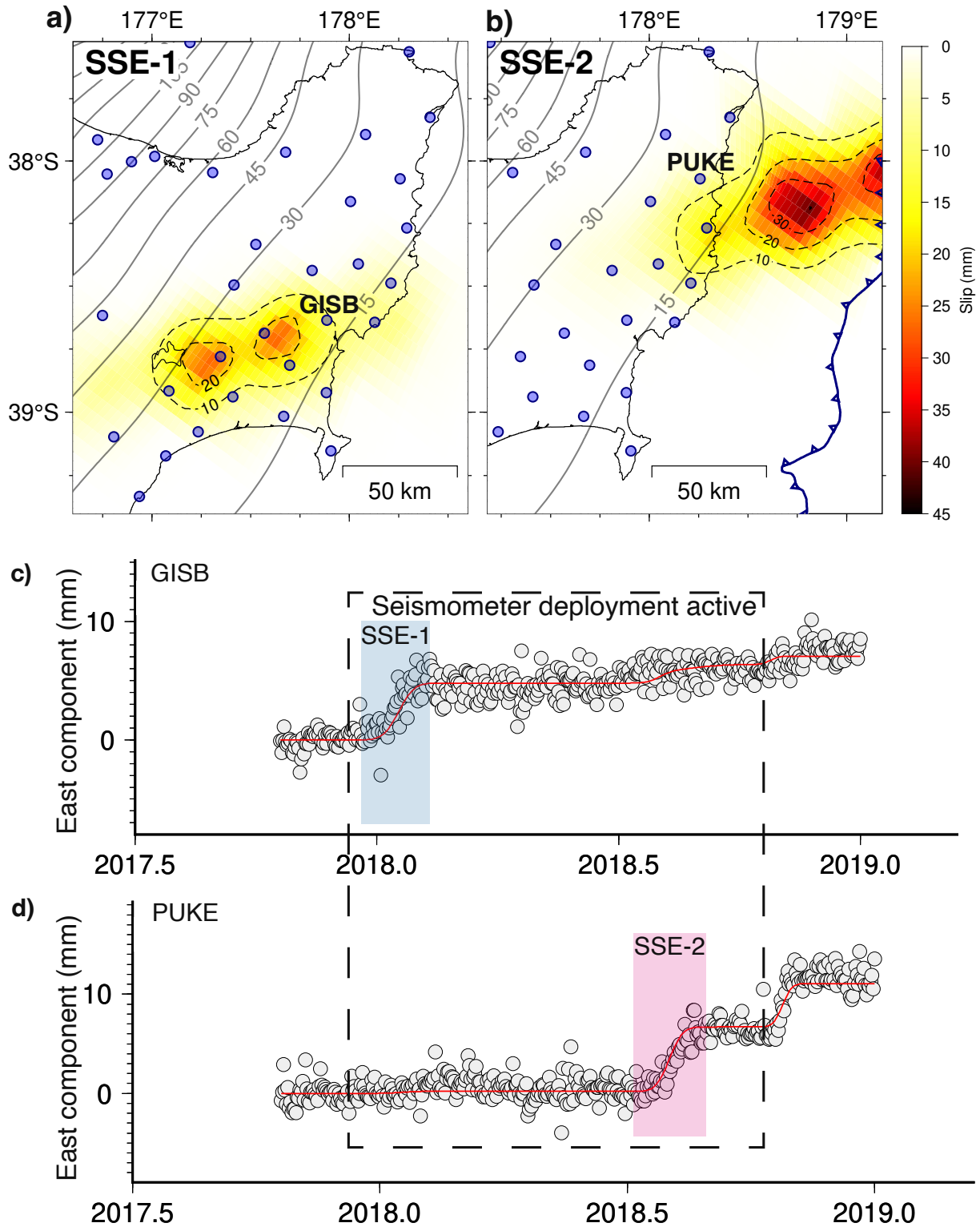


Figure 4.8: a) Slow slip patch for SSE-1 (recorded 2017-12-21 to 2018-02-01). Black dashed lines: slow slip contours (mm). Blue circles: GNSS stations. Gray lines: plate interface depth (km) (Williams *et al.*, 2013). b) As in panel a, for SSE-2 (recorded 2018-07-10-2018-09-01). Navy line: trench. c) Eastwards displacement (mm) for station GISB (see station location in panel a). GNSS time series model (red line) fit to data (circles). Dates are in decimal years. For three-component data, refer to supporting information. Shaded blue: SSE-1 duration. Black dashed line: temporary network deployment period analysed. d) Eastwards displacement (mm) for station PUKE (see station location in panel b). Shaded pink: SSE-2 duration.

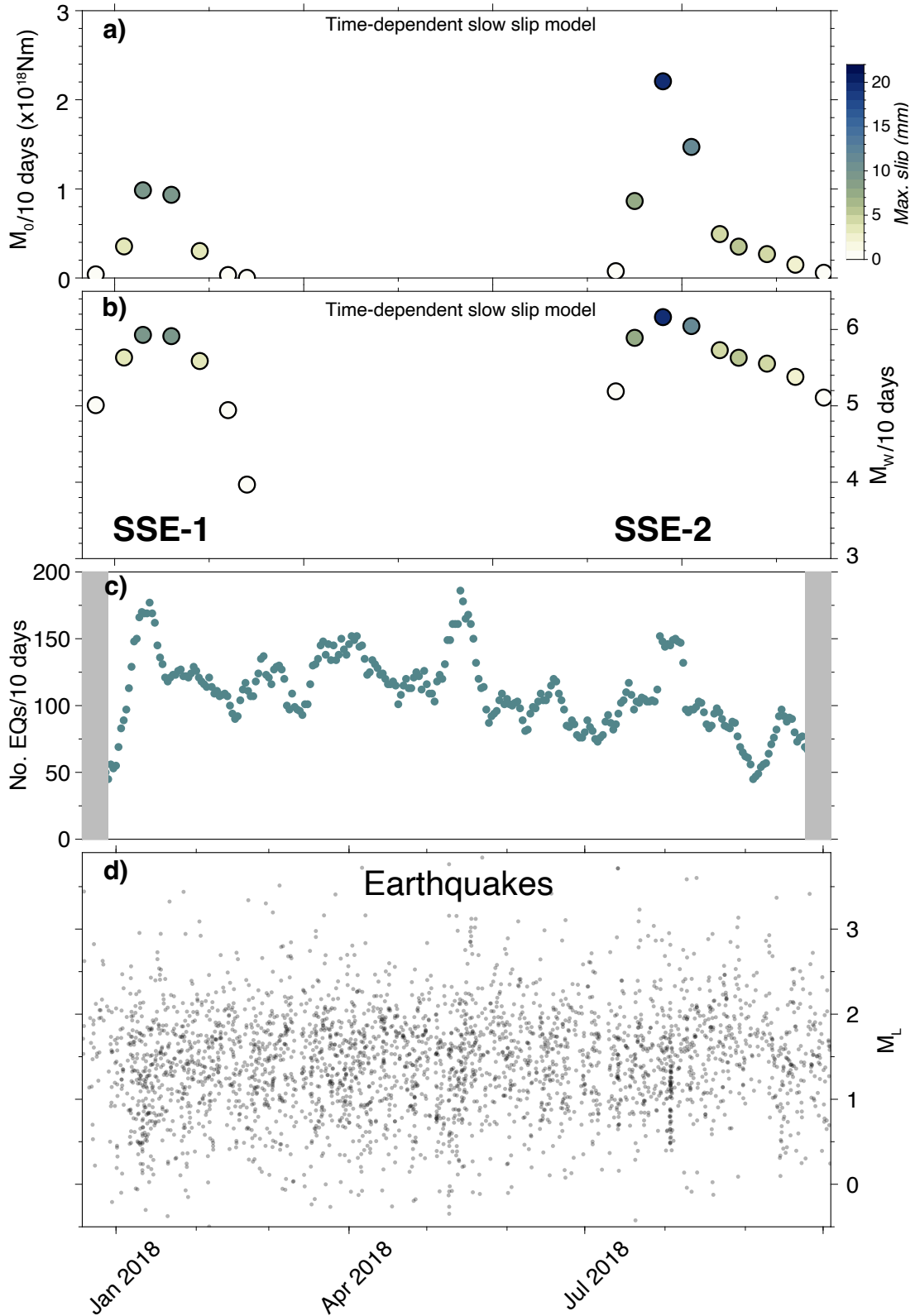


Figure 4.9: a) Seismic moment derived from the time-dependent slow slip model in ten-day slices, assuming a shear modulus of 30 GPa. b) Moment magnitude derived from the time-dependent slow slip model in ten-day slices. c) Number of earthquakes per ten days in sliding windows with increments of one day. First and last ten days of the catalogue are shaded in gray. d) Earthquake magnitudes through time.

mean slip (m). The slip inversions (Appendix D) provide slip per user-defined cell on the plate interface. We calculate the mean slip using the slip measurements on all cells with non-zero slip on the plate interface. The source area is approximated by a circle. Therefore the stress drop, $\Delta\sigma$ is

$$\Delta\sigma = \frac{7\pi\mu D}{16r}. \quad (4.2)$$

Stress drops for SSEs are typically an order of magnitude smaller than for earthquakes. Details are documented in Table 4.1. For completeness, we also list SSE-3 and SSE-4 from the broader time-dependent slip models, although our seismic analysis focuses on SSE-1 and SSE-2 during the temporary deployment.

Table 4.1: Summary of Slow Slip Event (SSE) parameters.

	Dates	Duration (days)	Area (km^2)	M_w	M_0 (Nm)	Mean slip (mm)	Stress drop (MPa)	Max. slip (mm)
SSE-1	21/12/17 - 01/02/18	42	3.34×10^4	6.22	3.53×10^{18}	2.7	0.00096	35.7
SSE-2	10/07/18 - 01/09/18	53	1.07×10^4	6.40	5.95×10^{18}	4.4	0.00152	53.4
SSE-3	21/07/18 - 10/10/18	81	6.17×10^3	6.13	2.00×10^{18}	8.1	0.00653	45.0
SSE-4	12/10/18 - 13/11/18	33	1.06×10^4	6.44	5.83×10^{18}	13.8	0.00850	54.1

4.3.4 Seismicity during SSE-1 and SSE-2

Increased seismicity rates during SSEs have been observed previously at the northern Hikurangi margin (e.g., [Delahaye et al., 2009](#); [Bartlow et al., 2014](#); [Jacobs et al., 2016](#); [Todd and Schwartz, 2016](#); [Yarce et al., 2019](#)). [Hughes et al. \(2021\)](#) demonstrated with a long-term SSE and seismicity catalogue that some SSEs are associated with increased seismicity. Larger magnitude SSEs that occur every 1–2 years have been the focus of previous seismicity and SSE studies at the northern Hikurangi margin (e.g., the 2014 SSE that recorded >200 mm of slip; [Yarce et al., 2019](#); [Todd et al., 2018](#)). However, temporal links between SSEs and seismicity are uncertain ([Chen et al., 2018](#); [Dascher-Cousineau and Bürgmann, 2024](#)), particularly due to the short (~20

year) SSE catalogue. More frequent but smaller-magnitude SSEs (e.g., [Perez-Silva et al., 2025](#)) can therefore provide fresh scope to analyse the relationship between SSEs and seismicity during deployments of dense temporary networks, such as that used here.

We extract time-dependent slow slip models throughout the entire observation period in ten-day slices. Slow slip rates peak approximately halfway through the SSEs. For the same period, we calculate the number of earthquakes per ten days, in sliding windows with increments of one day (Figure 4.9). Seismicity rates reach their maximum at the same time as the maximum slow slip rate.

A consistent delay between the onset of slow slip and the rise in seismicity is observed (Figure 4.9), matching patterns reported in other subduction zones (e.g., [Nakajima and Uchida, 2018](#)). This lag likely reflects fault-valving processes in which fluids are episodically transferred into the upper plate along fracture networks or the megathrust ([Sibson, 1990](#); [Warren-Smith et al., 2019](#)). Because our catalogue includes both interface and non-interface earthquakes, the elevated seismicity implies that SSE-induced stress or fluid changes extend beyond the modelled slow slip source region.

SSE-related seismicity is further characterised by pronounced spatiotemporal clustering during the SSE. We identify 54 earthquakes separated by <30 s and located within <10 km in depth and $<0.1^\circ$ laterally; manual inspection confirms they are distinct events. Such tight groupings may reflect stress transfer or fluid redistribution (e.g., [van der Elst and Brodsky, 2010](#); [Shaddox and Schwartz, 2019](#); [Hughes et al., 2021](#)). Dynamic triggering by teleseismic waves is unlikely to be the cause of increased spatiotemporal clustering, since no $M_w > 5.5$ earthquakes in the ISC catalogue occurred at times that would place their teleseismic surface-wave arrivals within the relevant triggering windows ([Lentas et al., 2019](#)), assuming surface wave speeds of 2.5–5 km/s ([Bormann et al., 2012](#); [Peña Castro et al., 2019](#)).

SSE-1 is associated with an updip concentration of earthquakes on the megathrust and within the subducting slab (Figure 4.10). This pattern resembles the 2014 SSE ([Shaddox and Schwartz, 2019](#)) and may be driven by pore-pressure increases from slab-derived fluids ([Nakajima and Uchida, 2018](#)) and/or by stress loading on faults already close to failure. The updip location also suggests the influence of plate interface roughness and subducted topography, which can

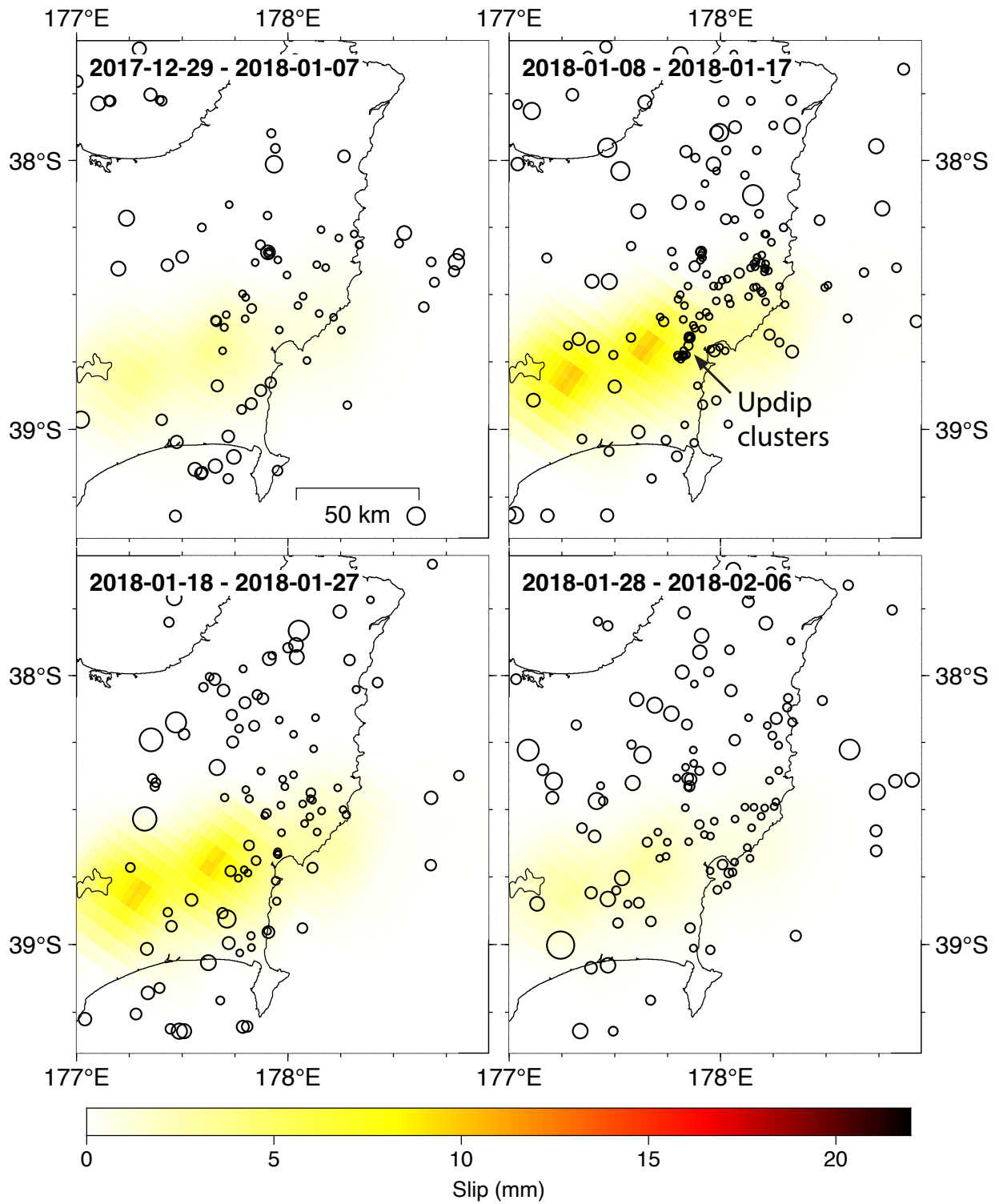


Figure 4.10: Ten-day windows of slow slip and seismicity (circles) distributions during SSE-1.

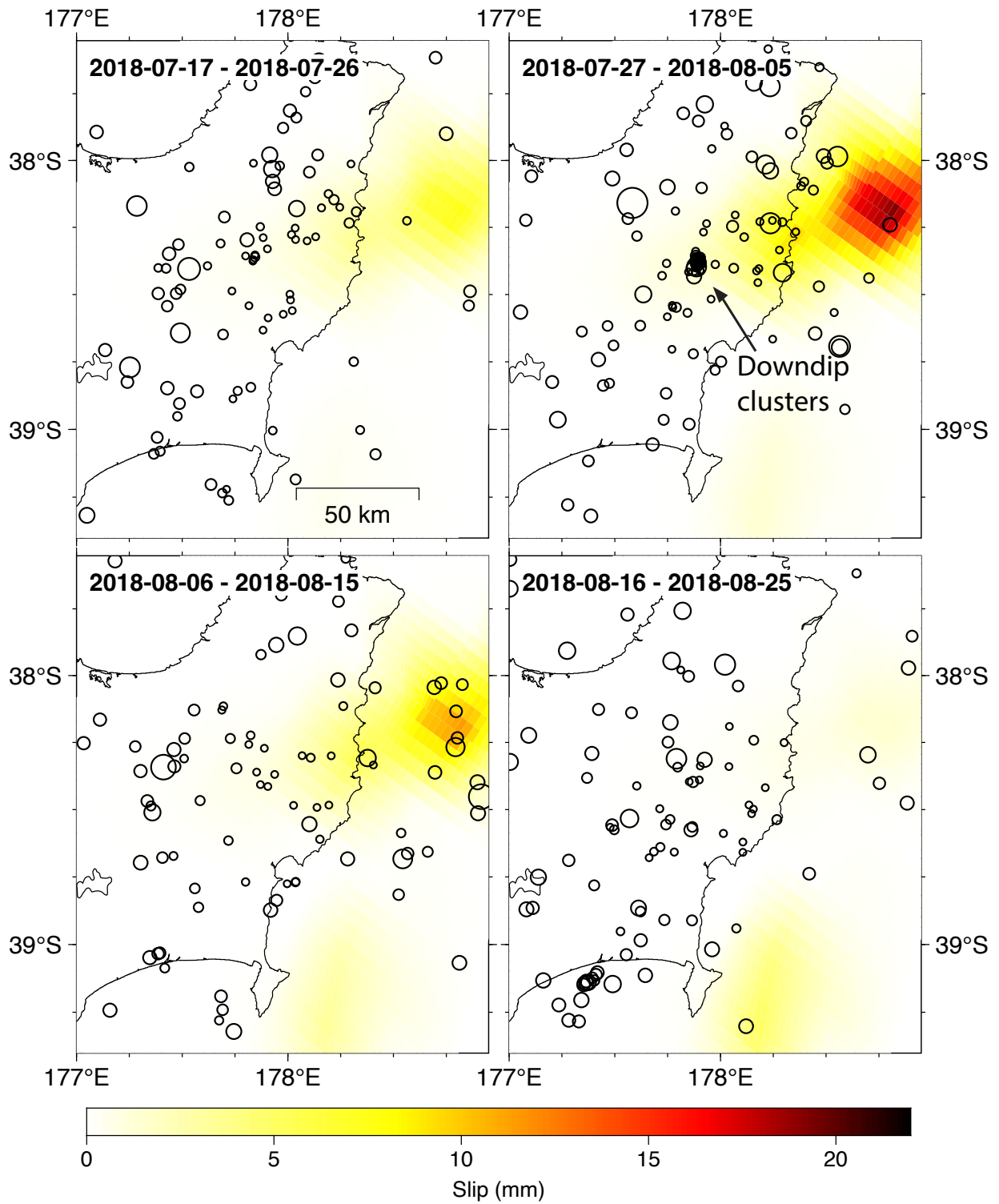


Figure 4.11: Ten-day windows of slow slip and seismicity distributions during SSE-2.

enhance fluid availability (*Bangs et al.*, 2023; *Yarce et al.*, 2019).

In contrast, SSE-2 produces a localised downdip cluster at 18–22 km depth (Figure 4.11), similar to patterns observed at the Hikurangi margin and other subduction zones (e.g., *Hughes et al.*, 2021). Earthquakes in this downdip cluster exhibit thrust-type mechanisms, consistent with megathrust slip, and the cluster recurs during both SSE-1 and SSE-2 despite differing slow-slip locations. Its persistence indicates that this segment of the interface is highly sensitive to regional stress changes. Similarly, a downdip cluster of seismicity occurs approximately 20 days after the onset of SSE-3, and continues for 20 days (Figure D.5). The cluster of earthquakes during SSE-3 is at ~ 20 km depth. We speculate that seismicity recurring at ~ 20 km depth may be due the presence of fluids at this depth that lower the effective stress, thus resulting in seismicity clusters during SSEs. We note however, that the 20 km plate interface depth region is well-resolved by our earthquake catalogue, and a lack of seismicity during SSEs at shallower and deeper depths may also be associated with poorer network coverage.

Taken together, these observations show that SSE-related seismicity at northern Hikurangi reflects a combination of stress transfer, fault-valving, and fluid migration, with their relative influence varying across the heterogeneous plate interface. Prior studies likewise report both updip (*Shaddox and Schwartz*, 2019) and downdip (*Hughes et al.*, 2021) clustering, underscoring the importance of SSE depth, magnitude, and local structure in governing the seismic response. Previous studies that have analysed the relationships between SSEs and variations in seismicity have focused on larger-magnitude SSEs, such as the M_w 6.8 2014 northern Hikurangi SSE (*Wallace et al.*, 2016). This study demonstrates that the smaller-magnitude SSEs result in similar seismicity patterns to those observed during larger-magnitude SSEs, demonstrating the value in studying the Hikurangi Margin in more detail, consistently through time, rather than just targeting the less frequent, larger SSEs. Longer-term SSE catalogues characterising their full depth extent, together with improved imaging of fluid distributions (e.g., V_P/V_S) and fluid migration rates (e.g., *Ozawa et al.*, 2024), will be crucial for constraining the dominant mechanisms.

From this study alone, we cannot discern the relative contributions of stress conditions, fault-valving and fluid migration phenomena, or a combination of these processes as a cause of

increased seismicity during SSEs. Repeating earthquake clusters reported in [Hughes et al. \(2021\)](#) during SSEs were variably located downdip, south of, or between slow slip patches. Conversely, [Shaddox and Schwartz \(2019\)](#) observe seismicity clusters updip of the slow slip patch. The complex interplay of a heterogeneous plate interface and variation in SSE characteristics (e.g., depth and magnitude) likely results in the varied seismicity response during SSEs ([Shaddox and Schwartz, 2019](#); [Hughes et al., 2021](#)). Future analysis of longer-term SSE and earthquake catalogues that bound the updip and downdip region of SSEs, with a particular focus on variations in V_P/V_S as a proxy for fluid distribution, as well as further investigation of the rate of fluid migration (e.g., [Ozawa et al., 2024](#)), will help to constrain the relative importance of the aforementioned mechanisms.

4.3.5 Fluids and SSEs

The majority of slow slip occurs at plate-interface depths shallower than ~20–30 km. Farther landward, V_P/V_S decreases ([Eberhart-Phillips and Bannister, 2015](#)), implying reduced fluid content and/or lower fracture connectivity. Consistent with this, surface seeps and slab normal faulting both cease ([Reyes et al., 2022](#), Figure 4.2). Plate roughness promotes fluid release and focusing via slab-penetrating faults, sediment and clay dehydration, and serpentinisation. These fluids, together with heterogeneous stresses from seamount subduction, likely facilitate shallow SSEs. The deeper slow slip identified in this study appears less frequent, perhaps because fluids escape at shallower depths or because higher temperatures (350°C, [McCaffrey et al., 2008](#)) promote more continuous creep; this is consistent with the Manawatū and Kapiti SSEs at the southern Hikurangi margin that have been shown to occur at 15–50 km depth every 4–5 years ([Wallace, 2020](#)).

In summary, subducted seamounts at the northern Hikurangi margin increase plate roughness, damage the upper plate, and modulate fluid supply. Localised uplift promotes normal faulting and enhances permeability along slab-penetrating faults, providing pathways for deep fluids to reach the surface, consistent with mantle-derived signatures at Te Puia and Mōrere hot springs.

4.4 Conclusions

Two slow slip events were detected with magnitudes M_w 6.2 and M_w 6.4 and durations of 6–8 weeks. SSE-1 was unusually deep, occurring onshore at plate-interface depths approaching 30 km, whereas SSE-2 was more consistent with previous slip observations, with slip concentrated offshore. Time-dependent slow slip modelling shows that seismicity rates increase during both events, with peak earthquake rates coinciding with maximum slow slip rates.

We derive 86 focal mechanisms from our catalogue, including 44 normal faulting earthquakes that define vertically aligned clusters within the slab lithosphere at depths ≥ 25 km below the plate interface. These normal faults, co-located with a purported subducting seamount, likely act as fluid conduits from the slab mantle toward the plate interface and upper plate. Slab bending and localised uplift around subducting seamounts promote normal faulting, increasing plate roughness and permeability and providing pathways for deep, mantle-derived fluids observed at surface seeps and springs. The downdip cessation of slab normal faulting and surface seeps coincides with the 20–30 km downdip limit of shallow SSEs, suggesting that insufficient fluids are available at greater depths to sustain slow slip. Together, these results indicate that the rough subducting plate interface at northern Hikurangi exerts strong control on forearc dewatering and the conditions required for SSE genesis.

Local earthquake tomography of the northern Hikurangi margin: plate roughness, fluid distributions, and controls on slow slip

The contents of this chapter are in review at the Journal of Geophysical Research: Solid Earth.

Woodward, A., I.D. Bastow, R. Bell, M. Paulatto, s. Henrys, K. Jacobs (*in review*). Local earthquake tomography of the northern Hikurangi margin: plate roughness, fluid distributions, and controls on slow slip. *Journal of Geophysical Research: Solid Earth*.

5.1 Introduction

5.1.1 Overview

Slow slip events (SSEs) are an important mode of strain release at subduction zones, but the factors that control where they occur, including the depth to which they persist, are still not fully understood. Globally, the majority of SSEs globally occur downdip of the seismogenic zone and have been attributed to transitional frictional properties. At some margins, including the northern Hikurangi margin, SSEs occur at shallower depths within the seismogenic zone ([Schwartz and Rokosky, 2007](#)). Fluids are widely thought to play a first-order role in shallow SSE genesis because they control effective stress (e.g., [Audet et al., 2009](#); [Ellis et al., 2015](#); [Sun et al., 2020](#)). Fault zone lithology and plate interface roughness may influence how fluids are trapped and released ([Barnes et al., 2020](#)). The northern Hikurangi margin is well-suited to investigating shallow SSE controls because it has a subaerial forearc and short trench-to-coastline distance ([Reyners et al., 1999](#)). Here, SSEs have been attributed to transitional

frictional properties, high fluid pressures and a rough plate interface (e.g., [Schwartz and Rokosky, 2007](#); [Shaddox and Schwartz, 2019](#)).

Most SSEs at the northern Hikurangi margin occur offshore at plate interface depths ≤ 20 km and occur every one to two years (Figure 5.1b). Smaller-magnitude SSEs have also been observed to depths of up to 35 km ([Wallace, 2020](#); [Woodward et al., 2026a](#)) and towards the central Kaimanawa region ([Wallace and Eberhart-Phillips, 2013](#)). This along-dip variation raises two questions: 1) Why does slow slip occur at such shallow depths? 2) What controls its downdip limit?

Previous tomographic studies have focused mainly on the offshore region (e.g., [Mochizuki et al., 2021](#); [Yarce et al., 2021](#); [Bassett et al., 2025](#)), using the permanent seismograph network (e.g., [Eberhart-Phillips and Bannister, 2015](#); [Chow et al., 2022b](#)) together with various temporary offshore ocean-bottom seismometer deployments (e.g., [Yarce et al., 2021](#); [Mochizuki et al., 2021](#)). Here we use local earthquake tomography to derive V_P and V_P/V_S models for the deeper part of the shallow slow slip region, where smaller and potentially less frequent SSEs occur, and to examine how along-dip variations in velocity structure relate to plate roughness and fluid distribution. We make use of a dense temporary onshore network deployed from December 2017 to October 2018 near Gisborne, combined with older data from the national permanent network and temporary Hikurangi Ocean Bottom Investigation of Tremor and Slow Slip (HOBITSS) network, deployed offshore from May 2014 to June 2015 (Figure 5.1c [Yarce et al., 2019](#)), to derive a localised, high-resolution tomographic model centred on the deeper part of the slow slip patch and including the region of greatest cumulative slow slip. In particular, we test whether high V_P anomalies support the presence of speculated seamounts beneath the onshore margin, and whether along-dip variation in V_P/V_S provides evidence for changes in slab hydration to help explain the depth distribution of SSEs.

5.1.2 Tectonic and geological background

At the Raukūmara Peninsula (Figure 5.1b), the Pacific Plate subducts obliquely beneath the Australian Plate at ~ 50 mm yr⁻¹ ([Wallace et al., 2012a](#)). In this region, SSEs recur every 6–24 months (e.g., [Perez-Silva et al., 2025](#); [Michel et al., 2025](#)). Compared with the central and

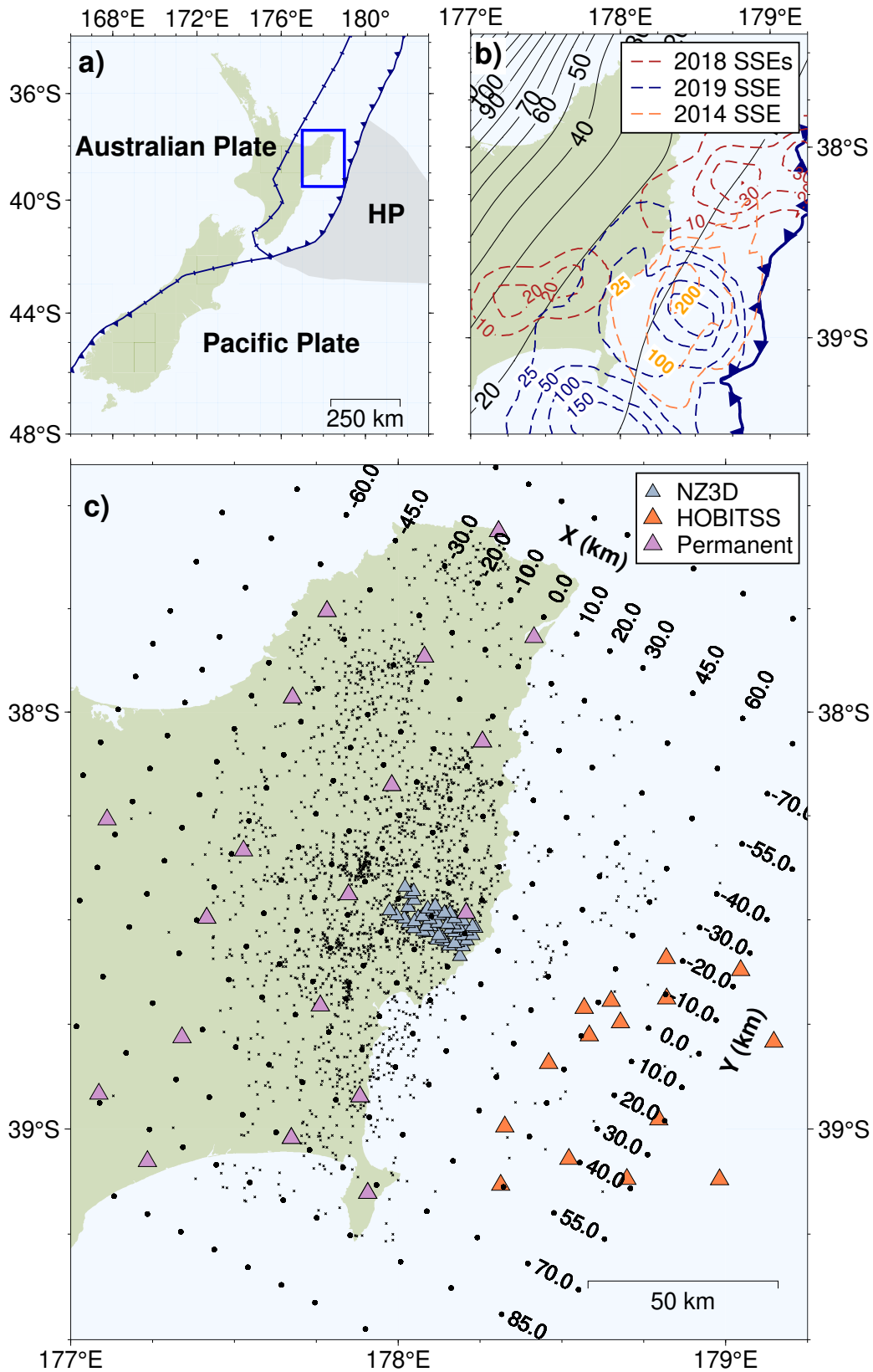


Figure 5.1: a) Tectonic map showing study region (blue rectangle) and major plate boundaries (Bird, 2003). b) Map of study region. Contours: previous SSEs (Wallace et al., 2016; Woods et al., 2024). c) Earthquakes (crosses) and seismometers (triangles) used in the tomographic inversion routine. Stations from the temporary Gisborne array (blue), HOBITSS network (orange), and permanent short-period and broadband instruments (pink) are used. Black circles: nodes used in finer 3-D inversion, with X and Y co-ordinates labelled.

southern Hikurangi margin, the northern margin is characterised by particularly shallow SSEs and relatively low interseismic coupling ([Wallace, 2020](#)). The Hikurangi Plateau, the large igneous province that forms the portion of the Pacific Plate subducting beneath North Island, is ~35 km thick, with an eclogitic lower crust ($V_P > 8.5 \text{ km s}^{-1}$ [Reyners et al., 2011](#); [Eberhart-Phillips and Bannister, 2015](#)). Subduction of the Hikurangi Plateau, to depths of up to 50 km, has resulted in a subaerial forearc ([Reyners et al., 2011](#)). At the coastline, where the dense, onshore network is located, the subducting slab is ~15 km depth. The Hikurangi Plateau is characterised by protruding seamounts, with dimensions of up to tens of kilometres laterally and a few kilometres vertically (Figure 5.2, e.g., [Barker et al., 2018](#)). Drilling of offshore, unsubducted seamounts demonstrates that these seamounts are formed of fractured and hydrated volcanoclastic sediments and basalt, with relatively low $V_P \approx 4 \text{ km/s}^{-1}$ ([Gase et al., 2023](#)). Numerical modelling and offshore seismic reflection surveys suggest that seamount subduction enhances the subduction of thick sediment packages to depth ([Bell et al., 2010](#); [Ellis et al., 2015](#)).

The offshore portion of the northern Hikurangi margin is characterised by low plate coupling and relatively thin (1 km) sediment cover. In contrast, the southern Hikurangi margin is characterised by higher interseismic coupling and up to 5 km sediment cover, with SSEs recurring every 4–5 years at 15–50 km depth ([Wallace et al., 2009](#); [Wallace, 2020](#); [Bassett et al., 2025](#)). Deeper SSEs were also recorded at the central Hikurangi margin in 2006 and 2008 ([Wallace and Eberhart-Phillips, 2013](#)). Subduction of rough seafloor relief in the north is thought to contribute to heterogeneous plate structure, upper-plate fracturing, and localised fluid pathways (e.g., [Gase et al., 2023](#); [Bangs et al., 2023](#)), though these interpretations are confined to the offshore region. The northern Hikurangi margin’s unusually shallow SSEs occur at low effective stress and lithostatic pressure, and fluids are thought to play a key role in enabling this behaviour (e.g., [Warren-Smith et al., 2019](#)), although the depth extent of those fluids, their sources, migration rates, and volumes remain uncertain.

Incoming plate seamounts have been identified from seismic and magnetic data (e.g., [Bell et al., 2010](#); [Barker et al., 2018](#); [Gase et al., 2024](#)), often colocated with observations of repeating earthquake clusters, tremor, and tsunami earthquakes (e.g., [Bell et al., 2014](#); [Todd et al., 2018](#);

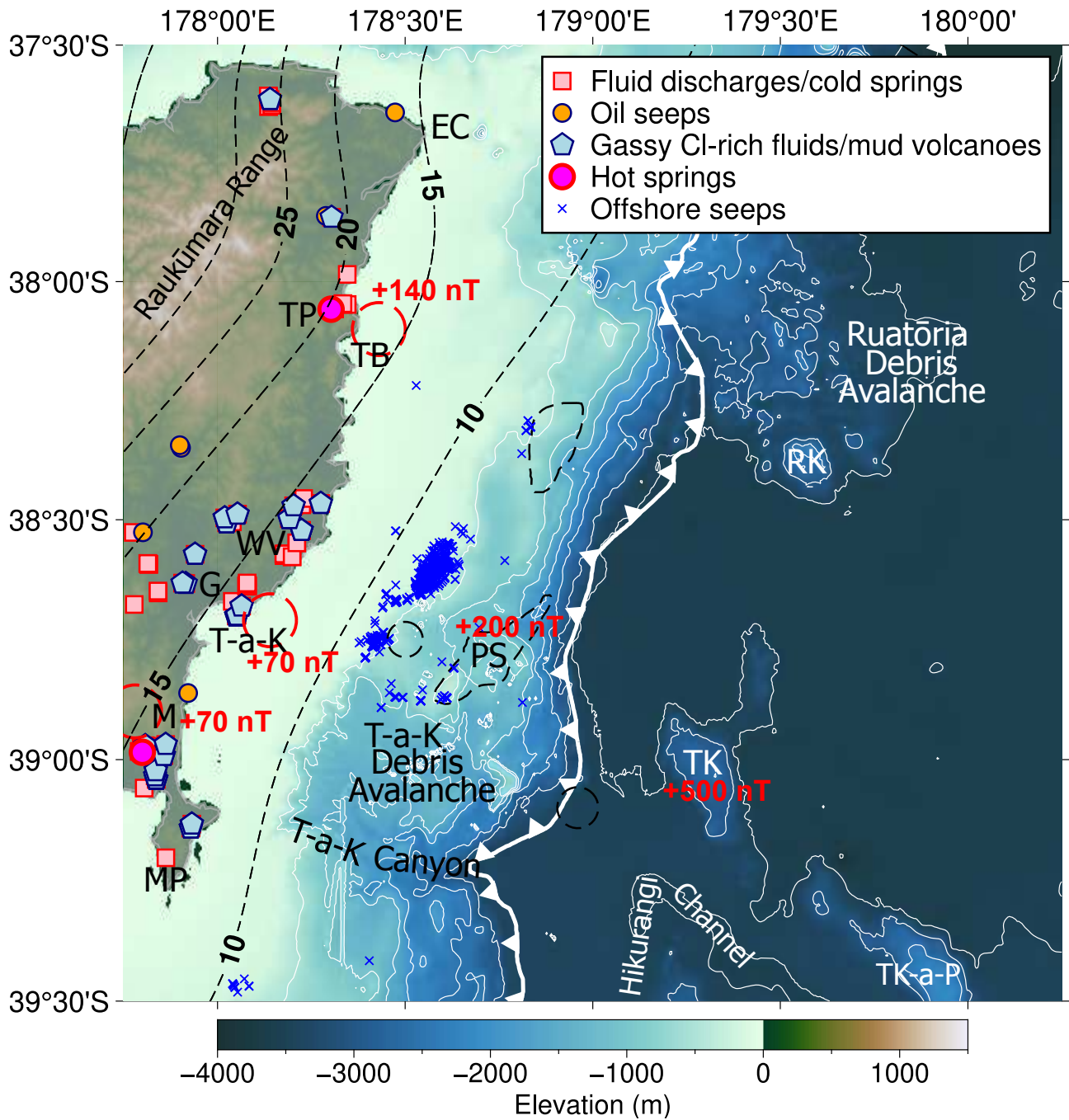


Figure 5.2: Relief map showing key features discussed in this study. Pink circles: hot springs. Blue pentagons: mud volcanoes. Black contours: plate interface depth (km, [Williams et al., 2013](#)). Magnetic anomalies that have been suggested to be seamounts in previous studies are shown; these have anomalies >60 nT, greater than the background magnetic signal. Dashed red circles: magnetic anomalies ([Hunt and Glover, 1995](#)). Dashed black shapes: subducted seamounts and their magnetic anomalies where available ([Bell et al., 2010](#); [Barker et al., 2018](#); [Bangs et al., 2023](#)). Abbreviations: EC, East Cape; G, Gisborne; Mahia Peninsula; M, Mōrere hot spring; MP, Mahia Peninsula; T-a-K, Tūranganui-a-Kiwa/Poverty Bay; PS, Pāpaku Seamount; RK, Ruatōria Knoll; TB, Tokomaru Bay; TK, Tūranganui Knoll; TK-a-P, Te Kuri-a-Paoa Seamount; TP, Te Puia hot spring; WV, Waimata Valley mud volcanoes. Topography and bathymetry data are from [Tozer et al. \(2019\)](#). Offshore seep data from [Watson et al. \(2020\)](#); onshore seep data from [Reyes et al. \(2022\)](#).

Shaddox and Schwartz, 2019; Hughes et al., 2021). Constraining the location of seamounts is important for understanding seismic hazard: the two 1947 tsunami earthquakes resulted in a ~ 10 m tsunami and likely nucleated at the Pāpaku seamount (*Bell et al., 2014*). The Pacific Plate has a relatively low, uniform background magnetic structure with the exception of large positive anomalies correlated with seamounts (e.g., *Bell et al., 2010; Barker et al., 2018*). For example, the offshore, unsubducted Tūranganui Knoll has a magnetic anomaly of +500 nT (labelled TK, Figure 5.2), and the subducted Pāpaku seamount (labelled PS, Figure 5.2) has a magnetic anomaly of +200 nT (*Barker et al., 2018*).

Seamount subduction has also been invoked to explain major forearc indentations, such as the Ruatoria and Tūranganui-a-Kiwa/Poverty Indentations, and debris avalanche processes (Figure 5.2 *Collot et al., 2001; Lewis et al., 2004; Pedley et al., 2010*). Previous studies have suggested that seamount subduction results in tunnelling and collapsing processes trailing updip of the seamount, resulting in sediment lenses and steep-sided indentations (*Collot et al., 1996, 2001; Lewis et al., 2004; Pedley et al., 2010; Bangs et al., 2023*). Experimental modelling suggests that the indentations, such as the 50-km wide Ruatoria Indentation and the 40-km wide Tūranganui-a-Kiwa/Poverty Indentation seen in bathymetry data, are similar in groove width to the diameter of seamounts (*Collot et al., 2001; Pedley et al., 2010*), providing constraints on expected seamount width downdip of the indentations, though no seamounts associated with these avalanches have yet been imaged.

Geologically, the Raukūmara peninsula can be divided into three main units from west to east: (1) the Early Cretaceous greywackes of the Raukūmara Range; (2) the East Coast Allochthon, thrust over the greywackes; and (3) Neogene marine sedimentary rocks overlying the allochthon to the east, which show margin-parallel normal faulting together with diapirism and mud volcanism (*Mazengarb and Speden, 2000*). The East Coast Allochthon (ECA) and Raukūmara Peninsula are shown in Figure 5.3. At the Raukūmara Peninsula, the upper plate is in a state of extension (*Mazengarb and Speden, 2000; Wallace et al., 2009; Henrys et al., 2025*), which may further influence fracture permeability and fluid migration through the forearc. The East Coast of the Raukūmara Peninsula and eastern offshore region hosts numerous hot springs, cold seeps and mud volcanoes (Figure 5.2), the likely surface expression of fluid

migration (e.g., [Giggenbach et al., 1993](#); [Reyes et al., 2010](#); [Watson et al., 2020](#); [Reyes et al., 2022](#)). This region is therefore a suitable target for a high-resolution V_P and V_P/V_S model that can constrain potential anomalies associated with these features.

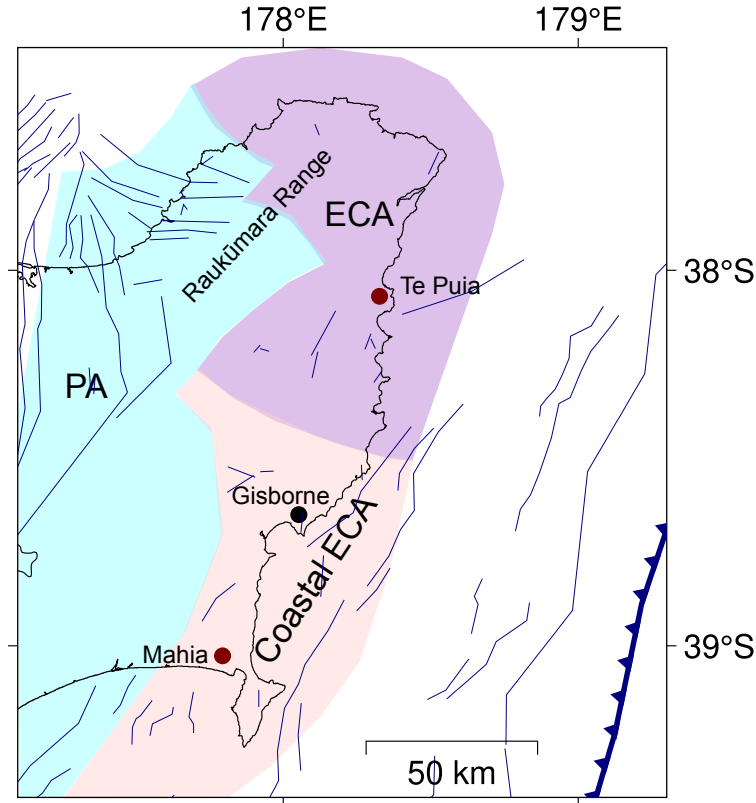


Figure 5.3: Simplified geologic map of the Raukūmara Peninsula based on [Mazengarb and Speden \(2000\)](#); [Eberhart-Phillips and Bannister \(2015\)](#). Abbreviations: PA, Pahau terrane (early Cretaceous); ECA, East Coast Allochthon. Brown circles: hot springs ([Reyes et al., 2022](#)). Navy lines: faults and plate boundaries ([Bird, 2003](#); [Seebeck et al., 2022](#)).

5.2 Local earthquake tomography

5.2.1 Fundamentals of tomographic inversion

Earthquake travel-times are calculated by assuming the ray path through a velocity model (the forward model), from which we can make interpretations about structure, geometry, and physical properties of the subsurface. In local earthquake tomography (LET), we measure P- and S-wave travel-times and invert for a velocity model. We use the SimulPS 2023 programme ([Thurber and Eberhart-Phillips, 1999](#); [Eberhart-Phillips et al., 2024](#)) to invert for V_P and V_P/V_S models. SimulPS makes use of a damped least squares, full matrix inversion using local

earthquake sources. An additional advantage of SimulPS is its ability to invert S-P travel times directly for a V_P/V_S model, rather than separately inverting for a V_P and V_S from which a V_P/V_S model is derived despite variation in resolution between the models. Inverting for V_P and V_S separately with P- and S-phase travel-times respectively, then calculating V_P/V_S at each node tends to introduce artifacts into the model.

Earthquake arrival times, t_{ij} , are defined as:

$$t_{ij} = \tau_{ij} + T_{ij}, \quad (5.1)$$

where τ_{ij} is the earthquake origin time and T_{ij} is the travel-time.

During the inversion procedure, V_P , V_P/V_S and earthquake locations are obtained by inverting P-wave travel-times and S-P differential travel-times:

$$T_{ij} = \int_{source}^{receiver} u, ds \quad (5.2)$$

where u is slowness (inverse of velocity) and ds is the ray path segment. Travel-times are a function of earthquake-receiver locations and the velocity model through which the ray path propagates. We therefore iteratively invert for earthquake hypocentres and velocity models. Predicted travel-times are calculated by assuming an initial velocity model and earthquake hypocentre, from which the residual travel-time can be calculated as:

$$r_{ij} = t_{ij}^{obs} - t_{ij}^{calc} \quad (5.3)$$

For a given earthquake-receiver pair, this can be linearly approximated using a Taylor series expansion to relate the hypocentre and velocity parameters:

$$r_{ij}^P = \frac{\partial T_{ij}}{\partial x} \Delta x + \frac{\partial T_{ij}}{\partial y} \Delta y + \frac{\partial T_{ij}}{\partial z} \Delta z + \Delta \tau_i + \int_{ij} \delta u^P(x, y, z) ds \quad (5.4)$$

where r_{ij}^P is the P-wave travel-time residual, and

$$\int_{ij} \delta u^P(x, y, z) ds = \sum_{n=1}^N \frac{\partial T_{ij}}{\partial m_n} \Delta m_n. \quad (5.5)$$

For differential travel-times,

$$r_{ij}^{s-p} = \int_{ij} \frac{\delta(\frac{V_P}{V_S})}{V_P} ds \quad (5.6)$$

where r_{ij}^{S-P} is the differential travel-time residual. x, y, z denote the earthquake coordinates and depth. The x, y, z partial derivatives are a function of velocity. For example, for the hypocentre x -coordinate:

$$\frac{\partial T_{ij}}{\partial x} = -\frac{1}{v} \left(\frac{dx}{ds} \right)_{source} \quad (5.7)$$

During LET, the velocity model is iteratively perturbed with the aim of minimising the misfit, r_{ij} . Each time the velocity model is perturbed, ray tracing is recalculated. Ray tracing is highly non-linear, due to the number of possible take-off angles and paths taken to the receiver. Ray bending approaches involve iterative adjustments of the ray geometry along the ray path, until Fermat's principle is satisfied. Simul2000 ([Thurber, 1983](#)) and its derivatives (e.g., SimulPS 2017, SimulPS 2023 [Eberhart-Phillips and Fry, 2017](#); [Eberhart-Phillips et al., 2024](#)) use pseudo-bending, which is based on Cartesian coordinate systems, thus is well-suited to small-scale local earthquake tomography problems. The ray equation does not need to be solved; rather, a three-point perturbation scheme is used where the travel-times are calculated along the ray path. Perturbations to the earliest-arriving ray path are performed, allowing for the first-arriving phase to be derived from a ray path that is not necessarily the shortest distance. Longer distance ray paths may arrive sooner due to propagation through faster velocity regions. Ray path segment perturbation can also account for station elevation, particularly useful for the Hikurangi margin subaerial forearc.

Pseudo-bending of the ray removes the need for directly solving the ray equation. The ray equation is then written as:

$$-\frac{\delta^2 r}{ds^2} = \frac{\nabla V - \frac{dV}{ds} \cdot \frac{d\mathbf{r}}{ds}}{V}, \quad (5.8)$$

where $\mathbf{r}$ is the position vector along the ray, s is the ray length, and V is the velocity.

The non-linear relationship between hypocentres and velocity model means that hypocentres must be perturbed as well as the velocity model to prevent production of artifacts into the velocity model and biasing towards the starting model used to initially locate the earthquakes. Hypocentre and velocity parameters must both be solved for, therefore:

$$\mathbf{d} = \mathbf{H}\Delta\mathbf{h} + \mathbf{V}\Delta\mathbf{v}, \quad (5.9)$$

where $\mathbf{d}$ represents the travel-time residuals vector, $\mathbf{H}$ and $\Delta\mathbf{h}$ represent the hypocentre partial derivatives and perturbations, and $\mathbf{V}$ and $\Delta\mathbf{v}$ represent the velocity model partial derivatives and perturbations. This equation is solved using a damped least-squares matrix inversion.

$$\mathbf{m} = (\mathbf{G}^T \mathbf{G} + \epsilon \mathbf{I})^{-1} \mathbf{G}^T \mathbf{d}, \quad (5.10)$$

where $\mathbf{m}$ is the model vector (hypocentre and velocity perturbations), $\mathbf{G}$ is the matrix of the travel-time partial derivatives, $\mathbf{d}$ is the data vector (travel-time residuals), ϵ is the damping parameter, discussed in section 5.3.2, and $\mathbf{I}$ is the identity matrix.

Figure 5.4 shows the effects of damping on the final velocity model.

5.3 Methods

The tomographic inversion scheme is summarised in the flowchart (Figure 5.5).

5.3.1 Data

To include ray coverage on- and offshore, we combined the earthquake catalogues of [Yarce et al. \(2019\)](#) and Chapter 3, derived from dense temporary offshore and onshore seismograph deployments respectively, supplemented in both cases by picks from the permanent seismo-

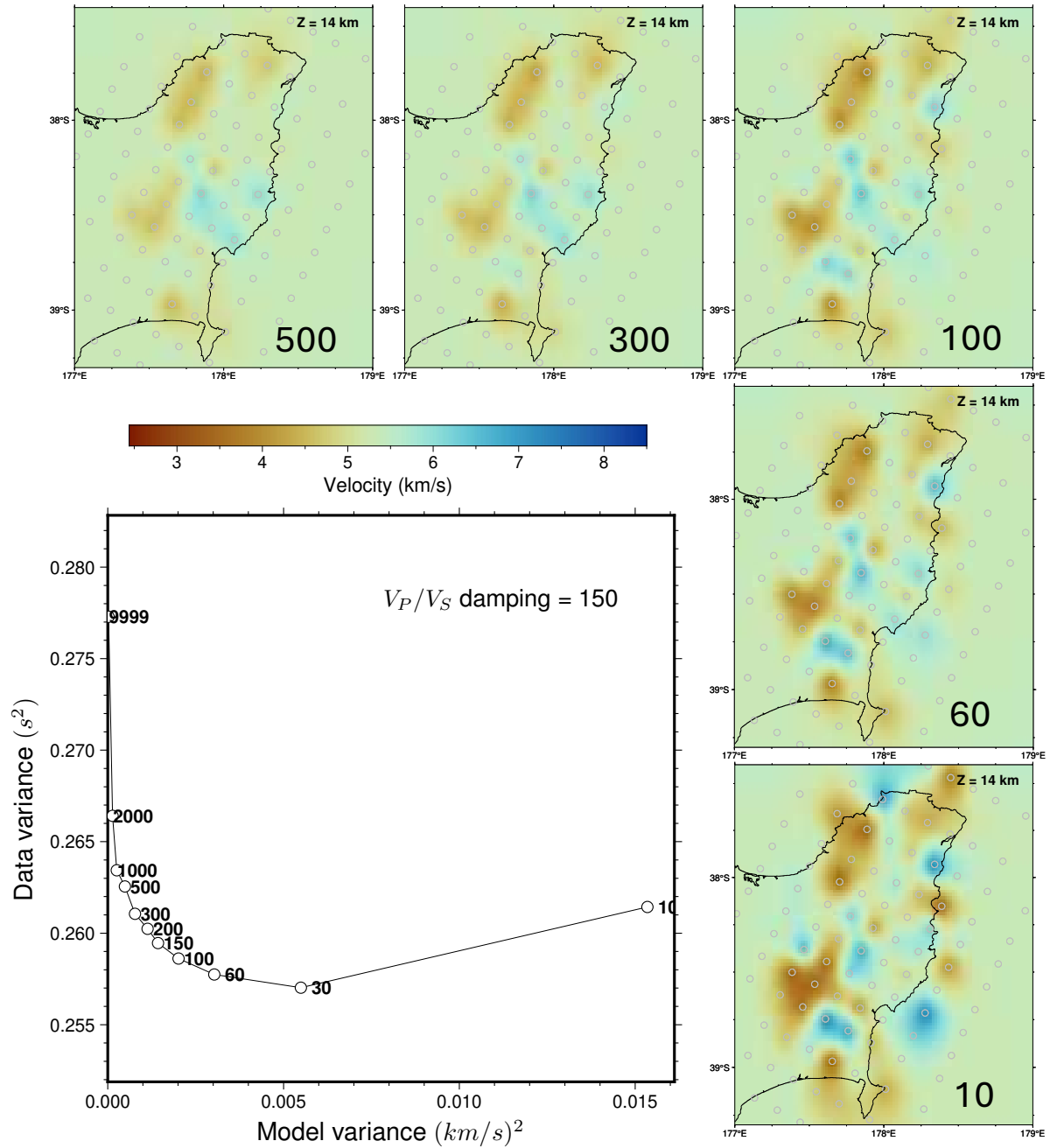


Figure 5.4: Trade-off curve for V_P damping parameters, with V_P/V_S damping fixed at 150. Maps show V_P models at 14 km depth for different damping parameters.

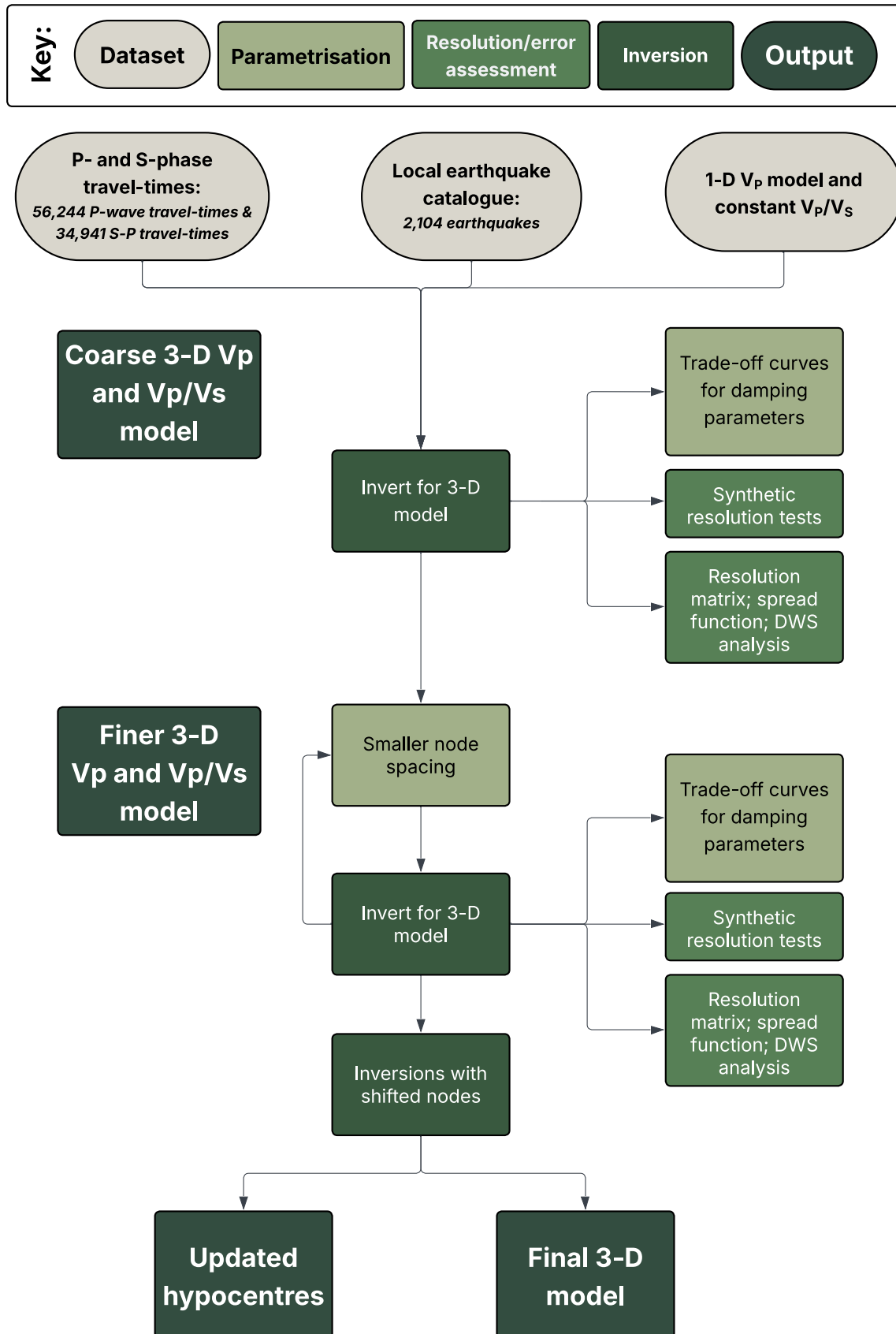


Figure 5.5: Diagrammatic summary of tomography workflow.

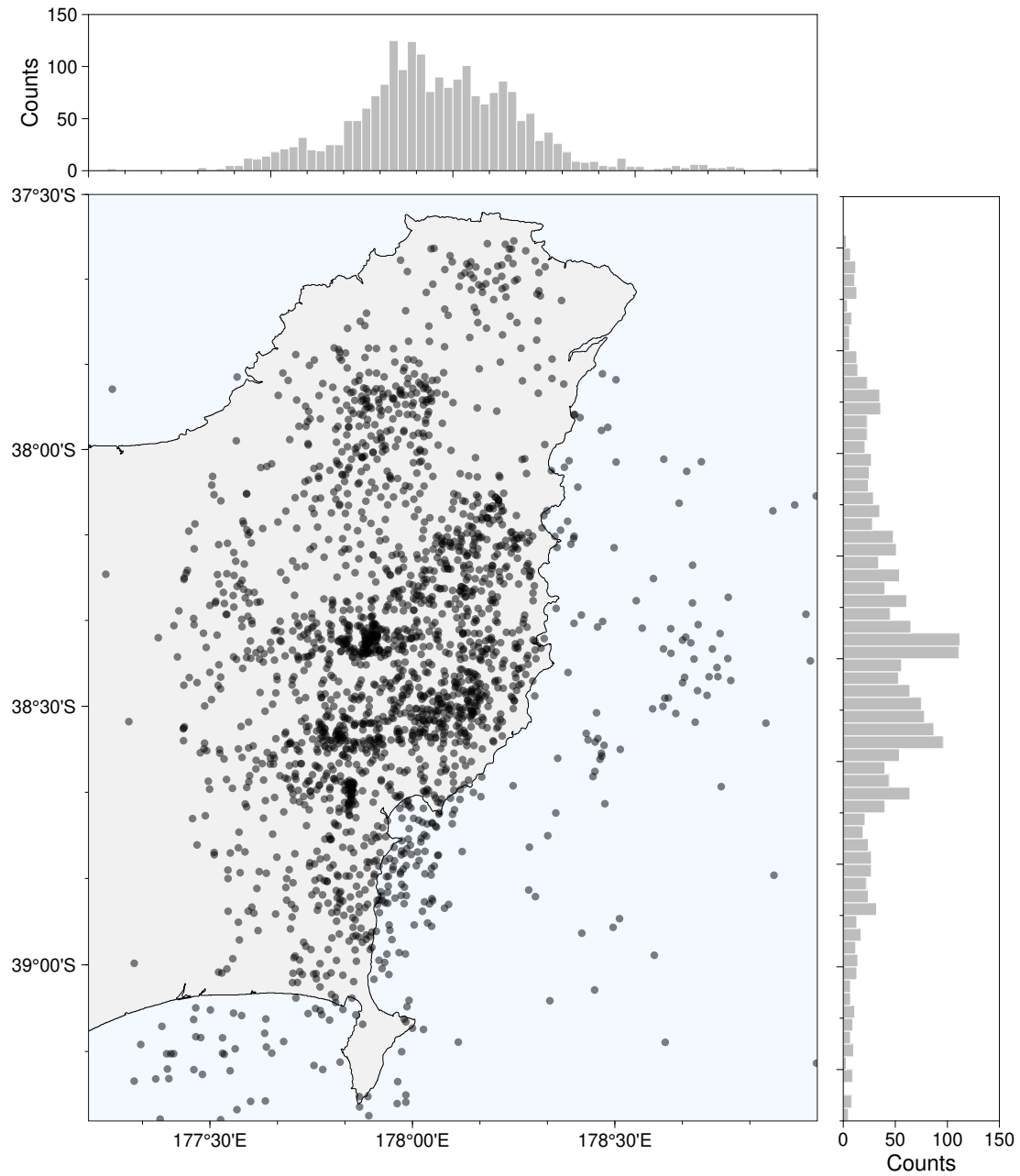


Figure 5.6: Map view of earthquakes used in SimulPS inversion. Histograms show longitude and latitude distributions of earthquakes.

graph network across Aotearoa New Zealand (Figure 5.1; *GNS Science* (2019, 2021)). The offshore catalogue was derived from the HOBITSS experiment (*Yarce et al.*, 2019), whereas the onshore catalogue was derived from a temporary land-based deployment near Gisborne deployed from December 2017 to October 2018 (NZ3D: Figure 5.1c). Combining these catalogues improved ray coverage (compared to using only the onshore catalogue) across the onshore-offshore transition and within the deeper part of the slow slip region. Phase picks and initial hypocentres were taken from the published catalogues and then filtered using the criteria below.

To select earthquakes suitable for tomography, we required each to have at least 12 P-wave travel-time measurements, at least one S-P travel-time measurement, an azimuthal gap $<210^\circ$, and a root mean square (RMS) travel-time misfit of <3 s. These criteria were chosen as, except for the RMS, they are independent of the initial velocity model used to locate the earthquakes. Earthquakes with RMS misfits >3 s are likely attributed to clock timing errors rather than Earth structure (though we observed no clock timing errors) or false positive picks or phase associations. The low S-P threshold was chosen to ensure that as many earthquakes from the HOBITSS network could be included; for earthquakes recorded by the onshore network, the mean number of S-P measurements per earthquake was 15.9.

These criteria yielded a final catalogue of 2,104 earthquakes, comprising 56,244 P-wave and 34,914 S-P travel-time measurements. Of these, 1,831 were recorded by the onshore network and 273 were recorded by the offshore network. The mean RMS travel-time misfit of the selected earthquakes was 0.46 s. The selected earthquakes span depths up to 140 km, with a mean depth of 28 km, and magnitudes of M_L 0.89–4.73. The retained earthquakes sample both the offshore shallow SSE region and the deeper onshore continuation of the subduction system.

5.3.2 Tomographic inversion scheme

To reduce the dependence of the final model on the starting model and initial node positions, we used a staggered inversion scheme (e.g., *Evans et al.*, 1994; *Haberland et al.*, 2009): we first derived a 1-D velocity model using VELEST (*Kissling et al.*, 1994), then inverted for coarse

then finer 3-D models using the SimulPS code (2023 release; [Thurber, 1993](#); [Eberhart-Phillips et al., 2024](#)). A gradational inversion minimises the effects of the 1-D model in poorly sampled regions, which can introduce artefacts into the model ([Evans et al., 1994](#)). SimulPS is a full matrix inversion method for the iterative joint inversion of 3-D V_P and V_P/V_S structure, and earthquake relocations. P-wave and S-P travel-times are inverted to update velocity models and hypocentres. We used the 1-D V_P velocity model derived from the same temporary catalogue in Chapter 3 as the V_P starting model.

Wadati analysis

We calculated a constant V_P/V_S value from Wadati-diagram analysis (Figure 5.7) as the starting V_P/V_S model value. The catalogue included some station-event pairs with anomalously high V_P/V_S values, clustered in the northern periphery of the inversion region (Figure 5.8). We calculated V_P/V_S for individual station-earthquake pairs and removed any measurements with values outside the range 1.5–2.0. A linear least-squares regression, with the best-fit line fixed through the origin, gave a starting V_P/V_S of 1.77 ± 0.04 (99 per cent confidence limit; Figure 5.7).

Note that the V_P/V_S derived in this chapter differs from the V_P/V_S derived in Chapter 3. In Chapter 3, the V_P/V_S of 1.73 was calculated using the whole catalogue of 3,071 earthquakes. In this chapter, we only include earthquakes that are located inside the network and pass the quality control criteria required for LET, which differs to the criteria which we required in Chapter 3. The lower V_P/V_S found in Chapter 3 is likely due to the inclusion of earthquakes located at distances >100 km landward of the trench, where upper plate fluid content and fault interconnectivity decreases, resulting in lower V_P/V_S ([Eberhart-Phillips et al., 2017](#)). In this Chapter, the focus on earthquakes located primarily along the fluid-rich east coast likely accounts for the slightly higher V_P/V_S value.

Figure 5.7 shows the Wadati diagram for the Gisborne catalogue and the combined Gisborne and HOBITSS catalogues. The inclusion of the HOBITSS data increases V_P/V_S slightly, likely due to the offshore location of the HOBITSS dataset, which therefore samples a more fluid-rich region. Figure 5.9 shows histograms of the travel-times for each catalogue, showing a slightly

slower distribution of S-P travel-times for the HOBITSS network. Nevertheless, the inclusion of the HOBITSS data results in a V_P/V_S consistent with the onshore catalogue alone, and the difference is below the resolution limit of the V_P/V_S model.

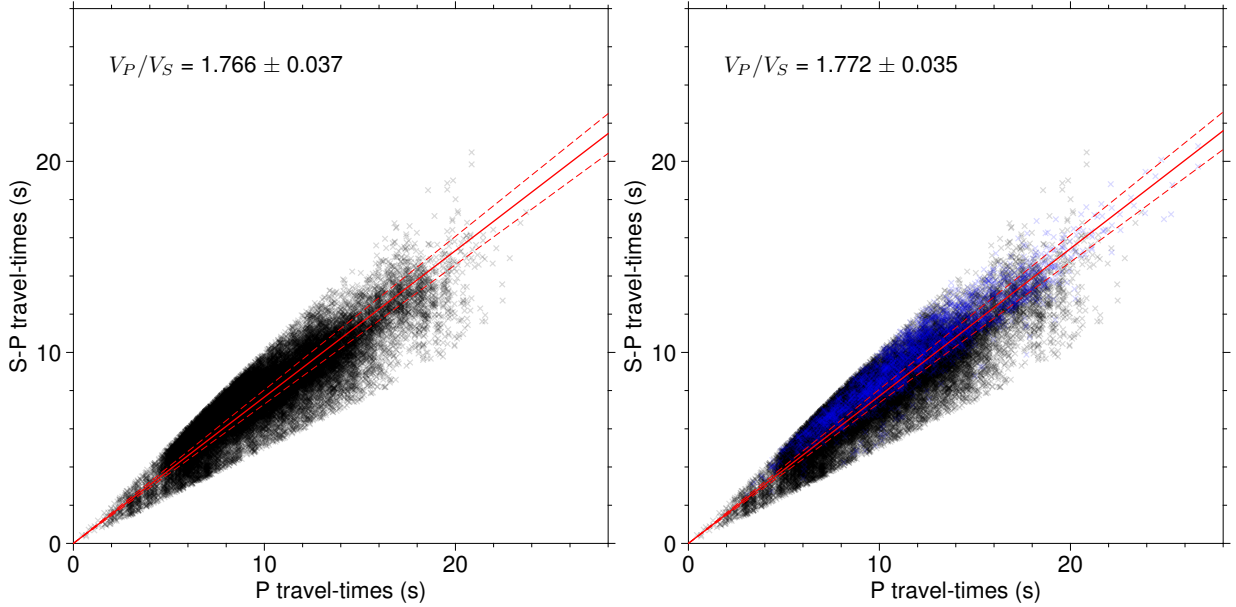


Figure 5.7: Wadati plot for travel-times used in LET for Gisborne array catalogue (left) and with the HOBITSS catalogue (right). Red line: linear least square regression solution with 99% confidence interval (dashed lines).

Model parametrisation

We defined a 3-D velocity model volume centred on 178°E, 38.45°S (Figure 5.1). The y -axis was aligned parallel to the northern Hikurangi margin, 155° anticlockwise from north, positive to the south-west. The x -axis was aligned margin-perpendicular, positive to the south-east, and the vertical axis was positive downwards.

The coarse 3-D model consisted of 12×14 x and y nodes, with a minimum 15 km node spacing and with vertical nodes at 1, 5, 9, 14, 20, 28, 36, 52 and 80 km depth. Additional planes of fixed nodes were placed at the outermost bounds of the model in all directions. The finer 3-D model consisted of 15×17 x and y nodes, with a minimum 10 km node spacing and with vertical nodes at 1, 4, 8, 12, 16, 20, 24, 28, 32, 36, 40, 50 and 80 km depth. Node spacing was decreased in the well-constrained depth regions, based on spread function analysis from the coarse 3-D model, and increased in lower-resolution parts of the model. We used variable node spacing to allow for the variation in ray coverage across the model volume. This node spacing

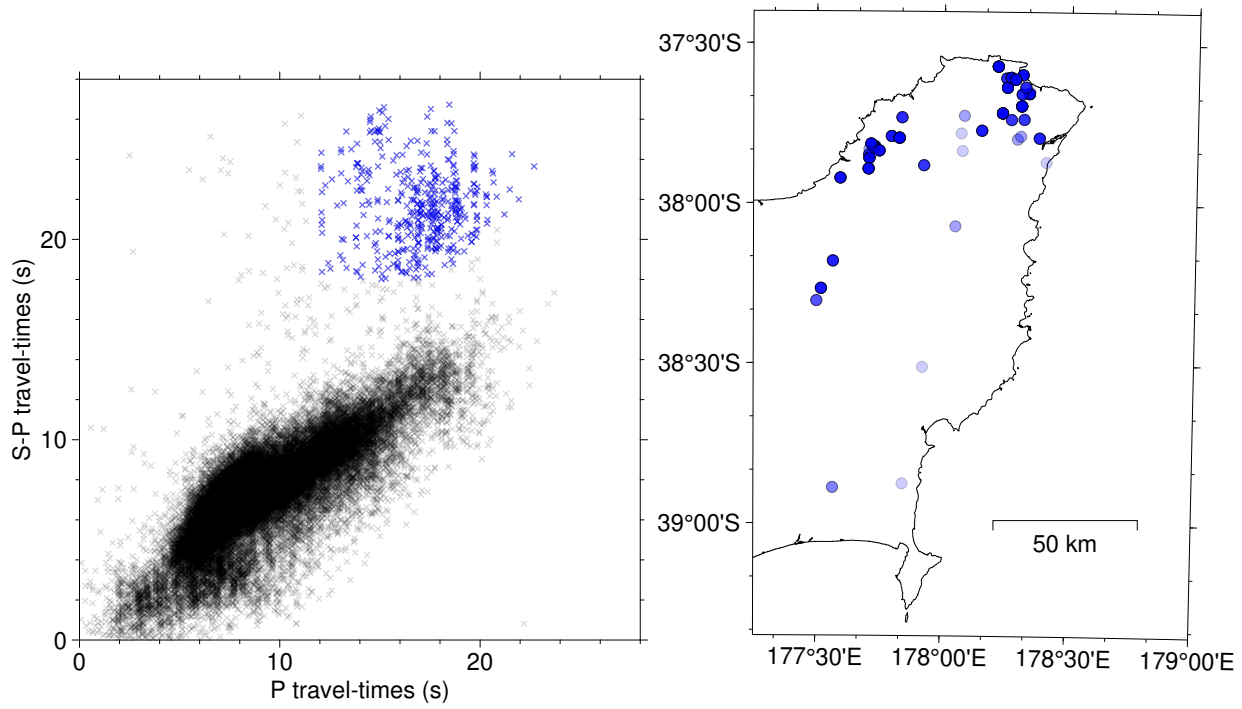


Figure 5.8: Left) Wadati diagram of all V_P/V_S measurements, with outlier V_P/V_S measurements in blue. Right) locations of corresponding earthquakes for outlier measurements with 80% transparency i.e., darker blue corresponds to greater number of outlier measurements per earthquake.

remains consistent with the nominal Fresnel zone limit of ~ 5 km, assuming a characteristic source frequency of 10 Hz and a depth limit of 40 km and assuming the cross-ray diameter of the Fresnel zone is on the order of $(\lambda L)^{1/2}$ for a wave of wavelength and ray length L (Montelli *et al.*, 2004).

We allowed up to 18 iterations, each involving earthquake relocation and velocity inversion. We selected the output from iteration 7 as the final model, because beyond this point, the reduction in data variance was small, the V_P and V_P/V_S models had converged, and the results were no longer sensitive to the maximum permissible model perturbation per iteration (Figure 5.10).

Damping parameters

Damping parameters are used to stabilise the inversion scheme, limiting how much the velocity model can be perturbed before solving for travel-times. Damping coefficients were defined for V_P and V_P/V_S to stabilise the 3-D model inversions. Damping parameters for the coarse and finer 3-D models were chosen by running seven-iteration inversions over a broad range of

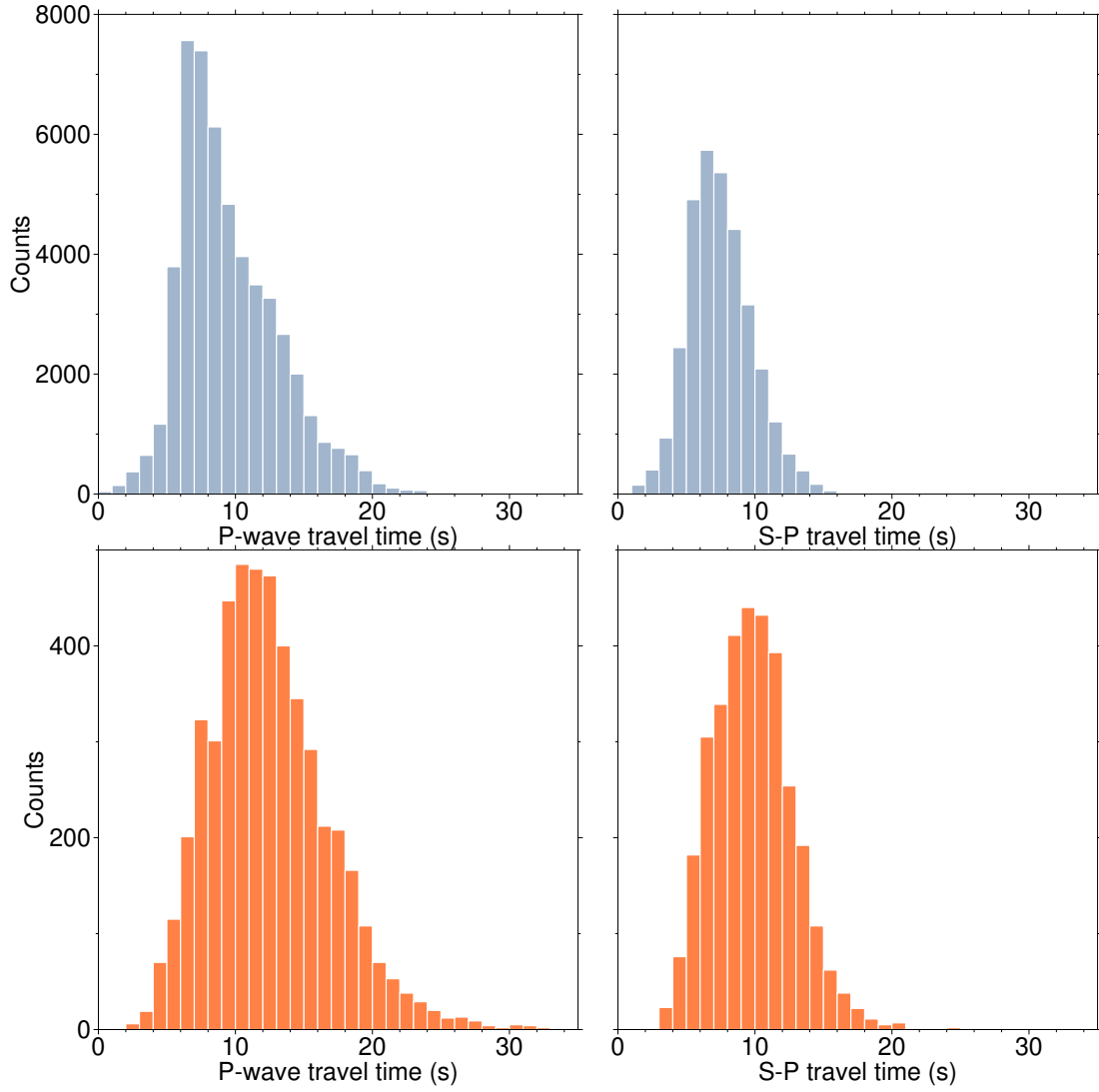


Figure 5.9: Top panels) P- and S-P travel-times for earthquakes used from the NZ3D catalogue. Lower panels) P- and S-P travel-times for earthquakes used from the HOBITS catalogue.

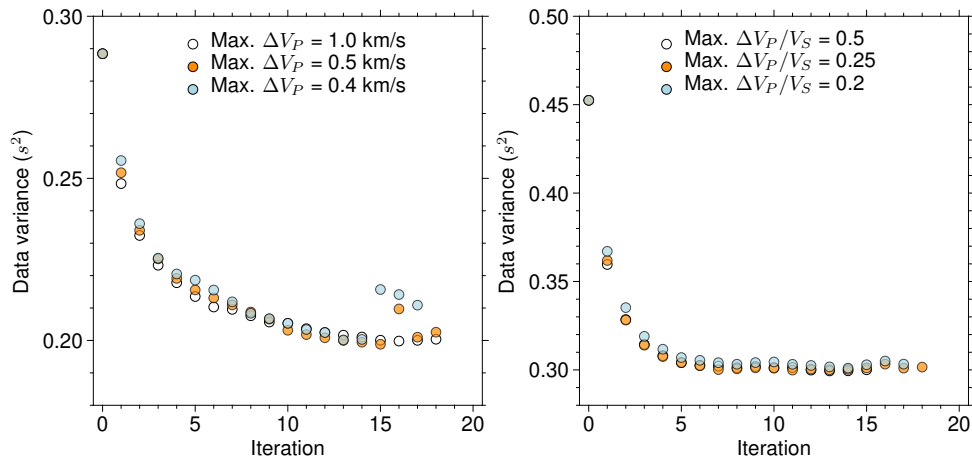


Figure 5.10: Iteration trade-off curve for V_P (left) and V_P/V_S (right). Colours correspond to maximum permissible change in V_P and V_P/V_S per iteration.

V_P and V_P/V_S damping values, and then examining trade-off curves of data variance versus model variance. Following [Eberhart-Phillips \(1986\)](#), we consider the inversion to be unstable for damping values at which the trade-off curve begins to deviate from the typical hyperbolic curve.

Following the parametrisation method of [Lin et al. \(2014\)](#), we chose the V_P damping parameter by holding the V_P/V_S damping parameter fixed at a large value (1000) to minimise the effect of the S-P travel-time data on the trade-off curve. From this, we selected a V_P damping value of 200 (Figure 5.11). We then ran the inversions for a large range of V_P/V_S values, while holding the V_P damping at a value of 200. From this we selected a value of 500. Finally, we reran the V_P inversions to plot a trade-off curve for a fixed V_P/V_S value of 500 to ensure that V_P damping of 200 was still at the ‘knee’ of the curve. The same process was repeated when we progressed to the finer 3-D model spacing, resulting in retention of values of 200 and 500 for V_P and V_P/V_S damping respectively. We did not invert for station corrections to avoid projecting resolvable velocity structure into station corrections (e.g., [Lin et al., 2014](#)).

For the same damping parameters, the inclusion of data from the HOBITSS reduces the data variance (Figure 5.12), demonstrating the improvement in residual reduction from the inclusion of additional data.

Interpolation methods

An easily overlooked stage of the LET workflow is the interpolation between nodes in order to visualise the model. SimulPS outputs a velocity at each node, spaced in this case a minimum of every 10 km laterally or 3 km vertically. Figure 5.13 shows the distribution of V_P values in a plotted model for different interpolation methods. SimulPS performs ray tracing in terms of slowness, in order to preserve travel-times, in ray segments defined by the user. Therefore, to replicate the interpolation method in SimulPS, we convert velocities to slowness and perform a trilinear interpolation, such that the velocity at any position in the model is a function of the velocity of, and distance from, the surrounding eight nodes.

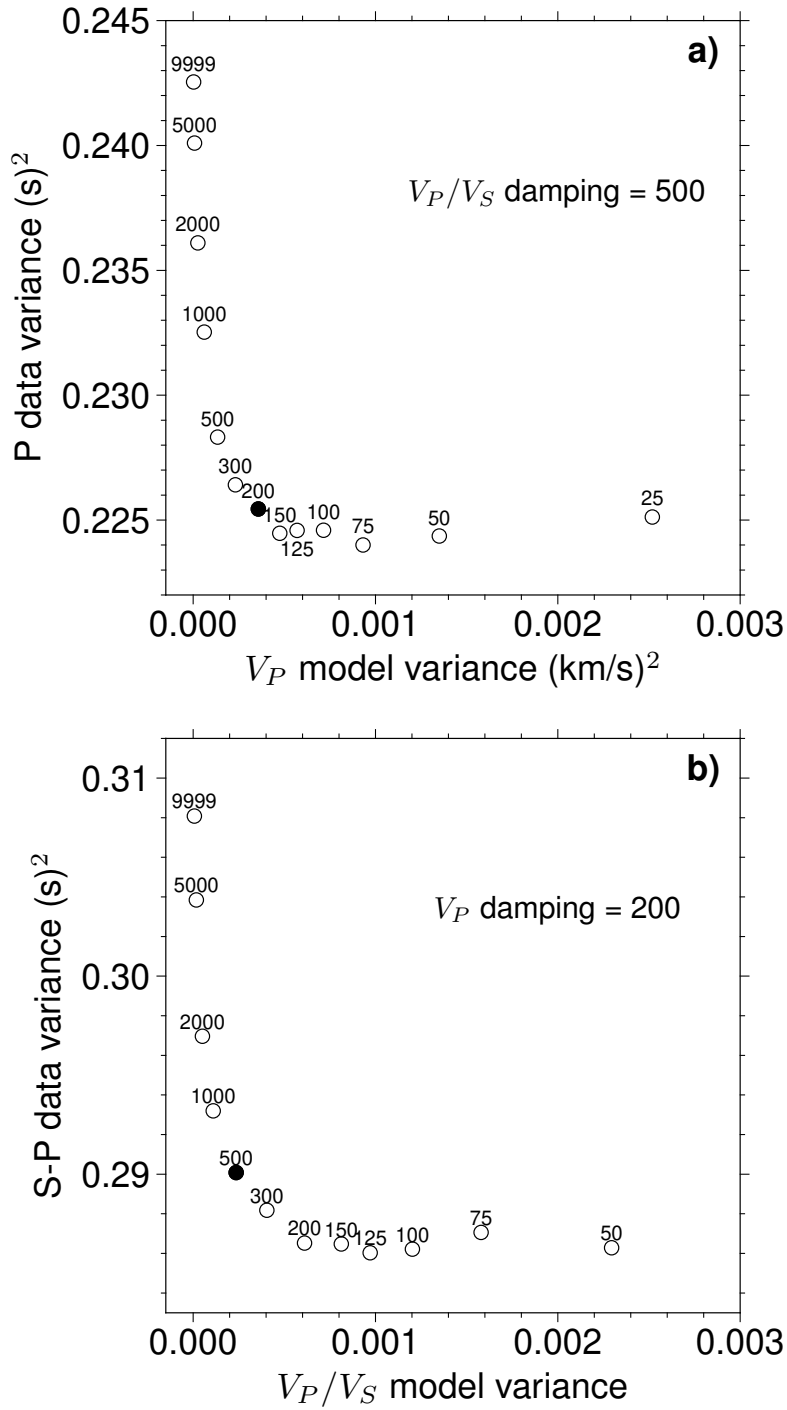


Figure 5.11: Trade-off curves of data variance versus model variance for final 3-D model (seven iteration inversions with finest node spacing) **a)** V_P model with V_P/V_S damping fixed at 500 and **b)** V_P/V_S model with V_P damping fixed at 200. Filled black circles: chosen damping parameter from each trade-off curve. Open black circles/numbers: damping parameters tested.

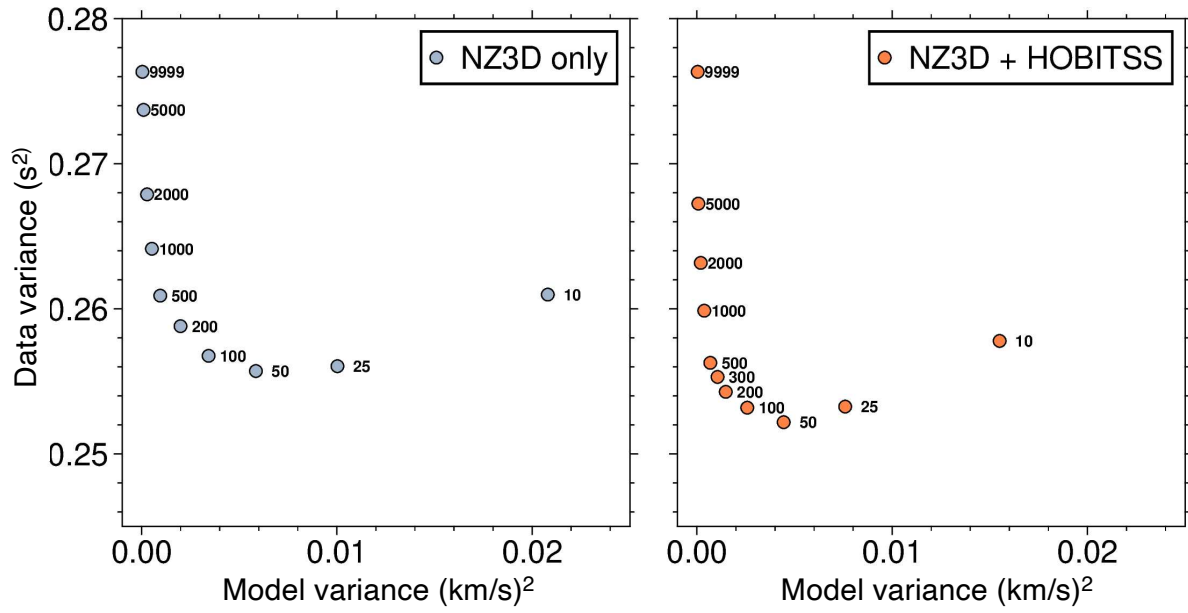


Figure 5.12: Example V_P trade-off curve with NZ3D data only (left) and with the combined catalogues of the NZ3D and HOBITSS networks (right).

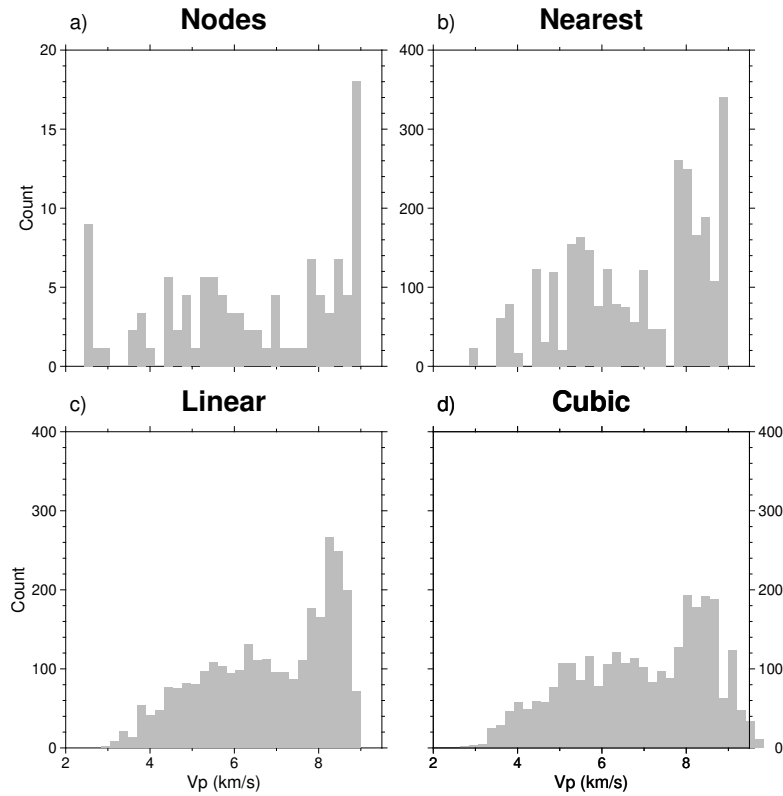


Figure 5.13: V_P distributions for cross-section view ($Y = 14 \text{ km}$) with different interpolation methods. **a)** Velocities output at nodes used in the inversion. **b)** Velocities interpolated based on the nearest node to the point in the mesh grid. **c)** Linear interpolation of velocities. **d)** Cubic interpolation of velocities.

Table 5.1: Summary of parameters used in SimulPS.

Parameter	Description	Value
neqs	Number of earthquakes	2141
nitloc	Number of attempts to adjust hypocentre/velocity model per iteration in order to reduce residuals	5
eigtol	Singular value decomposition cut-off in hypocentral adjustment; below this value, hypocentre is unchanged	0.001
rmscut	RMS residual below which hypocentral adjustments stop	0.01
zmin	Minimum hypocentre depth (set to highest station elevation)	1 km
dxmax	Maximum hypocentral adjustment allowed per iteration	1 km
nhitct	Minimum DWS for parameter to be included in the inversion	10
dvpmax	Maximum V_P adjustment per iteration	0.4 km/s
dvpvsmx	Maximum V_P/V_S adjustment per iteration	0.05
corrvpmax	Maximum V_P	9 km/s
corrvpmin	minimum V_P	1 km/s
vpdamp	V_P damping	200
vpvsdamp	V_P/V_S damping	500
stdamp	Station damping	1000
stepl	Step length along ray path for calculating partial derivatives along raypath	1 km
nitmax	Maximum number iterations	7
snrmct	Cut-off value for solution norm	0.001
tlim	Travel-time difference between successive iterations below which to terminate iterations	0.002 s
nitpb	Maximum number of iterations allowed during pseudo-bending per iteration	10

5.3.3 Summary of final parametrisation

For the coarse 3-D model, a damping value of 200 was used for V_P and 500 for V_P/V_S for the inversion of hypocentres and models with irregularly-spaced nodes (minimum 15 km horizontal node spacing). Perturbations of the velocity model and hypocentres were performed for seven iterations. Hypocentres were fixed for the first three iterations, based on the assumption that the initial hypocentres were well-constrained, then allowed to shift up to 1 km per iteration thereafter ([Hicks, 2015](#)). Raypaths were determined using the SimulPS implementation of the pseudo-bending ray tracing method of [Um and Thurber \(1987\)](#).

Travel-time residuals <0.75 s were weighted at 100 per cent in the inversion routine; the weighting of travel-time residuals from 0.75–3 s was linearly reduced from 100 to 2 per cent. P and S-P travel-time measurements were equally weighted. Pick errors were set to 0.1 s, equivalent to one quarter of the dominant period of a typical local earthquake seismogram recorded by the temporary network ([Diehl and Kissling, 2007](#)). Travel-time measurements for epicentral distances ≤ 75 km were weighted at 100 per cent; the weighting of epicentral distances from 75–200 km was linearly decreased from 100 to 50 per cent. At each iteration, the maximum allowed change in V_P was 0.4 km s^{-1} and V_P/V_S was 0.05.

We converted velocities at each node to slowness and then performed trilinear interpolation. We then resampled the interpolated model on a finer, irregularly spaced grid (minimum 10 km horizontal node spacing), following ([Evans et al., 1994](#); [Haberland et al., 2009](#)). This resampled model was used as the starting model for the finer 3-D model inversion, with recalculated damping parameters, and the same procedure as for the coarse model was repeated. Finally, we ‘smoothed’ the model by shifting the nodes by one-third of each node spacing in eight directions, following [Hicks et al. \(2014\)](#), effectively forming a cube of nodes around each original node position, and repeated the inversion on each of these shifted grids. For each grid, we trilinearly interpolated between nodes in increments of 0.5 km, then calculated the means and standard deviations at each point. While the grid shifts smooth potentially sharp velocity contrasts, they reduce the dependence of the final model on the chosen node positions.

5.4 Model resolution and error assessment

To assess the resolution and interpretability of the models, we analysed the derivative weight sum (DWS), performed checkerboard resolution tests, and evaluated the formal resolution and covariance matrices at each node.

The DWS provides a quickly computed, first-order estimate of resolution and measures the ray density near a node, weighted by the distance between the ray segment and the node. For a given node, the DWS is calculated using:

$$DWS(\alpha_n) = N \sum_i \sum_j \int_{P_{ij}} \omega_n(x) ds, \quad (5.11)$$

where i is the earthquake index, j is the station index, ω is the weighting as a function of coordinate position, P_{ij} is the earthquake-to-station ray path, N describes the volume affected by the n th velocity parameter α_n , and s is the step size length. In the inversion scheme, velocities were fixed at nodes where the DWS value fell below 10, following [Husen et al. \(2000\)](#).

5.4.1 Restoring resolution tests

Resolution tests involve setting up a synthetic velocity model, and determining which portions of the model can be recovered with the true earthquake and station locations. Commonly, checkerboard tests or velocity models that represent likely scales of subduction zone structures are used. This is useful for determining the length-scale of velocity anomalies that can be interpreted. Synthetic travel-time data are generated by assuming a synthetic velocity model e.g., a checkerboard of velocity perturbations, a single localised anomaly, or a likely feature in the model such as a subducting slab. Using the ray paths from the real earthquake-station pair dataset, the travel-times for these ray paths through the velocity model are calculated. These synthetic travel-times are then used to invert for a velocity model, to see how well the synthetic test is recovered. Checkerboard tests consist of a model with alternating fast and slow velocity anomalies, particularly useful for examining smearing and considering the relative resolution in different parts of the model.

To determine which areas of the model were well-resolved, we performed checkerboard resolution tests at 15 km and 10 km lateral length-scales. Starting from the 1-D velocity model, we constructed synthetic models with ± 10 per cent variations in V_P and V_P/V_S , varying with depth and every other horizontal node. This corresponds to horizontal anomalies of width 10 and 15 km, alternating between every other depth layer. Synthetic travel-times were calculated for the raypaths used in the inversion of the observed data, then inverted using the 1-D model as the starting model. Both checkerboard tests were performed using identical inversion parameters to those used in the final observed data inversion.

The 15 km checkerboard anomalies were well-recovered in the central onshore portion of the study region (Figure 5.14). The 10 km checkerboard anomalies were well-recovered only in a small region closest to the temporary onshore network. We therefore interpret only anomalies with length-scales of >15 km onshore, and >10 km in the immediate vicinity of the temporary network.

5.4.2 Spread function

We used the SimulPS implementation of the full model resolution matrix calculation (Thurber, 1993), from which the spread function (SF) was calculated (Toomey and Foulger, 1989). The SF compresses each row of the resolution matrix to a single value that represents the resolution at that node, while accounting for smearing between neighbouring nodes. The spread function (S_j) for a given node, j , is calculated using:

$$S_j = \log \left[|r_j|^{-1} \sum_{k=1}^N \left(\frac{r_{jk}}{|r_j|} \right)^2 D_{jk} \right], \quad (5.12)$$

where each element, r_{jk} , of a row of the matrix for node j is weighted by its distance D_{jk} from the node, and $|r_j|$ is the norm of the resolution matrix row. A node with low resolution and high dependence on other nodes has a large SF, and vice versa.

There is no universally applicable SF threshold because acceptable values depend on the data and node distribution in a given study. Our calculated SF values range between -0.24 and 4. We selected the maximum acceptable SF by examining the degree of smearing (Figure

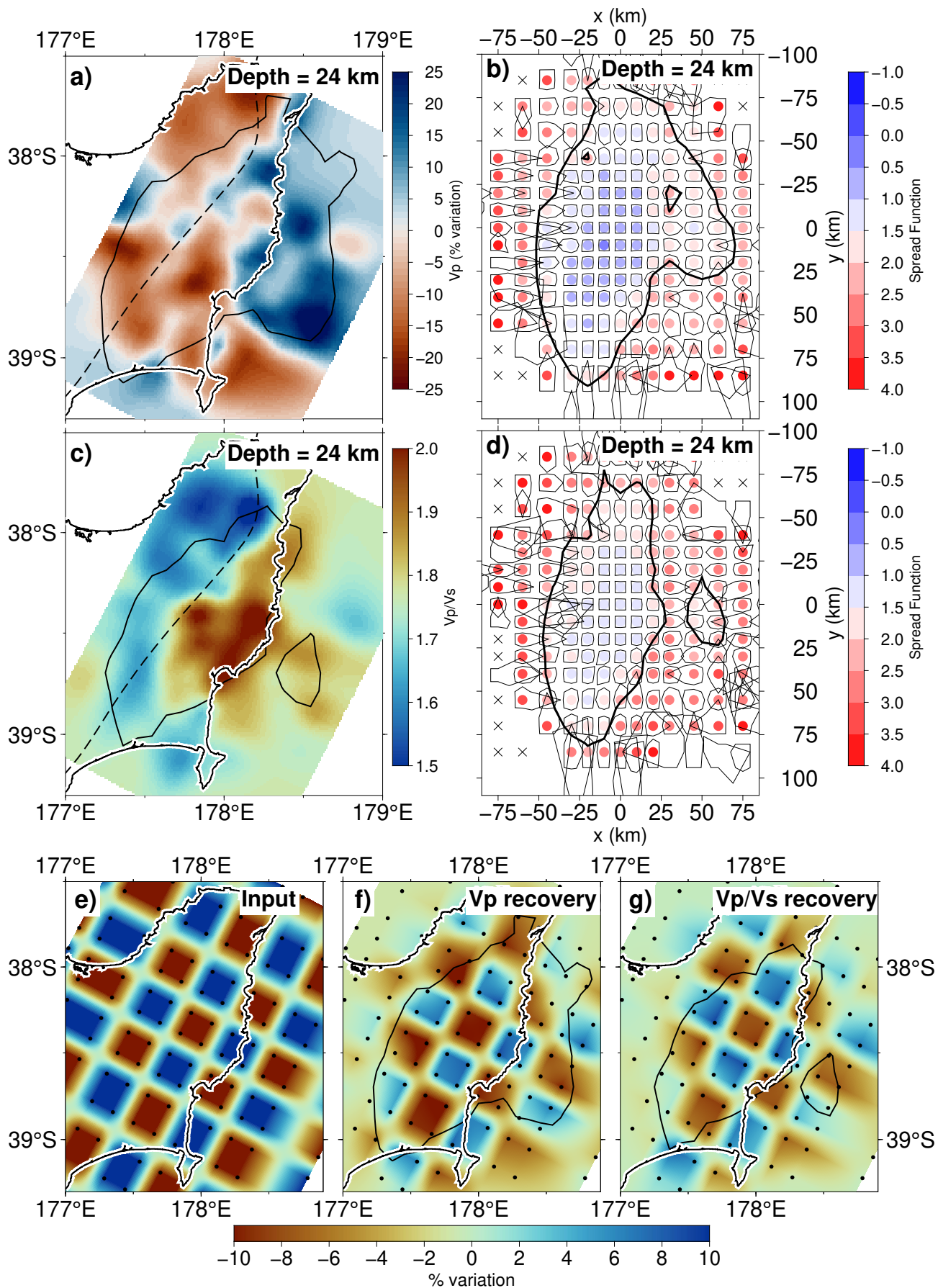


Figure caption continues on the next page

Figure 5.14: **a)** Final V_P model at 24 km depth. Black solid line: interpretable region defined by the spread function (contoured in geographic coordinates). Black dashed line: intersection of plate interface. **b)** Spread function for V_P model. Circles: spread function (SF) at each node. Polygons: 70 per cent contour of resolution kernel. Black solid line: interpretable region defined by $SF = 2.0$ (contoured in Cartesian coordinates). Crosses: fixed nodes. **c)** Final V_P/V_S model at 24 km depth. Black solid line: interpretable region defined by the spread function (contoured in geographic coordinates). Black dashed line: intersection of plate interface. **d)** Spread function for V_P/V_S model. Circles: spread function (SF) at each node. Polygons: 70 per cent contour of resolution kernel. Black solid line: interpretable region defined by $SF = 2.2$ (contoured in Cartesian coordinates). Crosses: fixed nodes. **e)** Input checkerboard ($\pm 10\%$) to calculate synthetic travel-times. **f)** 15 km anomaly checkerboard recovery for V_P from a 1-D starting model. Open circles: node positions. **g)** 15 km anomaly checkerboard recovery for V_P/V_S from a 1-D starting model. Open circles: node positions.

Table 5.2: Table summarising parameters of gradational inversion scheme

Model	Min. x-y spacing	Min. z spacing	No. earthquakes	Max. az. gap.	V_P damping	V_P/V_S damping
1-D	-	2 km	947	180	-	-
Coarse 3-D	15 km	4 km	2,104	210	200	500
Finer 3-D	10 km	3 km	2,104	210	200	500

5.14), and identifying the SF threshold that enclosed the well-resolved region recovered by the checkerboard tests. We consider nodes with SF values ≤ 2 for V_P and ≤ 2.2 for V_P/V_S to be interpretable. For further examples of spread function and checkerboard results, please refer to Appendix E.

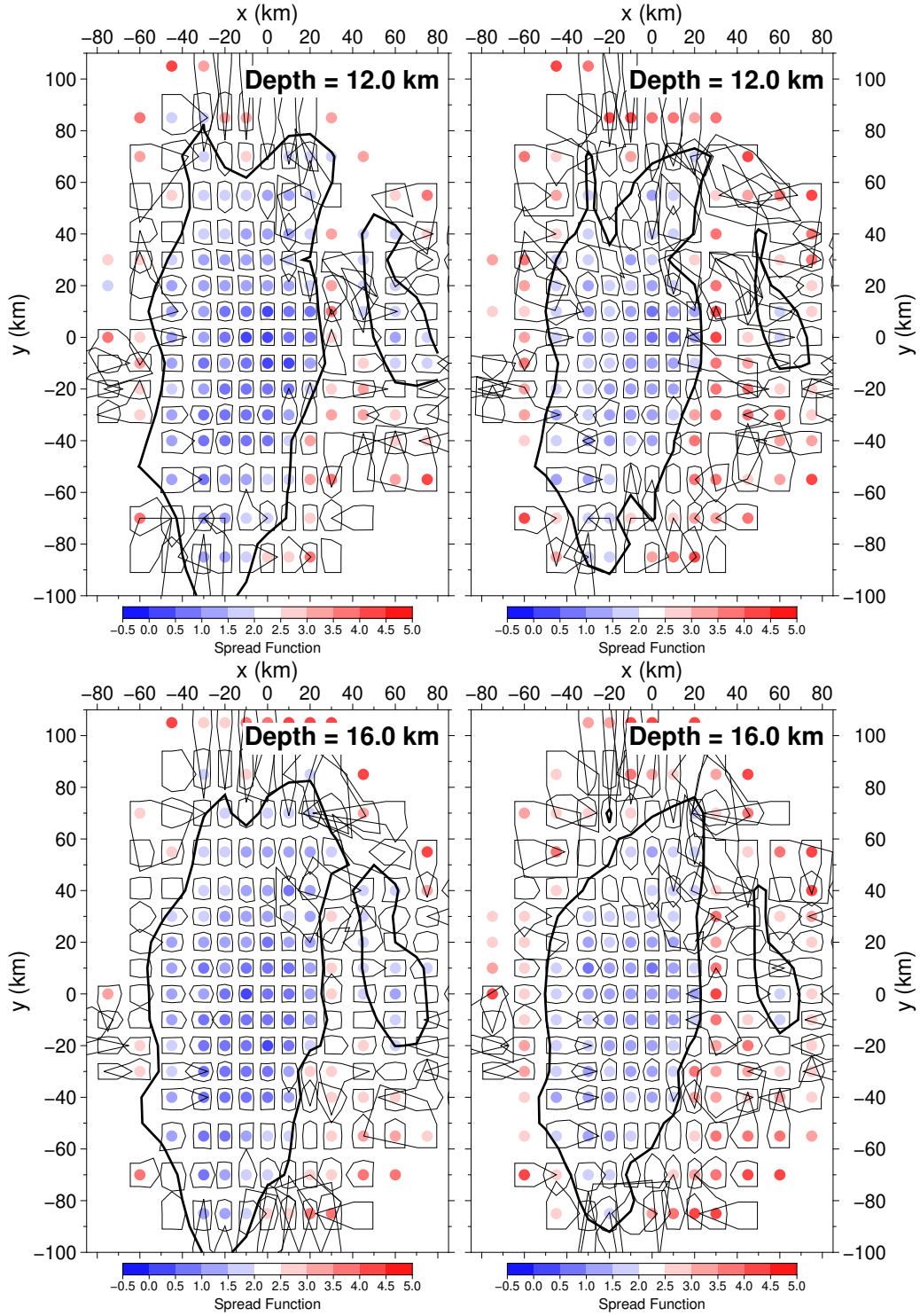


Figure 5.15: Left V_P spread functions at 12 and 16 km depth. Right V_P/V_S spread functions at 12 and 16 km depth. Refer to Figure 5.14 for further explanation.

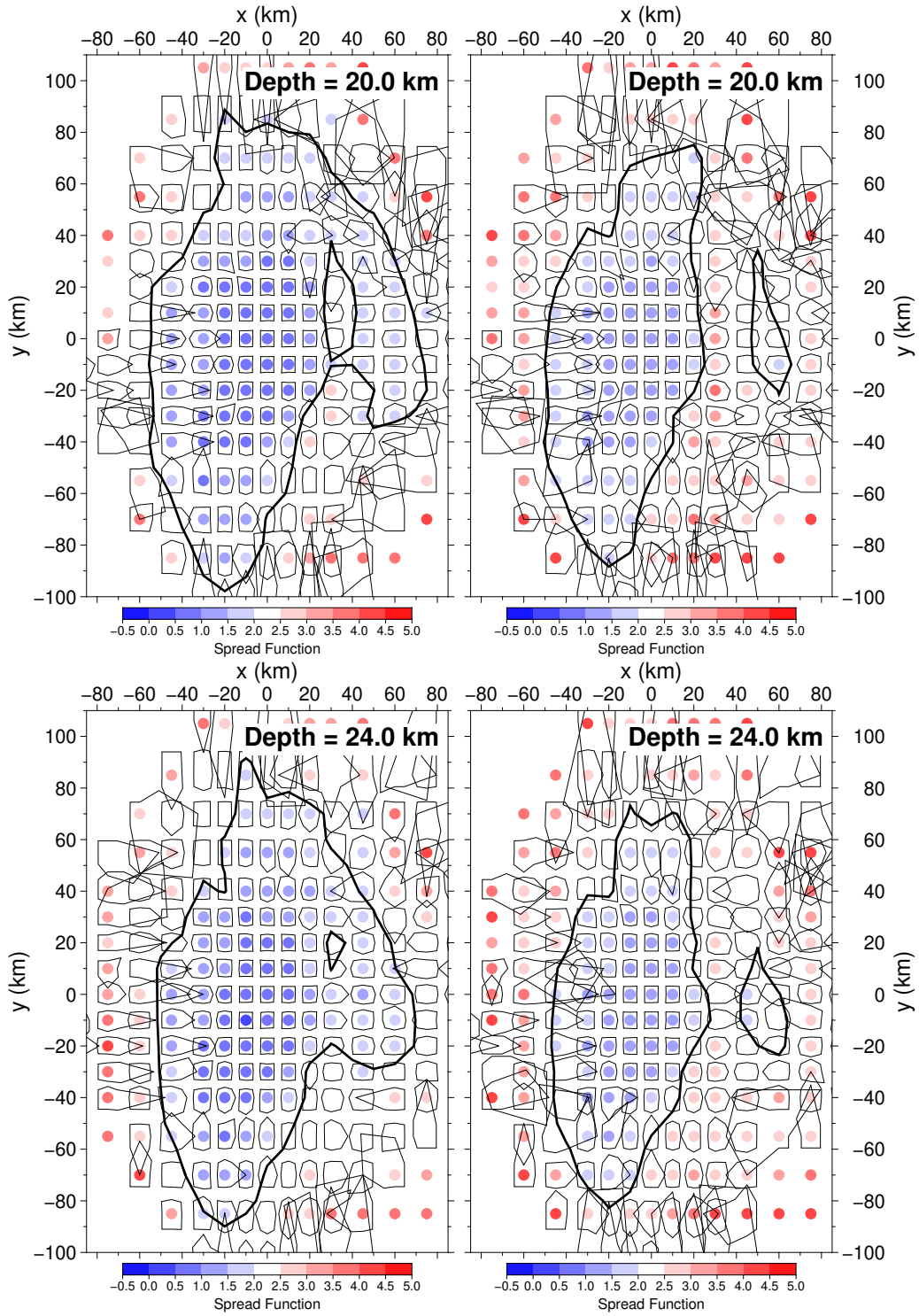


Figure 5.16: Left V_P spread functions at 20 and 24 km depth. Right V_P/V_S spread functions at 12 and 16 km depth

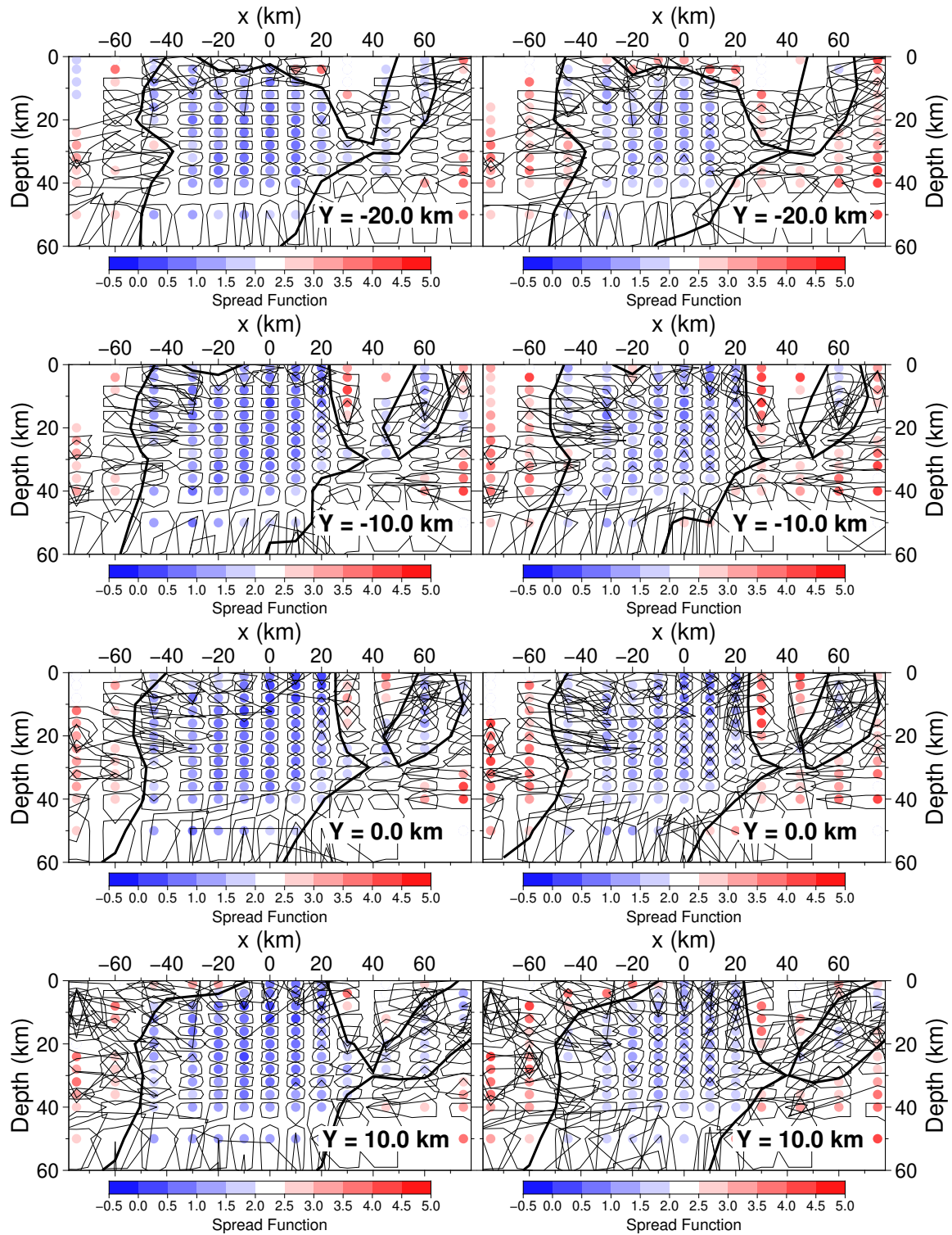


Figure 5.17: Left V_P model spread function at each node (circles) in cross section. Black contour: $SF = 2.0$. Right V_P/V_S model spread function at each node (circles) in cross section. Black contour: $SF = 2.2$.

5.4.3 Earthquake de-clustering

In general, it is recommended that tomographic inversions have a relatively uniform distribution of data in the model space, in terms of ray coverage, which is controlled by the distribution of earthquakes and stations. However, some studies have retained clusters of earthquake that are likely located in regions of high fluid pressure, interpreted from high V_P/V_S , resulting in low effective stress and abundant seismicity (e.g., [Mochizuki et al., 2021](#)). Similarly, dense networks of seismograph stations on volcanoes with clusters of seismicity are routinely used to determine V_P and V_P/V_S structure (e.g., [Lin et al., 2014](#)). Here, we test the effects of removing clusters of earthquakes on the final models.

We repeat the inversion procedure with a reduced catalogue where spatial clusters of earthquakes are removed to see the effects on the final model. To define earthquake clusters, we first calculate the minimum distance between all earthquake pairs, then use the Density-Based Spatial Clustering of Applications with Noise (DBSCAN) algorithm ([Ester et al., 1996](#)), which can identify irregularly-shaped clusters rather than one based on a centroid. We set the maximum distance between two earthquakes for one to be considered the neighbour of the other to 3 km. If 10 earthquakes are located within the neighbourhood of a core point, a cluster is identified. The combination of calculation of inter-event distances and clustering analysis allows us to retain small earthquake clusters of only a few events that are further away from other earthquakes. Figure 5.18 shows the initial catalogue, identified clusters, and final catalogue once clusters are removed. The clusters are primarily located close to the Gisborne array, therefore significantly reduces the number of rays in this region, without reducing data coverage (and therefore resolution) further afield. In total, eleven clusters are identified and removed, reducing the catalogue from 2,104 to 1,715 earthquakes. We still include all earthquakes recorded on the HOBITSS and permanent networks in the [Yarce et al. \(2019\)](#) catalogue, as used in the inversion with the full catalogue.

As in the full inversion routine, we shift the node positions to generate nine models, from which we calculate the mean velocity in increments of 0.5 km. We calculate the spread function and the 70 per cent contour of the resolution kernel, to compare node smearing between the two catalogues.

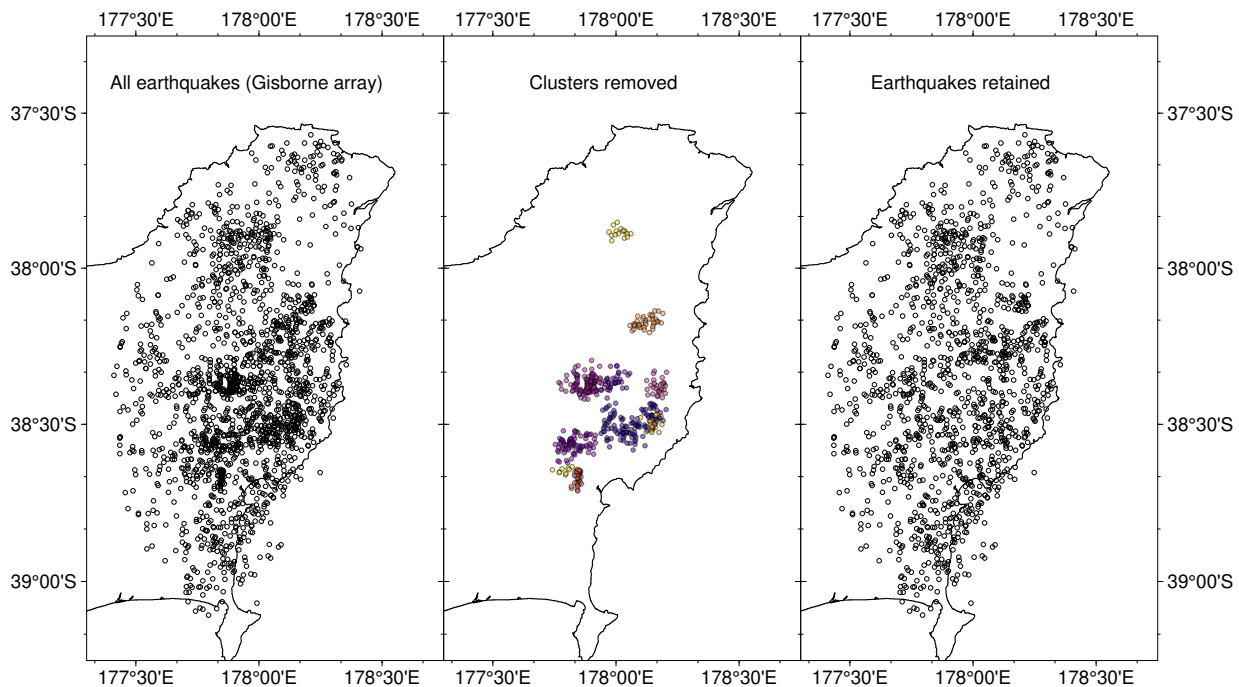


Figure 5.18: **Left panel)** Earthquakes with at least 12 P-picks and 1 S-P measurement. **Middle panel)** Earthquake clusters (11) removed following the DBSCAN procedure. Colours correspond to each cluster (plotted with 50% transparency). **Right panel)** Earthquakes retained for use in the inversion.

Inevitably when changing the data distribution in a tomographic inversion the final models change somewhat. However, features observed in the models from the full versus de-clustered catalogues are broadly consistent and the same interpretations can be made from each model. Here we show comparisons of model features interpreted in the discussion section, to demonstrate that these are not an artefact of earthquake clusters. We note that the structures and shapes of features are consistent between the models, but the de-clustered catalogue has reduced amplitude anomalies. Therefore, when we discuss features in the model, we focus on relative anomalies and interpret absolute values with caution.

Figure 5.19 shows the V_P model spread function and checkerboard recovery from the full and de-clustered catalogues. These are very consistent with each other, demonstrating that similar resolution and recovery is possible in both inversion routines and do not significantly skew the data. Red circles show the locations of key features discussed in Section 5.6, demonstrating that these features are similarly resolvable with both catalogues, and are likely not an artefact of the earthquake coverage.

In the next sections, we discuss and interpret the models derived from the full earthquake

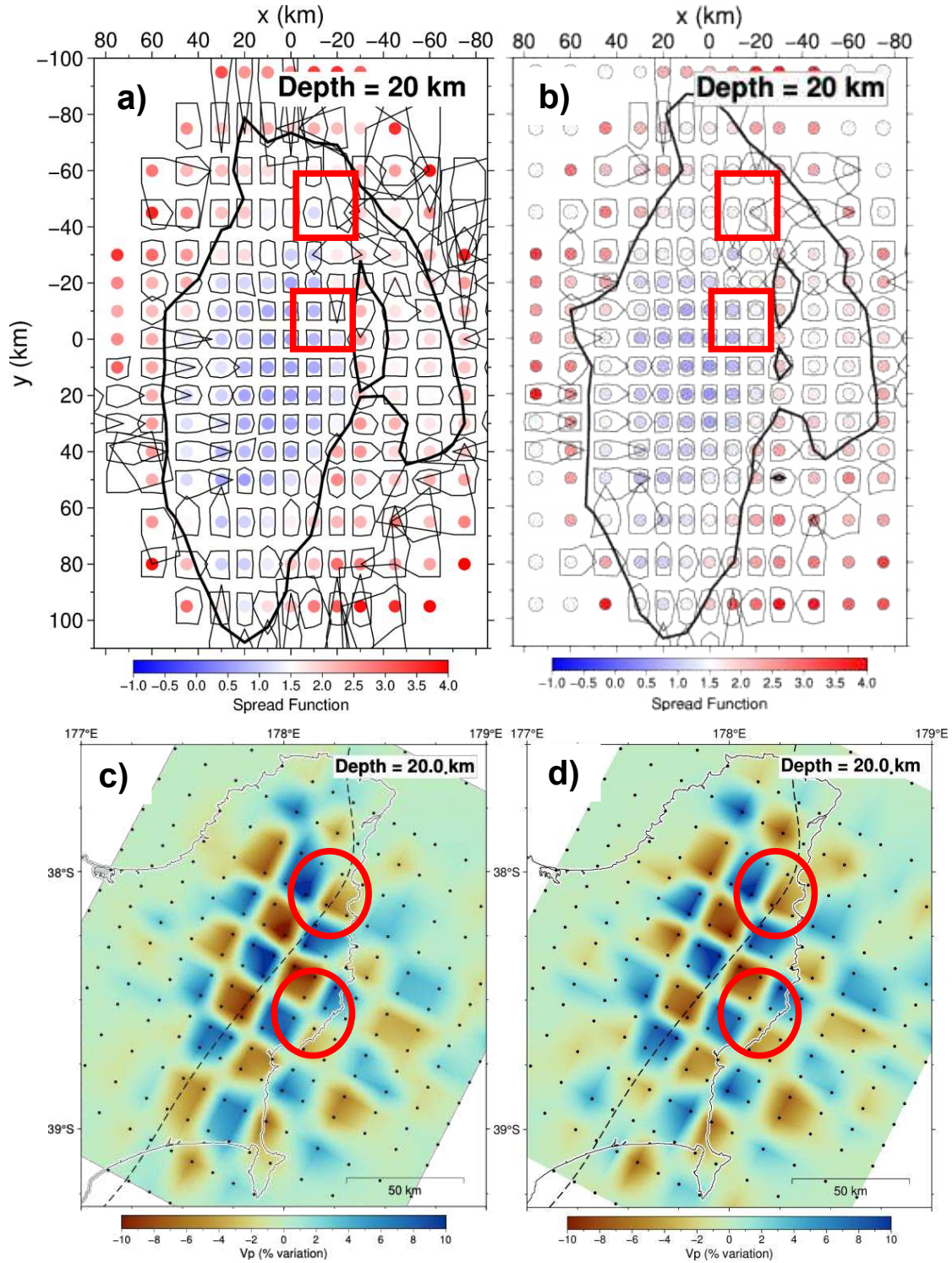


Figure 5.19: Full and de-clustered earthquake catalogue V_P spread function and checkerboard comparison. Red squares/circles: areas of model with features discussed in Section 5.6. **a)** Spread function for full earthquake catalogue inversion. **b)** Spread function for de-clustered earthquake catalogue inversion. **c)** Checkerboard recovery for full earthquake catalogue inversion. **d)** Checkerboard recovery for de-clustered earthquake catalogue.

catalogue, in favour of including as much data as possible, though additional tomograms are included in Appendix E for the de-clustered model outputs.

5.5 Results

5.5.1 Observed velocity structure

We consider the models to be interpretable to approximately 50 km depth. Laterally, interpretable V_P structure is concentrated in the onshore region to approximately 50 km inland, with an additional band of coverage near the offshore HOBITSS stations. Interpretable V_P/V_S structure is primarily onshore, due to fewer offshore S-P measurements. The gap in coverage between these two regions reflects limited station coverage (Figure 5.1) and a seismicity gap that has been interpreted as a region of higher plate coupling, both contributing to fewer ray paths sampling the region (Yarce *et al.*, 2019; Mochizuki *et al.*, 2021).

Map-view 3-D tomographic results are shown at depths of 12, 20, and 28 km (Figure 5.20). Margin-perpendicular cross-sections are shown with the plate interface model of Williams *et al.* (2013), derived from seismic reflection data and seismicity distributions (Figure 5.21).

Within the interpretable region, V_P generally increases with depth. A low- V_P anomaly is observed in the upper plate at approximately 15 km depth to the north-west of the Gisborne array (Figure 5.21). Upper-crustal V_P/V_S is heterogeneous, particularly near the coastline at depths shallower than ~8 km.

The subducting slab crust is characterised by high V_P/V_S (>1.8) to depths of ~30 km, below which V_P/V_S decreases. The high- V_P/V_S slab is strongly aligned with the regional plate interface model of Williams *et al.* (2013). The slab mantle, from ~15 km below the interface, generally has lower V_P/V_S with some localised regions of higher values (Figure 5.21). Along the plate interface, two pronounced high- V_P anomalies are observed near Te Puia and Gisborne (Figure 5.22).

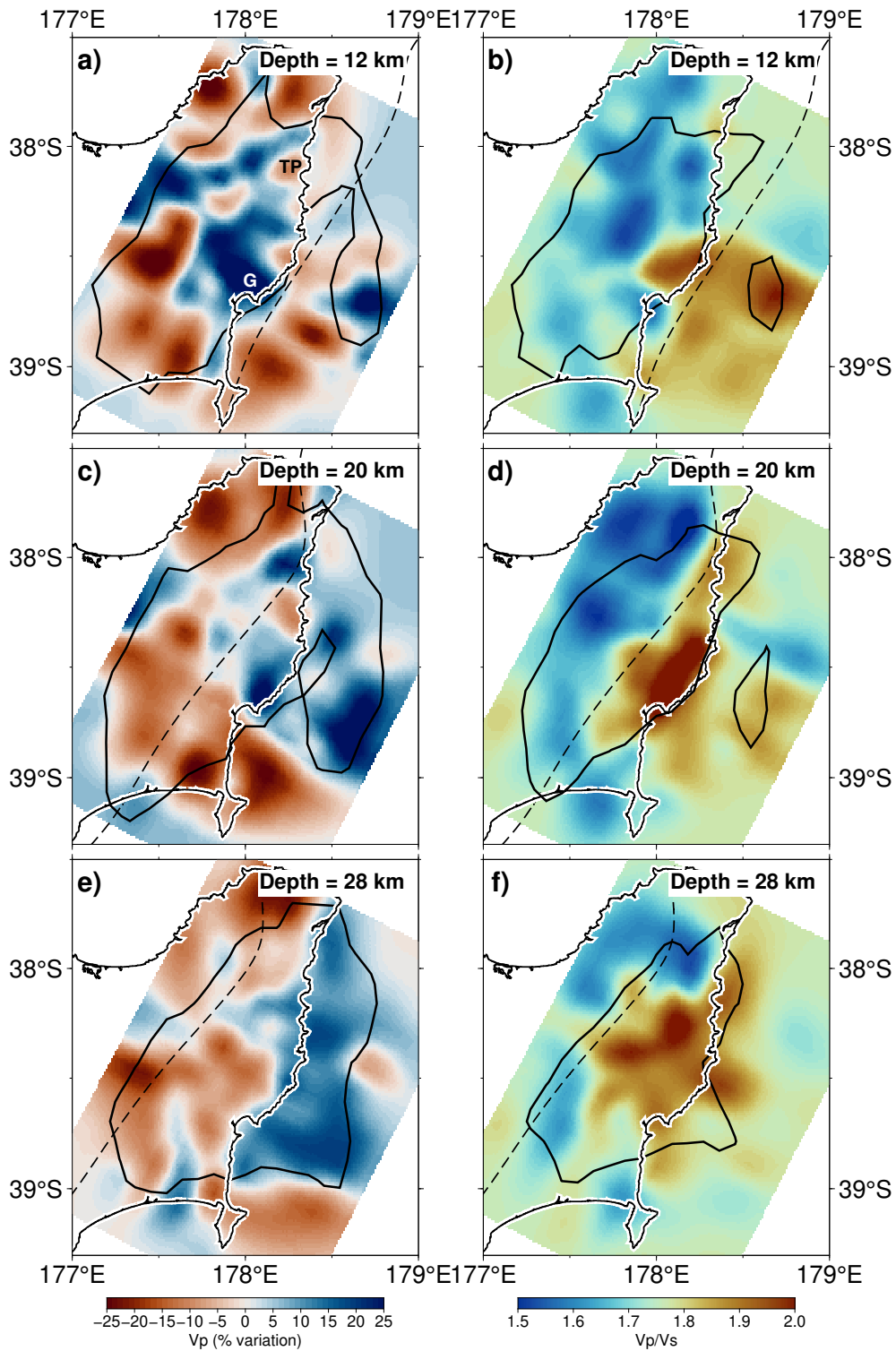


Figure 5.20: Map view V_P and V_P/V_S models for depth slices through the model. V_P models are shown as percentage variation relative to the mean depth for each layer across all nodes in that layer. For absolute V_P models, refer to supplementary materials. Black solid line: interpretable region defined by spread function. Black dashed line: intersection with plate interface model (Williams et al., 2013).

5.6 Discussion

The tomographic models show three striking features that are relevant to slow slip at the northern Hikurangi margin: heterogeneous upper plate structure, elevated slab V_P/V_S to depths of ~ 30 km, and high- V_P anomalies along the plate interface near Gisborne and Te Puia.

5.6.1 Upper plate

V_P broadly increases with depth across the region, but upper plate V_P and V_P/V_S are heterogeneous, particularly above plate interface depths of ~ 10 – 20 km, where surface seeps are observed. This heterogeneity may be due to the complex and overlapping interaction of several processes operating on different length-scales: eastward slumping of shallow sediments and the presence of clustered gas, oil, mud volcanoes, and hot springs (Figure 5.2; [Reyes et al. \(2010\)](#); [Litchfield et al. \(2022\)](#); [Reyes et al. \(2022\)](#)). Variable permeability pathways for fluid migration through the forearc also likely contribute to the observed heterogeneity, but the relative importance of each of these cannot be resolved unambiguously using passive source imaging.

A low V_P (label A; Figure 5.21) and low V_P/V_S (label B; Figure 5.21) anomaly is observed at ~ 15 km depth (Figure 5.21, $Y=0$, 10, and 20 km slices). Low V_P and low V_P/V_S is typically associated with the presence of gas or relatively dry fracture zones. However, as gas sources in this region are typically very shallow (e.g., [Crutchley et al., 2010](#)), this may be more likely associated with relatively dry fracture zones.

5.6.2 Subducting slab

In agreement with [Eberhart-Phillips and Bannister \(2015\)](#), the subducting slab crust generally has high V_P/V_S (>1.8), whereas the slab mantle, from ~ 15 km below the interface, has lower V_P/V_S with some localised regions of higher values (Figure 5.21). A similar V_P/V_S structure for the de-clustered catalogue is observed (see Appendix Figure E.21). The high V_P/V_S slab is strongly aligned with the regional plate interface model of [Williams et al. \(2013\)](#). Although that regional plate interface model is relatively smooth, plate interface structure is likely

to be heterogeneous on local scales of kilometres to tens of kilometres owing to subducted seamounts, underplated sediments, and variable fluid distributions (e.g., [Barnes et al., 2020](#); [Bangs et al., 2023](#); [Leah et al., 2022](#); [Reyes et al., 2022](#)). The high V_P/V_S is observed a few kilometres above the plate interface model, similar to [Eberhart-Phillips and Bannister \(2015\)](#), interpreted as subducted sediment and possible dewatering.

The downdip extent of the high V_P/V_S in the subducting slab varies along strike. Within the well-resolved part of the model (shown in (Figure 5.21 with further examples in Appendix E) the downdip limit of the high V_P/V_S region ranges between 30 and 40 km depth. Further north and south, the downdip limit of the high V_P/V_S is shallower, though we note that this is towards the resolution limit of the models.

5.6.3 Subducted seamounts

Our tomography shows two pronounced high- V_P anomalies along the plate interface near Te Puia and Gisborne (Figure 5.22). Both are spatially consistent with previous speculated subducted seamounts (Figure 5.2) based on magnetic, receiver function, and offshore seismic data ([Hunt and Glover, 1995](#); [Leah et al., 2022](#); [Bangs et al., 2023](#)). We discuss each region in turn below.

[Chesley et al. \(2021\)](#) demonstrated that the internal structure of seamounts is likely heterogeneous, consisting of a resistive core embedded in a matrix of conductive material. There is some debate regarding the V_P and V_P/V_S structure of subducted seamounts, and these effects likely vary with depth. [Arai et al. \(2020\)](#) presented two interpretations of the V_P structure associated with the offshore Pāpaku seamount (labelled P, Figure 5.22). The first interpretation suggested that subducted seamounts are fractured, low- V_P structures with high V_P observed downdip of the seamount due to strong compaction in the wedge ahead of the seamount. This is similar to observations at other subduction zones, such as the subduction of the Amami Plateau at the Ryukyu subduction zone and the subduction of the Louisville Ridge at the Tonga-Kermadec subduction zone ([Arai et al., 2017](#); [Bassett and Watts, 2015](#)). At the Hikurangi margin, this interpretation is consistent with strong positive magnetic anomalies and full waveform inversion studies for offshore seamounts ([Barker et al., 2018](#); [Bangs et al.,](#)

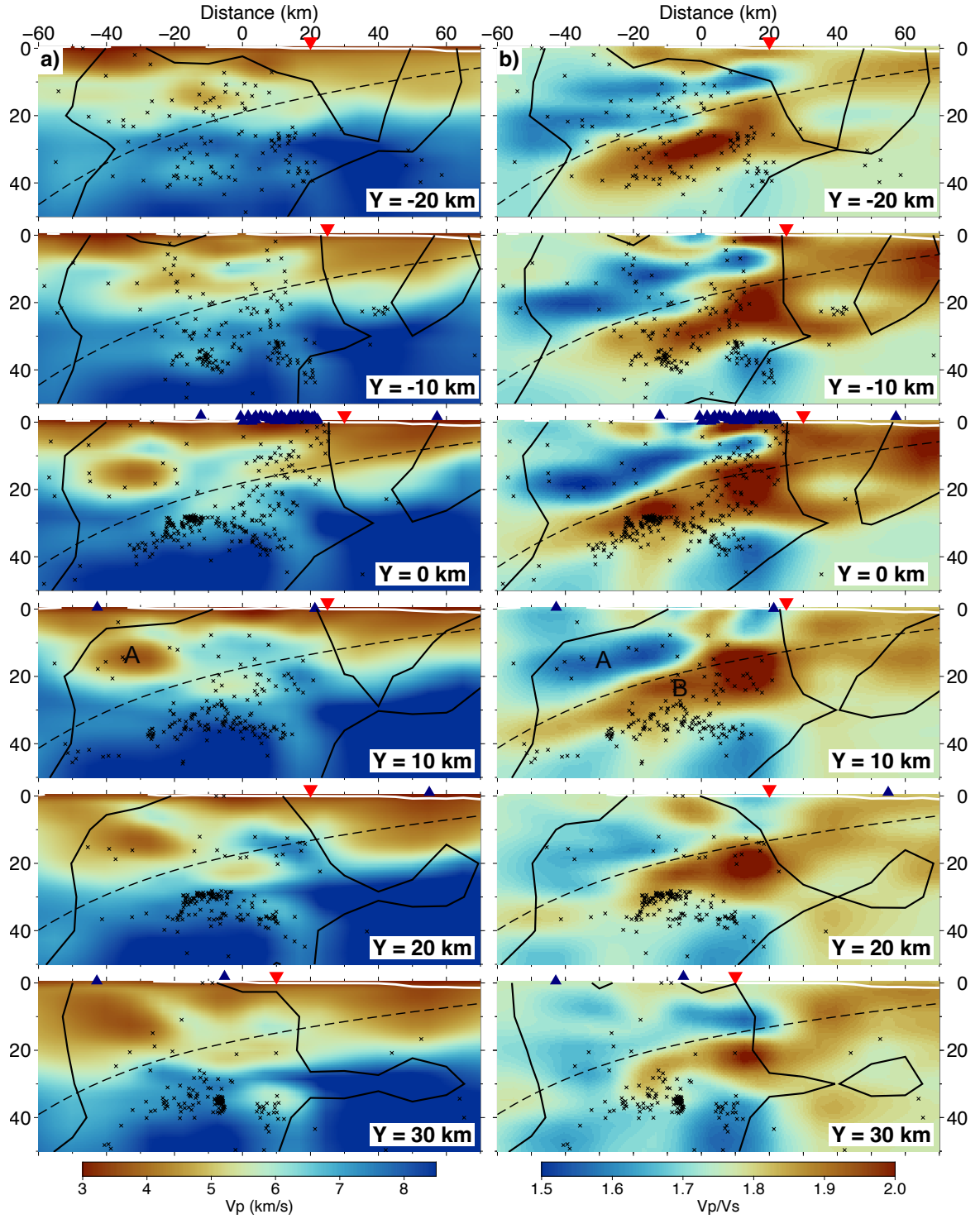


Figure 5.21: V_P (left) and V_P/V_S models in cross section from north to south. Dashed black line: plate interface model (Williams *et al.*, 2013). lines: spread function (SF) contours for V_P (SF = 2) and V_P/V_S (SF = 2.2). Navy triangles: seismometers. Inverted red triangle: coastline. Crosses: relocated earthquakes. A and B denote features discussed in 5.6.1.

2023), and drilling data suggesting that seamounts here are formed of fractured and hydrated smectite-rich volcanoclastic conglomerate interbedded with basalt with $V_P \approx 4 \text{ km s}^{-1}$ (Gase *et al.*, 2023). The second interpretation from Arai *et al.* (2020) suggested that seamounts consist of high V_P anomalies with an updip lower V_P channel in their wake, though this is not consistent with Hikurangi margin drilling data (Arai *et al.*, 2020; Gase *et al.*, 2023).

Gisborne seamount

Hunt and Glover (1995) report a positive magnetic anomaly of +70 nT near Gisborne, in Tūranganui-a-Kiwa/Poverty Bay (labelled G, Figure 5.22). We observe a high V_P anomaly downdip of this, which we suggest is downdip of a subducted seamount, in line with the velocity interpretation of seamounts in Arai *et al.* (2020). A potential seamount at this location was speculated by Leah *et al.* (2022) and Bangs *et al.* (2023). Leah *et al.* (2022) used P-to-S receiver functions to infer a local plate interface topographic high ($\leq 4 \text{ km}$) beneath the south-east part of the Gisborne network, whereas Bangs *et al.* (2023) interpreted an offshore sediment lens trailing a hypothesised seamount immediately offshore Gisborne. These studies place the possible seamounts within $\approx 10 \text{ km}$ of one another, which is consistent with different methods imaging different parts of a broader tens-of-kilometres-scale feature. This is consistent studies that suggest seamount width is similar to indentation width ($\sim 40 \text{ km}$, Pedley *et al.* (2010)) and the width of the V_P anomaly, and would be similar in width to the largest known subducted seamount, the Papāku seamount (Figure 5.2).

The Gisborne region also contains abundant surface mud volcanism in the Waimata Valley (Figure 5.2 Ridd, 1970; Leighton, 2020), which is likely the surface expression of overpressured clays along the East Coast Allochthon. The difference in surface seep compositions between the Waimata Valley and Te Puia region suggests a different fluid source. Te Puia hot springs have a higher proportion of mantle-derived fluids, while the Waimata Valley surface seeps are comprised primarily of mud volcanoes. One possible explanation is that the Gisborne seamount differs in composition or alteration state, resulting in different surface fluid signatures. Here, the resistivity model does not show high conductivity at the plate interface, but there is a conductive region at 7.5 km depth that may represent fluid conduits that feed the mud

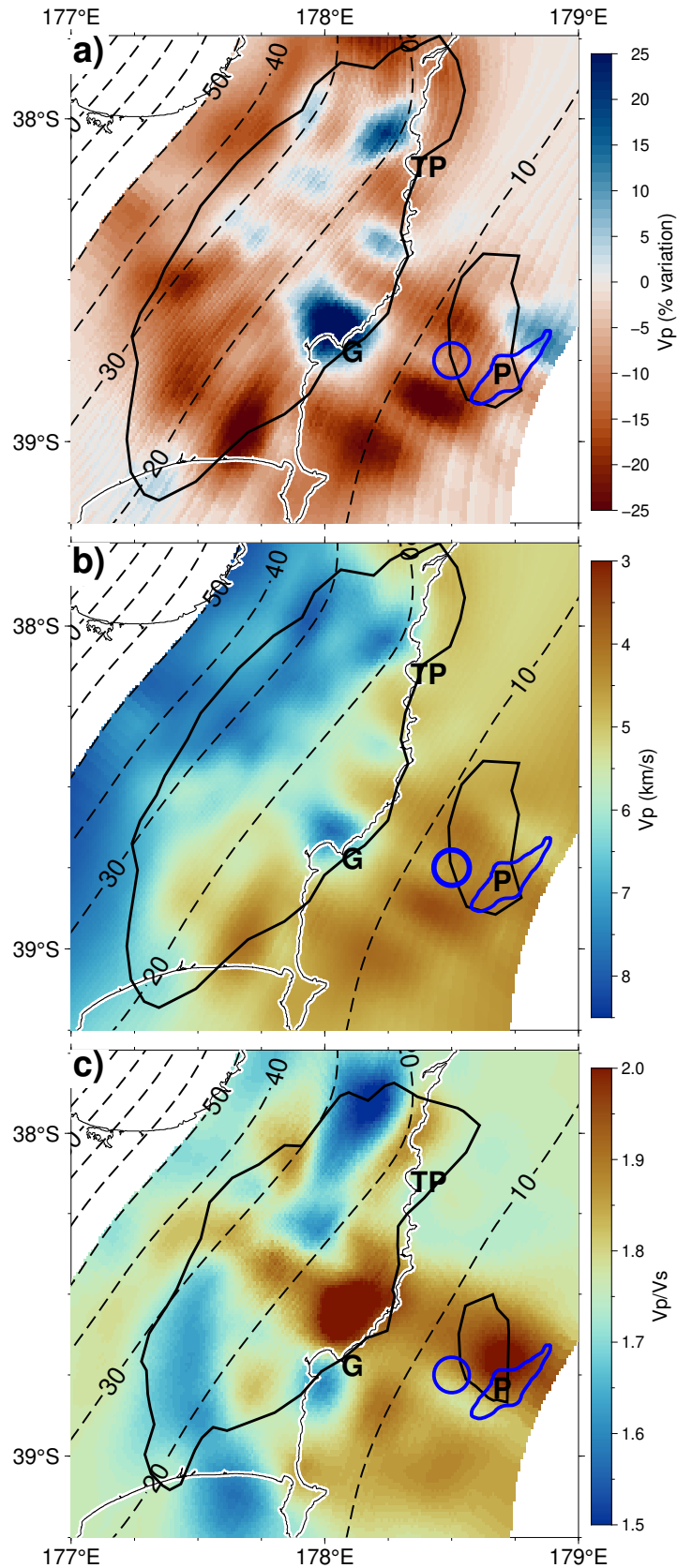


Figure 5.22: a) V_P model across plate interface defined by *Williams et al. (2013)*, from 4–50 km depth as a percentage variation in V_P relative to the starting 1-D model. Black solid line: spliced spread function contours in 4 km depth slice increments. Black dashed line: plate interface depth. Blue solid shapes: mapped seamounts within the interpretable region of our V_P model (*Barker et al., 2018; Bangs et al., 2023*). P: Pāpaku seamount. b) Absolute values of V_P model across the plate interface. c) V_P/V_S across the plate interface.

volcanoes ([Heise et al., 2023](#)). Margin-parallel normal faults, collocated with mud-volcano eruption sites ([Mazengarb and Speden, 2000](#)), may be enhanced by seamount subduction and associated uplift in the upper plate.

The Tūranganui-a-Kiwa/Poverty Indentation lies to the south-east of the purported Gisborne seamount (Figure 5.2). This $\sim 4000 \text{ km}^2$ structural re-entrant includes the Tūranganui-a-Kiwa/Poverty Canyon system and has been interpreted as the result of seamount subduction-induced debris flow and avalanche processes ([Pedley et al., 2010](#)). [Pedley et al. \(2010\)](#) suggested that a train of seamounts has subducted here since $\sim 1\text{--}2 \text{ Ma}$. Based on the margin convergence rate, that interpretation is consistent with one seamount presently beneath the land and further seamounts offshore ([Pedley et al., 2010](#)). This is broadly consistent with our inferred Gisborne seamount (Figure 5.22), and with the additional offshore seamounts interpreted by [Bangs et al. \(2023\)](#). The orientation of this train of seamounts is consistent with the plate convergence direction, and the orientation of the Tūranganui-a-Kiwa/Poverty Indentation. Experimental modelling suggests that indentation width is similar to seamount diameter, suggesting that the subducted seamount would be $\approx 40 \text{ km}$ wide ([Collot et al., 2001](#)).

Te Puia seamount

In the V_P model inverted from the full earthquake catalogue, we observe a high V_P anomaly downdip of a speculated seamount near Te Puia hot springs. This anomaly is absent in the inversion for V_P from the de-clustered catalogue, though this may be because this significantly reduces the number of raypaths towards the station closest to the anomaly, which is at the edge of the network. Both models display good checkerboard recovery and spread functions in this region of the model, yet display different V_P signals. However, a positive V_P anomaly here is consistent with complementary geophysical studies. For example, [Hunt and Glover \(1995\)](#) report a magnetic anomaly of $+140 \text{ nT}$ near Te Puia in Tokomaru Bay (labelled TP, Figure 5.22). Their modelling suggests that this anomaly could be explained by a seamount at 15 km depth, $\approx 65 \text{ km}$ in diameter with a relief of 5 km ; the absence of a significant local gravity anomaly supports a source at depth ([Hunt and Glover, 1995](#)). In the same location, [Heise et al. \(2023\)](#) observe a conductive patch ($\sim 30 \Omega\text{m}$), interpreted as a fluid and sediment rich area.

These would place the inferred seamount immediately updip of the high V_P anomaly that we observe from 18–24 km depth, consistent with the modelling of [Hunt and Glover \(1995\)](#).

Geochemical data from the Te Puia hot springs demonstrate mantle-derived helium isotopes, indicating fluid pathways from the slab mantle to the surface ([Reyes et al., 2022](#)). Localised uplift of subducted seamounts due to their relatively buoyancy (e.g., [Arai et al., 2017](#)) and the consequent development of complex fracture networks (e.g., [Wang and Bilek, 2011](#)) may act to enhance these fluid pathways both from slab mantle to plate interface and from the plate interface to the upper plate.

The $\sim 3,100 \text{ km}^3$ Ruatoria debris avalanche (Figure 5.2) may have been triggered by seamount subduction that destabilised the East Coast ([Collot et al., 2001](#)), although [Lewis et al. \(2004\)](#) argued that the relevant seamount may have been located further north, near East Cape, $\sim 20 \text{ km}$ north of the Te Puia seamount. No seamount has been interpreted in seismic reflection data at East Cape so far. Our tomography does not resolve the broader question of whether the avalanche was caused by destabilisation due to a subducted seamount, but it is consistent with substantial interface roughness in the Te Puia region. Furthermore, the Te Puia hot springs are located in a region experiencing active sliding of Late Cretaceous blocks towards the coast at a rate of 55 mm/yr ([Mazengarb and Speden, 2000](#)) and the lubrication of overpressured fluid clays along the East Coast Allochthon has been described as a contributing factor to the destabilisation of the East Coast resulting in the Ruatoria landslide ([Lewis et al., 2004](#)).

Implications for persistent interface roughness

The offshore Pāpaku seamount has been characterised with magnetic and controlled-source seismic data (e.g., [Barker et al., 2018](#); [Davy et al., 2021](#)). In our model, we do not observe a strong, high V_P anomaly downdip of Pāpaku, although the seamount lies close to the edge of the interpretable region of our model and so this absence should be treated cautiously. We do, however, observe a high V_P/V_S anomaly at and downdip of Pāpaku. Similar anomalies are also present at and downdip of the inferred Te Puia and Gisborne seamounts. Taken together, these observations are consistent with a possible seamount-related influence on forearc elastic properties, but all three features lie within a broader region of elevated background V_P/V_S , so

at the present model resolution, we cannot isolate a unique seamount-derived contribution in the V_P/V_S model.

While we cannot distinguish whether the high V_P anomalies that we observe represent the seamounts themselves or compacted sediment downdip of them, our results support the existence of subducted seamounts near Te Puia and Gisborne. More generally, they suggest that plate interface roughness persists to at least 20 km depth beneath the onshore margin. This has implications for geodetic and other inversion schemes that depend on the assumed geometry and elastic heterogeneity of the plate interface, as demonstrated by [Williams and Wallace \(2015\)](#).

5.6.4 Slow slip events controlled by slab hydration

The majority of SSEs at the northern Hikurangi margin occur offshore to depths of ~2–15 km depth and possibly all the way to the trench (Figure 4.1; [Wallace et al., 2016](#)), with occasional deeper events extending to 20–35 km depth (e.g., Chapter 4; [Wallace, 2020](#); [Woodward et al., 2026b](#)). SSE characteristics also vary along-strike, with increasing recurrence interval towards the south ([Perez-Silva et al., 2022](#)), possibly associated with variation in slab geometry, plate convergence rate, upper plate terrane, or pore pressure ([Li et al., 2018](#); [Perez-Silva et al., 2022](#)). We observe a trench-perpendicular contrast in V_P and V_P/V_S trending NE-SW, between Gisborne and Mahia Peninsula at 20 km depth (e.g., Figure 5.20 c/d) and in the plate interface V_P/V_S model. It is possible that this is associated with along-strike variation in fluid content, though this contrast is towards the edge of the interpretable region and strongest outside our interpretable region so should be treated with caution. Future work to constrain along-strike V_P and V_P/V_S variation, encompassing the entire central-northern Hikurangi shallow SSE region may help to constrain along-strike variation in V_P and V_P/V_S and possible relationship to SSE segmentation.

Offshore SSEs (Figure 4.1) recorded during temporary ocean-bottom seismometer and absolute pressure gauge deployments have been particularly well characterised (e.g., [Woods et al., 2024](#)), associated with temporal variations in fluid pressure and microseismicity that have been interpreted as a fault-valving mechanism ([Warren-Smith et al., 2019](#); [Zal et al., 2020](#)).

We observe maximum V_P/V_S at the plate interface and within the slab at ~ 20 km depth, close to the downdip limit of the shallowest SSEs (Figure 5.21). V_P/V_S then decreases downdip, transitioning to a low V_P/V_S region by ~ 35 – 40 km depth, coincident with the downdip limit of the less frequent, deeper SSEs. This is consistent with plate interface resistivity, which is conductive to ~ 25 km depth, then more resistive downdip, interpreted as a lack of fluid-rich conductive sediments (Heise *et al.*, 2023). These patterns are broadly consistent with a reduction in slab-derived fluid availability with depth.

Eberhart-Phillips and Bannister (2015) and Mochizuki *et al.* (2021) observed high seismicity in the high V_P/V_S slab crust and interpreted these regions as fluid-rich. We observe a similar relation (e.g., Figure 5.21, panel Y=0 km). Eberhart-Phillips and Bannister (2015) suggested that the highest slip rates occur where the slab may act as a reservoir where fluids accumulate and dilatant strengthening is enhanced. Similarly to Mochizuki *et al.* (2021), we observe high seismicity in the region of highest V_P/V_S , which they speculated represented vertical extensional fracture zones that discharge overpressured fluids. The high V_P/V_S in the subducting slab may represent high fluid pressures and dehydration fluids trapped below an impermeable plate interface (Reyners and Eberhart-Phillips, 2009). We note, however, that while fluids exert first-order control on V_P/V_S , fluid concentrations and distributions may vary through time due to fault-valving processes related to SSE cyclicity and our earthquake travel-time data includes earthquakes during and between SSEs (e.g., Warren-Smith *et al.*, 2019; Zal *et al.*, 2020).

Mochizuki *et al.* (2021) observe high V_P/V_S in the offshore portion of the subducting slab. Though this is outside the interpretable region in our models, this is consistent with high fluid content in the slab at all depths at which SSEs are observed, and suggestions that slow slip may continue all the way to the trench (Wallace *et al.*, 2016). The broadly high V_P/V_S is also consistent with numerical modelling that suggested low V_S , likely due to sediments, could explain the unusually long-duration ground motion recorded along the East Coast following the 2016 M_w 7.8 Kaikōura earthquake (Kaneko *et al.*, 2019).

Subduction zone fluids are from a variety of depth-dependent sources, including meteoric water (uppermost km), subducted seawater, clay dehydration (initiating >10 km depth) and mantle-derived gases (Ellis *et al.*, 2015; Reyes *et al.*, 2022). Their relative contributions vary

downdip as temperature and pressure increase.

Hot springs in this region contain relatively high proportions of mantle-derived fluids ([Reyes et al., 2022](#)). [Hunt and Glover \(1995\)](#) also noted that strongly mineralised waters surface along the East Coast in areas associated with positive magnetic anomalies and subducted seamounts. This suggests that rough plate interface structure and fluid availability may not be independent controls on slow slip. Instead, seamount subduction may help to maintain fluid pathways and to localise regions of elevated fluid pressure at the plate interface, as summarised in the schematic shown in Figure 6.3. Furthermore, the seamounts themselves are formed of fluid-rich volcanoclastics, which contribute additional fluids, while their buoyancy (relative to surrounding oceanic crust) likely enhances plate roughness through localised uplift. [Chesley et al. \(2021\)](#) suggested that seamounts of dimensions similar to Tūranganui Knoll may subduct 3.2–4.7 times more water compared with unfaulted oceanic lithosphere. Similar associations between seamount subduction and shallow SSE occurrence have been suggested in other subduction zones, such as the central Ecuadorian subduction zone and the Nankai trough off Muroto ([Segovia et al., 2018](#); [Flores et al., 2024](#)). Numerical modelling of subduction of successive seamounts, similar to that seen near Gisborne, has also simulated shallow SSE occurrence ([Gauthier et al., 2026](#)).

In contrast, at the southern Hikurangi margin, where SSEs are deeper and less frequent, the margin is locked at shallow depths. Here, any seamounts (if they exist) are obscured from seismic imaging by <5 km sediment cover. Compared to the northern margin, samples from surface seeps at the southern margin suggest increased residence time, proportions of seawater and modified porewater fluids, and a decreased relative percentage of waters with chemical signatures associated with clay dehydration and serpentinisation ([Reyes et al., 2022](#)). This suggests that seamount subduction plays a key role in fluid cycling, concentration, and compositions, resulting in varied SSE characteristics.

Taken together, our results suggest that the largest and shallowest SSEs occur where slab V_P/V_S is highest (~20 km depth), whereas deeper and less frequent SSEs occur where slab V_P/V_S is lower but still elevated relative to the surrounding rock. Below ~35 km depth, slab V_P/V_S decreases further and observed SSEs become rare. This is consistent with fluid availability

contributing to the downdip limit of SSEs, although V_P/V_S alone cannot uniquely identify fluid source or abundance. Future work that better constrains smaller-magnitude SSEs may help test whether SSE frequency and magnitude vary systematically with depth-dependent fluid availability.

5.7 Conclusions

We present a high-resolution 3-D V_P and V_P/V_S model for the Raukūmara Peninsula using a temporary onshore seismograph network. The model is interpretable from 8–50 km depth, ~60 km onshore, and up to 40 km offshore. It therefore complements previous tomographic studies based on temporary offshore seismograph networks and controlled-source seismic sources, and improves constraints on the deeper part of the northern Hikurangi slow slip region.

We observe a high V_P anomaly at the plate interface near Gisborne, which we suggest are associated with subducted seamounts. This suggests that plate interface roughness persists to at least 15 km depth beneath the onshore margin; this roughness may provide regions of favourable fluid storage and enhance fluid migration from the subducting slab into the upper plate. Plate roughness may directly account for the heterogeneity of surface seeps observed along the East Coast. The locations of surface seeps are correlated with the subducted seamounts, suggesting that seamounts enhance fluid pathways from a range of depths, accounting for the variable geochemical signature of the seeps. Previous figures are summarised in Figure 5.23, showing these observations and interpretations side by side.

The region in which SSEs occur, where the plate interface is shallower than 35 km, is characterised by high V_P/V_S structure, which we interpret as evidence for high fluid pressures in the subducting slab crust and at the plate interface. Slab V_P/V_S is highest at plate interface depths of ~20 km, and decreases downdip. We interpret this pattern to indicate that a hydrated and fractured subducting slab supplies fluids to the plate interface in the shallow slow slip region, enhanced where the plate interface is roughest, helping to generate the high fluid pressures associated with SSEs and related microseismicity. The downdip decrease in slab V_P/V_S suggests a reduction in slab-derived fluids at depths where SSEs become less frequent

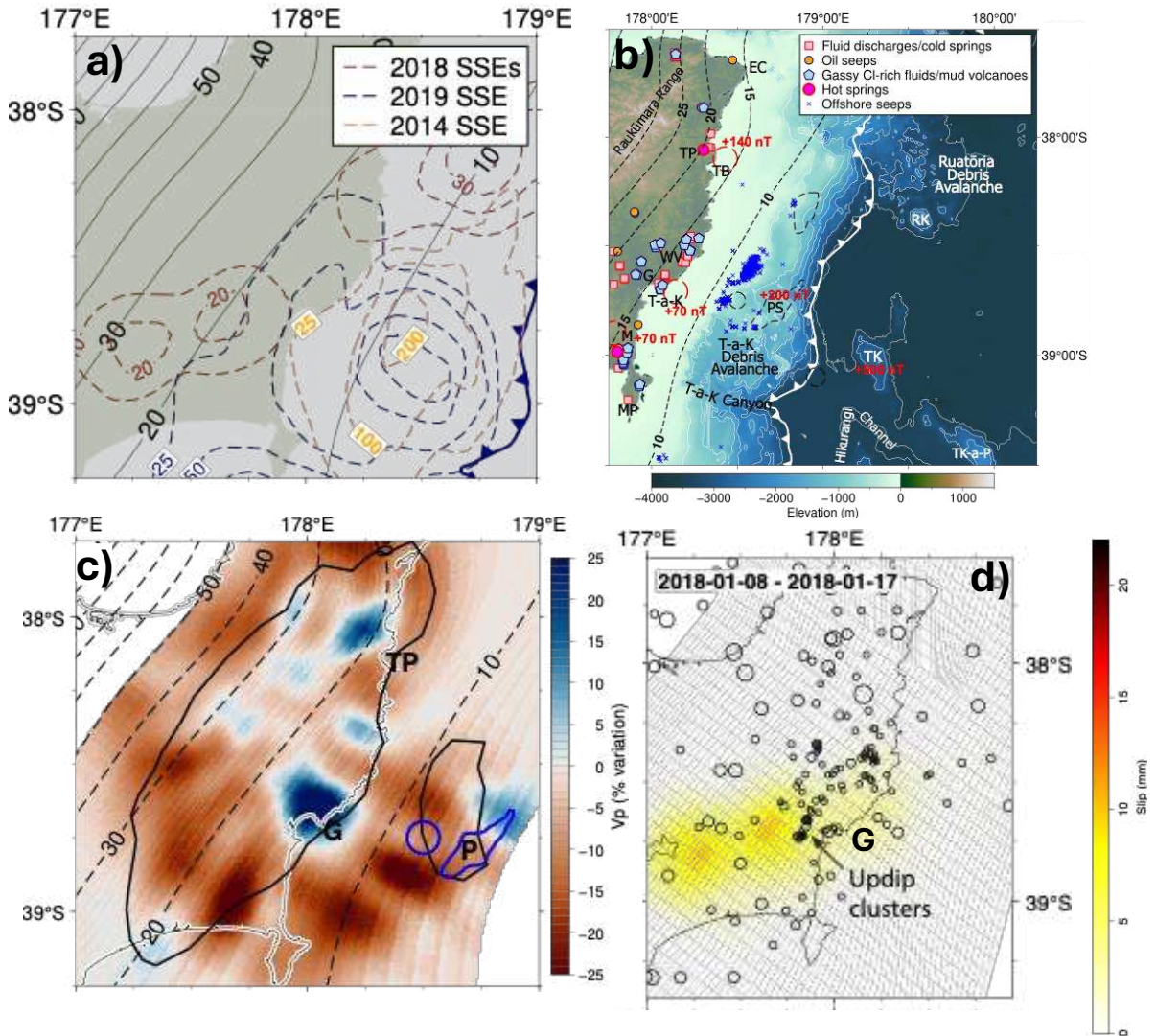


Figure 5.23: Summary figure of previous observations and interpretations. **a)** Navy line: trench (Bird, 2003). Black lines: plate interface depth (Williams et al., 2013). Contours: SSEs from previous studies and reported in this thesis (Woods et al., 2024; Wallace et al., 2016). **b)** Bathymetry and magnetic anomaly map. Dashed black lines: seamount locations constrained from magnetic or seismic reflection data (Bell et al., 2010; Barker et al., 2018; Bangs et al., 2023). Abbreviations: EC: East Coast; G: Gisborne; M: Morere; MP: Mahia Peninsula; RK: Ruatōria Knoll; T-a-K: Tūranganui-a-Kiwa; TB: Tokomaru Bay; TK: Tūranganui Knoll; TP: Te Puia; WV: Waimata Valley. **c)** Plate interface V_P model as percentage relative to depth-layer mean. Solid black line: well-resolved region as defined by the spread function. Solid blue lines: seamount locations (Bell et al., 2010; Barker et al., 2018). Abbreviations: TP: Te Puia; G: Gisborne; P: Pāpaku. **d)** Location of earthquake clusters relative to SSE-1 and Gisborne (labelled G).

and are no longer observed, and indicates that fluid availability is likely an important control on the downdip limit of SSEs.

CHAPTER 6

Concluding remarks

In this thesis, we addressed the following research questions:

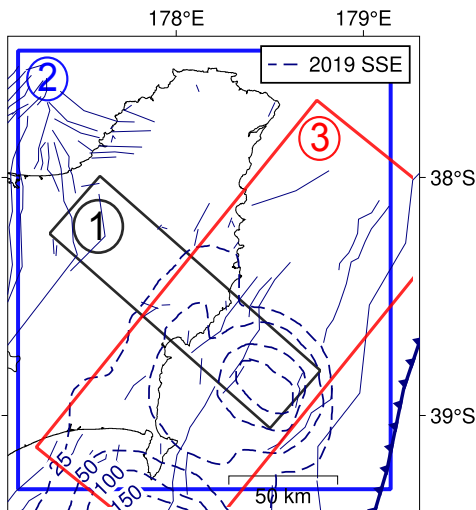
1. What is the character of the deeper (>10 km) part of the SSE zone, and how does it compare with the well-constrained shallow zone?
2. How do microseismicity and SSEs interact and evolve spatially and temporally?
3. What controls the downdip limit of shallow SSEs?

Here, we summarise the main findings of the thesis, then discuss the extent to which each of the research questions have been addressed and remaining areas for future work.

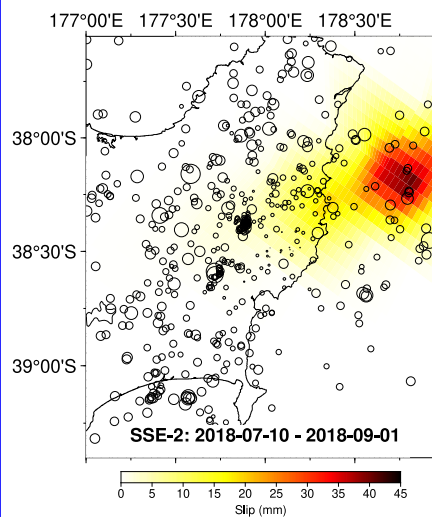
6.1 Summary

We conducted multiple investigations to characterise the slow slip region of the northern Hikurangi margin using a dense, temporary seismograph network. We used these data to constrain the relationship between SSEs and microseismicity, and possible connections between slab processes and surface fluid discharge. Our analysis complements the well-constrained offshore observations of the large 2014 and 2019 SSEs and their associated seismic signatures (e.g., [Todd et al., 2018](#); [Yarce et al., 2019](#); [Woods et al., 2024](#)). We demonstrate that smaller magnitude, understudied, and under-constrained SSEs provide additional opportunities to characterise and understand slip mechanisms in a region currently limited by a relatively short observation record.

We created a new earthquake catalogue, focal mechanism catalogue, and 3-D V_P and V_P/V_S models for the Raukūmara Peninsula. Our analysis focused on earthquakes and two $M_w \sim 6$ SSEs that occurred between December 2017 and October 2018, complementing existing



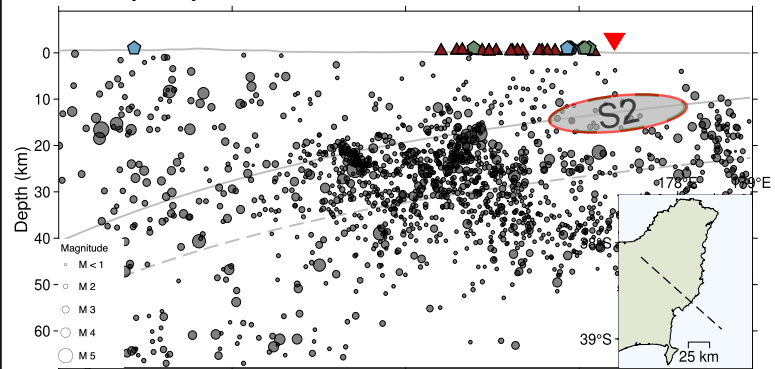
② How do microseismicity & SSEs interact & evolve?



Seismicity rates increase during Mw6 SSEs, similar to observations of larger offshore SSEs. Seismicity clustering peaks halfway through each SSE.

High V_p/V_s to ≈ 35 km depth suggests hydrated subducted oceanic crust and the presence of fluids. Magnetic and bathymetry data, correlated with high V_p anomalies, suggest the presence of seamounts, which provide additional fluids. The downdip limit of SSEs coincide with the cessation of normal faulting and surface observations of seeps.

① What is the seismological character of the deeper part of the shallow SSE zone?



Linear bands of seismicity in the subducting slab crust and mantle and normal-type focal mechanisms consistent with offshore observations.

③ What controls the downdip limit of SSEs?

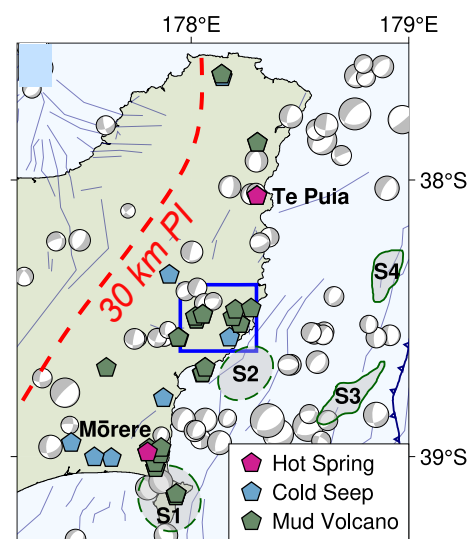
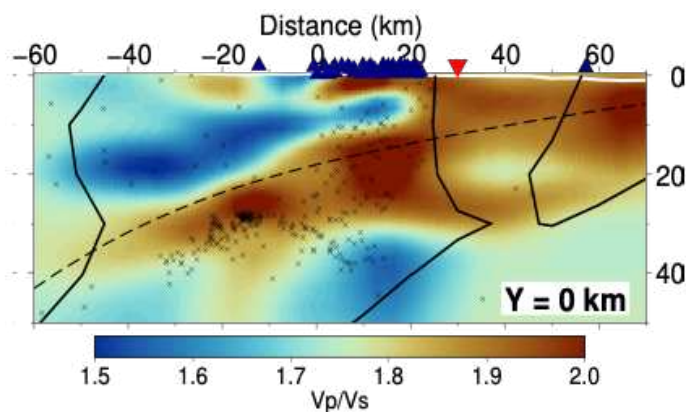


Figure 6.1: Summary figure showing seismological constraints provided in relation to each of the research questions. PI: plate interface.

studies that focused on the $M_w \sim 7$ SSEs in 2014 and 2019. Our earthquake catalogue includes 3,071 earthquakes with hypocentre uncertainties ≤ 5 km, demonstrating active faulting in the upper plate, subducting slab, and along the plate interface. We reduced the magnitude of completeness to 1.7, compared with 2.0 for the GeoNet catalogue. We also constrained focal mechanisms, including numerous normal faults in the subducting slab, consistent with longer-term seismological records from the ISC catalogue and previous studies (e.g., [François-Holden et al., 2008](#); [Lentas et al., 2019](#)).

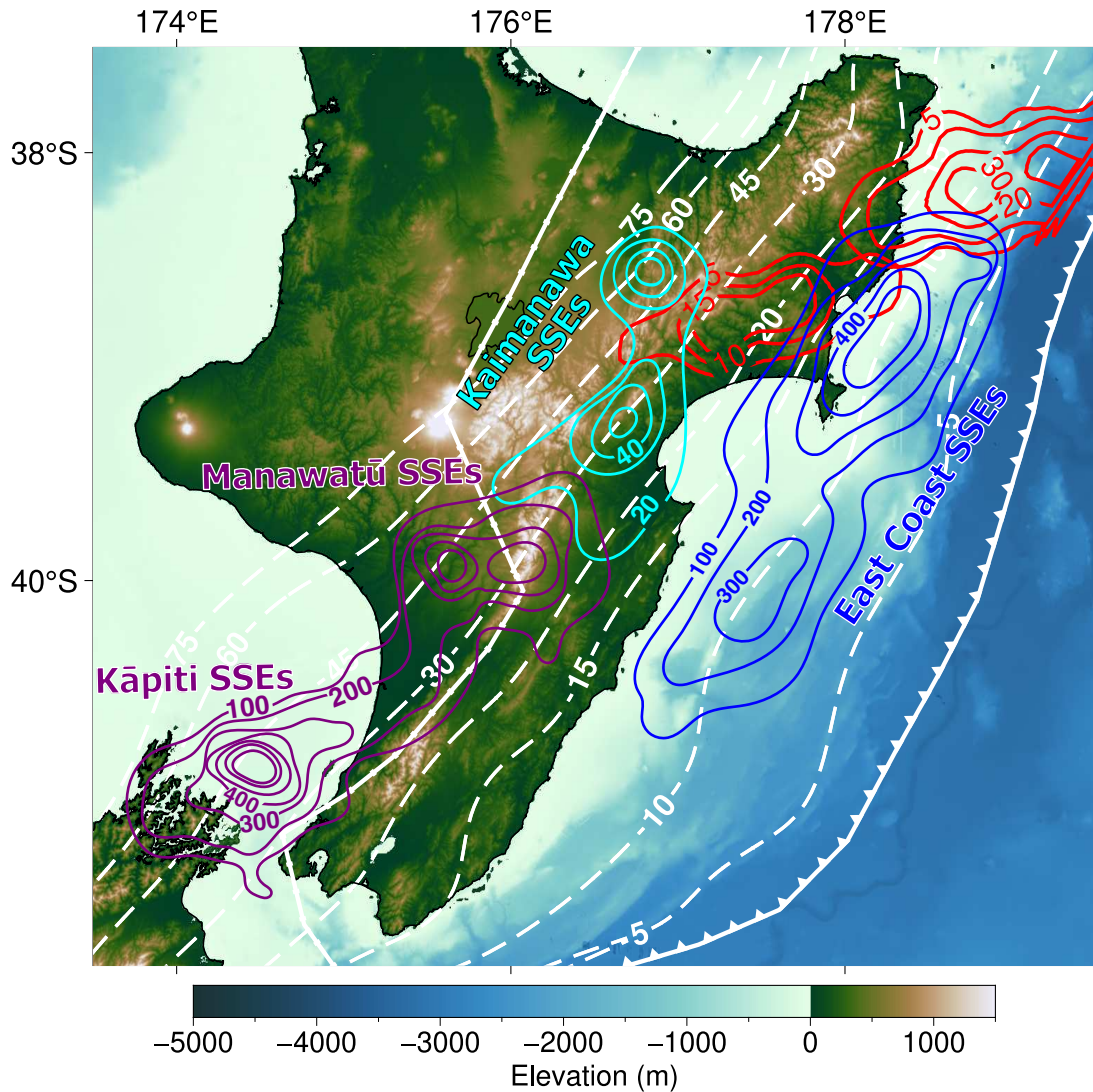


Figure 6.2: Cumulative slow slip contours as in Figure 1.2, with 2018 SSEs (red contours) as discussed in Chapters 4 and 5.

We constrained two previously unreported SSEs using daily displacement data from GNSS stations (Figure 6.2). Analysing our earthquake catalogue in relation to time-dependent slip patches during these SSEs demonstrates that seismicity rates increase and peak during SSEs,

coinciding with the maximum slow slip rate. This shows that earthquake-SSE interactions during smaller magnitude SSEs are broadly consistent with those observed during larger magnitude SSEs, demonstrating the value of constraining smaller events. Given the short observation period and small number of SSEs available for analysis, a reappraisal of the long-term GNSS record for additional small magnitude events would be valuable. A more complete SSE catalogue would allow the statistical significance of earthquake-SSE interactions to be tested more robustly, and would help determine whether the ~20 km-depth seismicity patch is consistently susceptible to SSE-induced earthquake clustering, regardless of SSE location.

Earthquakes and focal mechanisms are observed in the subducting slab crust and mantle. Earthquakes define linear seismicity features up to ≥ 25 km below the plate interface. These features are colocated with observations of onshore surface seeps, including mud volcanoes and hot springs. These hot springs have mantle-derived helium isotope signatures, suggesting that normal faults represent fluid pathways from slab mantle to plate interface and eventually the surface. These surface seeps are clustered above locations of speculated seamounts, based on observations of debris avalanches, positive magnetic anomalies, sediment lenses, and receiver function analysis. We suggest that the subduction of seamounts, which are formed of relatively buoyant, volcanoclastic sediment, enhanced plate roughness and permeability through localised uplift, in addition to existing slab bending processes. Similar observations are seen at the Ryukyu trench in seismic reflection data, and are consistent with seismic reflection data offshore (e.g., [Arai et al., 2017](#)). To further investigate the location of these speculated seamounts, we then used the earthquake catalogue to explore the V_P and V_P/V_S structure.

We combined the NZ3D and offshore HOBITSS earthquake catalogues to derive a high-resolution 3-D V_P and V_P/V_S model. The model was interpretable to ~50 km depth, thereby encompassing the downdip limit of observed SSEs and complementing existing tomographic models focused on the shallower slow slip patch. We observe high V_P anomalies downdip of the locations of speculated seamounts at 15–20 km depths near the Gisborne mud volcanoes and possibly at Te Puia hot springs. We suggest that these represent consolidated sediment

downdip of subducting seamounts, consistent with numerical and experimental modelling, interpretations of seismic reflection data, and magnetic data. These observations suggest that the plate interface remains rough to at least 15 km depth. The Gisborne high V_P anomaly would place the subducted seamount immediately offshore, consistent with normal faulting and linear bands of seismicity observed in our catalogue. Seamount subduction has likely contributed to the destabilisation of the East Coast, and enhanced fluid pathways from slab to the surface, represented by surface observations of seeps with heterogeneous chemical signatures.

We observe high V_P/V_S in the subducting slab, in close agreement with existing plate interface models, which we interpret as a fluid-rich oceanic crust and representative of overpressured fluids at the plate interface. V_P/V_S is highest at ~ 20 km depth, suggesting that this is where fluid concentrations overlap. At ~ 20 km depth, fluids may be sourced from sediment dewatering, clay dehydration, and serpentinisation, accounting for the heterogeneous fluid signatures at the surface. Slab V_P/V_S decreases downdip to ~ 35 km, suggesting a relative reduction in slab-derived fluids coinciding with where SSEs are less frequent and are no longer observed. The high V_P/V_S signature beneath Gisborne suggests that the fluid concentration is highest where the slab is roughest, suggesting that high fluid pressures associated with SSEs and microseismicity are controlled by slab hydration and fracturing, particularly where seamounts are located. Fluid availability is therefore likely a key control on the downdip limit of SSEs.

The downdip limit of SSEs coincides with the cessation of normal faults and surface seeps, and reduced V_P/V_S in the slab. Taken together, our findings suggest that plate roughness plays a key role in controlling fluid distribution, and thus the downdip limit of SSEs. These interpretations are summarised in Figure 6.3. In summary, we suggest that seamount subduction results in sediment consolidation downdip of the seamounts, and concentrated fluids in and around them, increasing fluid pressure. This is represented seismologically by high-angle normal faults, structures linking the slab mantle, plate interface, and upper plate, and high V_P/V_S interpreted as fluid rich material. Previous studies have also identified tremor downdip of seamounts during SSEs (Todd *et al.*, 2018). This manifests as the migration of fluids from the slab mantle to the surface, resulting in surface observations of seeps with varying chemical

signatures along the margin, and destabilisation and eastwards slumping of the East Coast.

6.2 Contrasting the southern Hikurangi margin

The Hikurangi margin displays significant variation in SSE depth (Figure 6.2), magnitude, and recurrence interval, correlated with variations in permeability, plate interface roughness and sediment cover. At the southern Hikurangi margin, SSEs are deeper (15–50 km), likely due to temperature-controlled metamorphic dehydration reactions. The southern Hikurangi margin lacks observable protruding seamounts due to <5 km sediment cover ([Bassett et al., 2025](#)). Lower volumes of fluids reach the surface, and surface seeps have significantly different compositions, notably a lack of mantle-derived isotope signatures ([Reyes et al., 2022](#)). Suggestions of a fluid-poor basement rock are supported by observations of higher V_P , lower V_P/V_S higher resistivity, and higher Q , compared to the northern Hikurangi margin ([Eberhart-Phillips et al., 2008](#); [Eberhart-Phillips and Fry, 2017](#); [Eberhart-Phillips et al., 2020](#); [Heise et al., 2023](#); [Bassett et al., 2023, 2025](#)). The transpressional upper plate has lower permeability than the northern margin, likely resulting in increased plate interface fluid pressures, accounting for the deeper interseismic coupling ([Henry et al., 2025](#)). Bathymetric data show that the outer rise at the southern margin lacks protruding seamounts ([Tozer et al., 2019](#)), suggesting that the plate interface is relatively smooth at depth. Taken together, these properties and features form a subduction environment that is more conducive to deeper SSEs, associated with temperature-controlled dehydration reactions.

6.3 Returning to the original research questions

6.3.1 What is the character of the deeper (>10 km) part of the SSE zone, and how does it compare with the well-constrained shallow zone?

In Chapters 3 and 4, we derived an earthquake and focal mechanism catalogue using the high-density, temporary, onshore seismograph network. Linear bands of seismicity suggest the presence of faults that penetrate the slab crust and mantle, and bound the location of imaged or suggested seamounts Figure 6.3 (label 1). The focal mechanism catalogue reveals normal

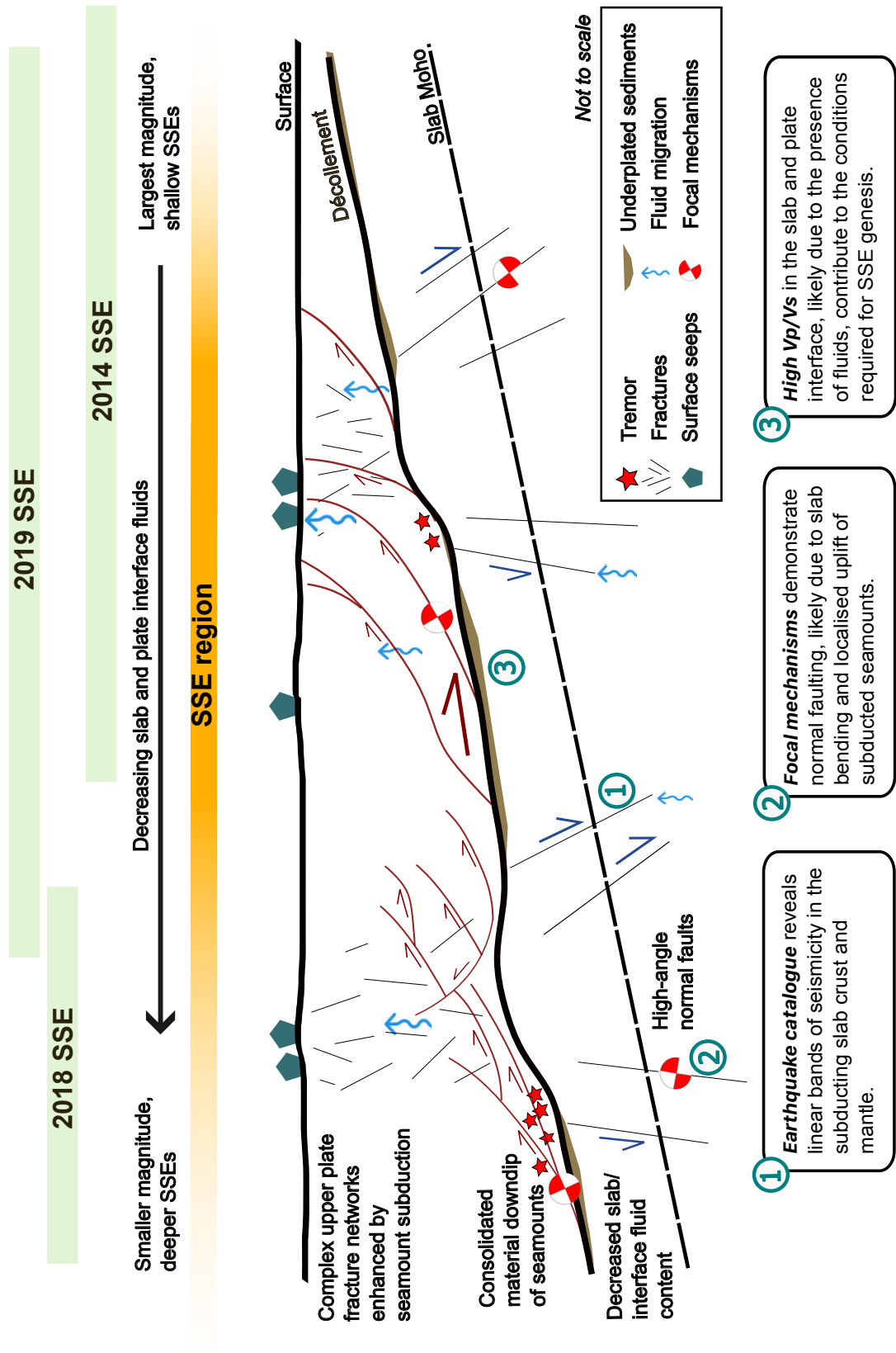


Figure 6.3: Schematic summary of processes discussed in main text.

faulting within the slab lithosphere at depths ≥ 25 km below the plate interface. High-angle normal faults also occur in the region of subducted seamounts, consistent with observations at similar subduction zones (e.g., the Ryukyu trench [Arai et al., 2017](#)). Normal faulting is likely due to a combination of slab bending and localised uplift of relatively buoyant, subducted seamounts. These processes are summarised in Figure 6.3 (label 2). These observations are consistent with offshore seismic reflection and earthquake observations, that show seismicity clusters co-located with seamounts ([Bell et al., 2010](#); [Todd et al., 2018](#)).

In Chapter 5, we used local earthquake tomography to derive an updated high-resolution V_P and V_P/V_S model, centred on Gisborne. We observe a high V_P anomaly near Gisborne, downdip of the suggested location of subducted seamounts, based on high, positive magnetic anomaly, bathymetric data, seismic reflection, and receiver function studies ([Hunt and Glover, 1995](#); [Pedley et al., 2010](#); [Leah et al., 2022](#); [Bangs et al., 2023](#)). This high V_P anomaly aligns well with the linear bands of seismicity and normal faulting, supporting our suggestion that offshore seamount buoyancy-related normal faulting continues to at least 15 km depth. These observations are consistent with offshore seismic reflection and full waveform inversion studies, which interpret high V_P anomalies downdip of seamounts ([Bangs et al., 2023](#)).

Offshore and onshore seeps are clustered above subducted seamounts identified in this and previous studies (Figure 5.2). The onshore seeps, uncontaminated by seawater, allow for constraints on fluid sources at depth. High proportions of mantle-derived fluids at the Te Puia and Mahia Peninsula hot springs support observations of slab mantle normal faults that likely represent fluid pathways from the slab to the surface. Fluid pressures are a key control on SSE occurrence, and these observations suggest that plate interface roughness contributes to fluid availability in the deeper part of the northern Hikurangi slow slip region. This is consistent with offshore data that show lithologic and geometric heterogeneity at cm-to-km scale ([Barnes et al., 2020](#)). The subaerial forearc at the Hikurangi margin allows for the analysis of surface seep composition; at other subduction zones, such as central Ecuador, seeps are at the ocean floor so are contaminated by seawater. It is therefore possible to constrain how fluid sources vary along-strike and with depth at the Hikurangi margin, and relate this to slow slip distribution and occurrence ([Reyes et al., 2022](#)).

6.3.2 How do microseismicity and SSE interact and evolve spatially and temporally?

In Chapter 4, we provided initial constraints on two SSEs during the deployment of the temporary seismograph network. Analysing our earthquake distributions with time-dependent slow slip models demonstrates that seismicity evolves in space and time in relation to SSEs. The first SSE occurred 8–35 km depth, with seismicity clustered updip at ≈ 20 km depth. The second SSE occurred at 5–20 km depth, with seismicity clustered at ≈ 20 km depth.

During both SSEs, seismicity rates peaked mid-way through the SSE, coinciding with the period of maximum slow slip rate. This suggests that SSEs influence regional stress, thus inducing seismicity clusters. For both SSEs, despite the difference in slip depths, seismicity clusters occurred at ~ 20 km depth, suggesting that the spatial clustering of this seismicity is controlled by local conditions and the temporal clustering of seismicity is controlled by SSEs. The region may have lower effective stress, possibly due to higher fluid overpressures, subducted fluid-rich topography, underthrust and dehydrated subducted sediments, dehydrated slab minerals, or a combination of these processes.

Seismicity clusters occur in a region of high V_P/V_S (as shown in Chapter 5), supporting the idea that microseismicity clusters are associated with higher fluid concentrations, similar to [Mochizuki et al. \(2021\)](#). This is also supported by independent constraints that find that microseismicity and tremor clusters during SSEs are located at or near subducted seamounts, rather than at consistent locations relative to the slow slip patch, demonstrating that plate interface geometry and lithology control seismicity distributions ([Todd and Schwartz, 2016](#); [Todd et al., 2018](#); [Shaddox and Schwartz, 2019](#)).

Our observations characterise the deeper part of the northern Hikurangi margin slow slip region, demonstrating that offshore and onshore observations of plate interface roughness and high fluid concentrations are relatively continuous, and coincide with the downdip limit of SSEs. This suggests that shallow SSEs are controlled by fluid availability and plate roughness.

6.3.3 What controls the downdip limit of shallow SSEs?

In Chapter 5, we used local earthquake tomography to investigate the wavespeed structure of the slow slip region. High V_P anomalies are interpreted as consolidated material downdip of subducted seamounts up to 20 km depth. This suggests that the plate interface remains rough and fluid-rich to at least these depths due to the subduction of fluid-rich topography, with the buoyancy and uplift effects of seamounts further enhancing roughness and providing upward fluid pathways.

High V_P/V_S (>1.8) is observed in the slab, highest at 20 km depth and decreasing downdip. Independent offshore models also show high V_P/V_S <20 km depth (Mochizuki *et al.*, 2021; Yarce *et al.*, 2021). The high V_P/V_S regions are correlated with observations of slab normal faulting and surface seeps, shown by our focal mechanisms, seismic reflection data (Figure 4.2), and previous studies (Reyes *et al.*, 2022; Bangs *et al.*, 2023). The downdip limit of SSEs coincides with the cessation of observations of normal faults and surface seeps, suggesting that fluid availability may become insufficient for SSE genesis.

Fluids are likely also sourced from clay dehydration and serpentinisation reactions, which initiate from 10 and 13–21 km depth respectively. The presence of fluids is supported by the high V_P/V_S observed in the subducting slab, which closely aligns with existing plate interface models (Williams *et al.*, 2013). At greater depths, fluid availability may therefore become insufficient for shallow SSE genesis.

6.4 Future work

6.4.1 Observations of an unusually deep SSE at the northern Hikurangi margin

The first SSE in 2018 was unusually deep ($>\sim 20$ km), and was an order of magnitude smaller than most of the SSEs constrained and studied at this margin. It remains unknown whether this was an anomalous event, or whether it suggests that there may be further undetected SSEs at these depths. We note also that this SSE occurred further west of typical East Coast

SSEs, overlapping with the Kaimanawa SSE region, suggesting that it may be associated with transitional properties between the East Coast and Kaimanawa type SSEs or that the two regions may interact.

Unlike fast earthquakes, SSEs often recur with sufficient regularity to allow targeted observations. Deep, southern Hikurangi SSEs occur every 4–5 years while shallow, northern Hikurangi SSEs occur every 1–2 years ([Wallace, 2020](#)). A reappraisal of the long-term GNSS record to determine if there have been additional >20 km depth northern Hikurangi SSEs would determine the recurrence interval of such events. At distances >100 km landward of the trench, reduced upper plate fluid content and fault interconnectivity result in decreased V_P/V_S as seen in our models and [Eberhart-Phillips and Bannister \(2015\)](#). If SSE occurrence is driven by a fault-valving process (e.g., [Warren-Smith et al., 2019](#)), we might expect that because this region has less fluid than the shallower portion of the margin, fluids take longer to accumulate and reach overpressured conditions, resulting in longer recurrence intervals.

6.4.2 Plate interface heterogeneity at different spatial scales

We have explored the extent to which plate interface heterogeneity controls conditions that result in SSE genesis at km–tens of km scales that can be constrained by seismological methods. Offshore data has shown that plate interface heterogeneity occurs on scales of cm to tens of km ([Barnes et al., 2020](#)). Additional constraints on onshore plate interface heterogeneity would be useful to explore the extent to which smaller-scale plate interface heterogeneity continues at depth, and the relative importance of large-scale versus small-scale features. It is likely that small-scale heterogeneity persists to similar depths to large-scale, because it seems unlikely that slow slip conditions are enabled only by large, discrete heterogeneities separated by tens of km. In particular, the Gisborne network could be used for land-based full waveform inversion to determine the structure of the upper plate near the mud volcanoes, and possibly how these connect to the subducted seamount near Gisborne.

The V_P and V_P/V_S models derived in Chapter 5 made use of multiple local earthquake catalogues. The inclusion of controlled-source seismic data from the NZ3D experiment could help improve resolution of the shallowest structure beneath the Waimata Valley mud volcanoes,

the potential relationships between plate interface structure and upper plate conduits (e.g., [Arai et al., 2023](#)), and the seamount that we and others have suggested is located directly beneath the Gisborne network. Additionally, if we were to include teleseismic phase arrival times, we could improve resolution in the outer parts of the model or potentially increase coverage of the model to greater depths.

6.4.3 Surface seep compositions

The abundance of surface mud volcanoes and hot springs in regions of suspected subducted seamounts is consistent with observations at other subduction zones (e.g., [Arai et al., 2017](#)). The Hikurangi margin is unique, however, because the subaerial forearc allows for sampling of surface seeps that are not contaminated by seawater. At other subduction zones that host shallow SSEs, though there are many offshore seeps that support the idea of high fluid flow through the forearc, direct sampling is not possible.

Additionally, why the composition of seeps at the Hikurangi margin varies significantly over a ~150 km lateral extent along-strike remains uncertain. The Te Puia and Mahia hot springs are rich in mantle-derived fluids, while the Waimata Valley mud volcanoes are dominated by muddy clays. While the higher proportions of mantle-derived fluids at the northern Hikurangi margin compared with the southern Hikurangi margin seaward of the Australian Plate mantle wedge demonstrates that surface seeps are linked to processes at significant depths ([Reyes et al., 2022](#)), the reason for significant heterogeneity within the northern Hikurangi margin region remains unclear. Use of the 143 short-period DATA-CUBE3 and GSX3 instruments centred on the Waimata Valley mud volcanoes could provide additional constraints on potential upper plate conduits and the source of the fluid heterogeneity. Initial ambient noise tomography results suggest the presence of ~5 km low V_S anomalies beneath the mud volcanoes ([Goerzen, 2021](#)). Joint consideration of subsurface velocity structure and mud volcanism activity would be helpful to constrain fluid pathways (e.g., [Ferrari et al., 2026](#)).

Bibliography

- Aden-Antoniów, F., W. B. Frank, C. J. Chamberlain, J. Townend, L. M. Wallace, and S. Bannister (2024), Low-Frequency Earthquakes Downdip of Deep Slow Slip Beneath the North Island of New Zealand, *Journal of Geophysical Research: Solid Earth*, 129(5), e2023JB027971, doi:[10.1029/2023JB027971](https://doi.org/10.1029/2023JB027971), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2023JB027971>.
- Aki, K. (1965), Maximum likelihood estimate of b in the formula $\log N = a - bM$ and its confidence limits, *Bulletin of the Earthquake Research Institute*, 43, 237–239.
- Anderson, H., E. Smith, and R. Robinson (1990), Normal faulting in a back arc basin: Seismological characteristics of the March 2, 1987, Edgcumbe, New Zealand, Earthquake, *Journal of Geophysical Research: Solid Earth*, 95(B4), 4709–4723, doi:[10.1029/JB095iB04p04709](https://doi.org/10.1029/JB095iB04p04709), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/JB095iB04p04709>.
- Ando, R., N. Takeda, and T. Yamashita (2012), Propagation dynamics of seismic and aseismic slip governed by fault heterogeneity and Newtonian rheology, *Journal of Geophysical Research: Solid Earth*, 117(B11), 2012JB009532, doi:[10.1029/2012JB009532](https://doi.org/10.1029/2012JB009532).
- Arai, R., S. Kodaira, T. Yamada, T. Takahashi, S. Miura, Y. Kaneda, A. Nishizawa, and M. Oikawa (2017), Subduction of thick oceanic plateau and high-angle normal-fault earthquakes intersecting the slab, *Geophysical Research Letters*, 44(12), 6109–6115, doi:[10.1002/2017GL073789](https://doi.org/10.1002/2017GL073789), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2017GL073789>.
- Arai, R., S. Kodaira, S. Henrys, N. Bangs, K. Obana, G. Fujie, S. Miura, D. Barker, D. Bassett, R. Bell, K. Mochizuki, R. Kellett, V. Stucker, B. Fry, and N. Team (2020), Three-Dimensional P Wave Velocity Structure of the Northern Hikurangi Margin From the

- NZ3D Experiment: Evidence for Fault-Bound Anisotropy, *Journal of Geophysical Research: Solid Earth*, 125(12), e2020JB020433, doi:[10.1029/2020JB020433](https://doi.org/10.1029/2020JB020433), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JB020433>.
- Arai, R., S. Miura, Y. Nakamura, G. Fujie, S. Kodaira, Y. Kaiho, K. Mochizuki, R. Nakata, M. Kinoshita, Y. Hashimoto, Y. Hamada, and K. Okino (2023), Upper-plate conduits linked to plate boundary that hosts slow earthquakes, *Nature Communications*, 14(1), 5101, doi:[10.1038/s41467-023-40762-4](https://doi.org/10.1038/s41467-023-40762-4), number: 1.
- Araki, E., D. M. Saffer, A. J. Kopf, L. M. Wallace, T. Kimura, Y. Machida, S. Ide, E. Davis, IODP Expedition 365 shipboard scientists, S. Toczko, S. Carr, C. Kinoshita, R. Kobayashi, and A. Rösner (2017), Recurring and triggered slow-slip events near the trench at the Nankai Trough subduction megathrust, *Science*, 356(6343), 1157–1160, doi:[10.1126/science.aan3120](https://doi.org/10.1126/science.aan3120).
- Arnulf, A. F., J. Biemiller, L. Lavier, L. M. Wallace, D. Bassett, S. Henrys, I. Pecher, G. Crutchley, and A. Plaza Faverola (2021), Physical conditions and frictional properties in the source region of a slow-slip event, *Nature Geoscience*, 14(5), 334–340, doi:[10.1038/s41561-021-00741-0](https://doi.org/10.1038/s41561-021-00741-0).
- Audet, P., and Y. Kim (2016), Teleseismic constraints on the geological environment of deep episodic slow earthquakes in subduction zone forearcs: A review, *Tectonophysics*, 670, 1–15, doi:[10.1016/j.tecto.2016.01.005](https://doi.org/10.1016/j.tecto.2016.01.005).
- Audet, P., M. G. Bostock, N. I. Christensen, and S. M. Peacock (2009), Seismic evidence for overpressured subducted oceanic crust and megathrust fault sealing, *Nature*, 457(7225), 76–78, doi:[10.1038/nature07650](https://doi.org/10.1038/nature07650).
- Ballance, P. F. (1976), Evolution of the Upper Cenozoic magmatic arc and plate boundary in northern New Zealand, *Earth and Planetary Science Letters*, 28, 356–370.
- Bangs, N. L., J. K. Morgan, R. E. Bell, S. Han, R. Arai, S. Kodaira, A. C. Gase, X. Wu, R. Davy, L. Frahm, H. L. Tilley, D. H. N. Barker, J. H. Edwards, H. J. Tobin, T. J. Reston, S. A. Henrys, G. F. Moore, D. Bassett, R. Kellett, V. Stucker, and B. Fry (2023), Slow slip along

the Hikurangi margin linked to fluid-rich sediments trailing subducting seamounts, *Nature Geoscience*, 16(6), 505–512, doi:[10.1038/s41561-023-01186-3](https://doi.org/10.1038/s41561-023-01186-3).

Bannister, S. C., B. J. Perin, and T. H. Webb (1989), Normal faulting through subducted oceanic crust: the 19 July 1985 earthquake of Hawke's Bay, New Zealand, *Tectonophysics*, 162(3-4), 303–313, doi:[10.1016/0040-1951\(89\)90251-5](https://doi.org/10.1016/0040-1951(89)90251-5).

Barker, D. H. N., R. Sutherland, S. Henrys, and S. Bannister (2009), Geometry of the Hikurangi subduction thrust and upper plate, North Island, New Zealand, *Geochemistry, Geophysics, Geosystems*, 10(2), doi:[10.1029/2008GC002153](https://doi.org/10.1029/2008GC002153), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2008GC002153>.

Barker, D. H. N., S. Henrys, F. Caratori Tontini, P. M. Barnes, D. Bassett, E. Todd, and L. Wallace (2018), Geophysical Constraints on the Relationship Between Seamount Subduction, Slow Slip, and Tremor at the North Hikurangi Subduction Zone, New Zealand, *Geophysical Research Letters*, 45(23), 12,804–12,813, doi:[10.1029/2018GL080259](https://doi.org/10.1029/2018GL080259), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2018GL080259>.

Barnes, J. D., J. Cullen, S. Barker, S. Agostini, S. Penniston-Dorland, J. C. Lassiter, A. Klügel, and L. Wallace (2019), The role of the upper plate in controlling fluid-mobile element (Cl, Li, B) cycling through subduction zones: Hikurangi forearc, New Zealand, *Geosphere*, 15(3), 642–658, doi:[10.1130/GES02057.1](https://doi.org/10.1130/GES02057.1).

Barnes, P. M., and B. M. De Lépinay (1997), Rates and mechanics of rapid frontal accretion along the very obliquely convergent southern Hikurangi margin, New Zealand, *Journal of Geophysical Research: Solid Earth*, 102(B11), 24,931–24,952, doi:[10.1029/97JB01384](https://doi.org/10.1029/97JB01384).

Barnes, P. M., G. Lamarche, J. Bialas, S. Henrys, I. Pecher, G. L. Netzeband, J. Greinert, J. J. Mountjoy, K. Pedley, and G. Crutchley (2010), Tectonic and geological framework for gas hydrates and cold seeps on the Hikurangi subduction margin, New Zealand, *Marine Geology*, 272(1-4), 26–48, doi:[10.1016/j.margeo.2009.03.012](https://doi.org/10.1016/j.margeo.2009.03.012).

Barnes, P. M., L. M. Wallace, D. M. Saffer, R. E. Bell, M. B. Underwood, A. Fagereng, F. Meneghini, H. M. Savage, H. S. Rabinowitz, J. K. Morgan, H. Kitajima, S. Kutterolf, Y. Hashimoto,

- C. H. Engelmann de Oliveira, A. Noda, M. P. Crundwell, C. L. Shepherd, A. D. Woodhouse, R. N. Harris, M. Wang, S. Henrys, D. H. Barker, K. E. Petronotis, S. M. Bourlange, M. B. Clennell, A. E. Cook, B. E. Dugan, J. Elger, P. M. Fulton, D. Gamboa, A. Greve, S. Han, A. Hüpers, M. J. Ikari, Y. Ito, G. Y. Kim, H. Koge, H. Lee, X. Li, M. Luo, P. R. Malie, G. F. Moore, J. J. Mountjoy, D. D. McNamara, M. Paganoni, E. J. Screaton, U. Shankar, S. Shreedharan, E. A. Solomon, X. Wang, H.-Y. Wu, I. A. Pecher, L. J. LeVay, and IODP Expedition 372 Scientists (2020), Slow slip source characterized by lithological and geometric heterogeneity, *Science Advances*, 6(13), eaay3314, doi:[10.1126/sciadv.aay3314](https://doi.org/10.1126/sciadv.aay3314).
- Bartlow, N. M., S. Miyazaki, A. M. Bradley, and P. Segall (2011), Space-time correlation of slip and tremor during the 2009 Cascadia slow slip event, *Geophysical Research Letters*, 38(18), doi:[10.1029/2011GL048714](https://doi.org/10.1029/2011GL048714), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2011GL048714>.
- Bartlow, N. M., L. M. Wallace, R. J. Beavan, S. Bannister, and P. Segall (2014), Time-dependent modeling of slow slip events and associated seismicity and tremor at the Hikurangi subduction zone, New Zealand, *Journal of Geophysical Research: Solid Earth*, 119(1), 734–753, doi:[10.1002/2013JB010609](https://doi.org/10.1002/2013JB010609), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2013JB010609>.
- Bassett, D., and A. B. Watts (2015), Gravity anomalies, crustal structure, and seismicity at subduction zones: 1. Seafloor roughness and subducting relief, *Geochemistry, Geophysics, Geosystems*, 16(5), 1508–1540, doi:[10.1002/2014GC005684](https://doi.org/10.1002/2014GC005684).
- Bassett, D., R. Sutherland, and S. Henrys (2014), Slow wavespeeds and fluid overpressure in a region of shallow geodetic locking and slow slip, Hikurangi subduction margin, New Zealand, *Earth and Planetary Science Letters*, 389, 1–13, doi:[10.1016/j.epsl.2013.12.021](https://doi.org/10.1016/j.epsl.2013.12.021).
- Bassett, D., A. Arnulf, S. Kodaira, A. Nakanishi, A. Harding, and G. Moore (2022), Crustal Structure of the Nankai Subduction Zone Revealed by Two Decades of Onshore-Offshore and Ocean-Bottom Seismic Data: Implications for the Dimensions and Slip Behavior of the Seismogenic Zone, *Journal of Geophysical Re-*

search: *Solid Earth*, 127(10), e2022JB024,992, doi:[10.1029/2022JB024992](https://doi.org/10.1029/2022JB024992), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2022JB024992>.

Bassett, D., G. Fujie, S. Kodaira, R. Arai, Y. Yamamoto, S. Henrys, D. Barker, A. Gase, H. van Avendonk, N. Bangs, H. Seebeck, B. Tozer, K. Jacobs, T. Luckie, D. Okaya, and K. Mochizuki (2023), Heterogeneous Crustal Structure of the Hikurangi Plateau Revealed by SHIRE Seismic Data: Origin and Implications for Plate Boundary Tectonics, *Geophysical Research Letters*, 50(22), e2023GL105,674, doi:[10.1029/2023GL105674](https://doi.org/10.1029/2023GL105674), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2023GL105674>.

Bassett, D., S. Henrys, B. Tozer, H. van Avendonk, A. Gase, N. Bangs, S. Kodaira, D. Okaya, K. Jacobs, R. Sutherland, H. Seebeck, D. Barker, G. Fujie, R. Arai, A. Seaward, K. Mochizuki, M. Savage, T. Stern, and T. Luckie (2025), Crustal Structure of the Hikurangi Subduction Zone Revealed by Four Decades of Onshore-Offshore Seismic Data: Implications for the Dimensions and Slip Behavior of the Seismogenic Zone, *Journal of Geophysical Research: Solid Earth*, 130(1), e2024JB030,268, doi:[10.1029/2024JB030268](https://doi.org/10.1029/2024JB030268), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2024JB030268>.

Behboudi, E., D. D. McNamara, and I. Lokmer (2023), Shallow Tectonic Stress Magnitudes at the Hikurangi Subduction Margin, New Zealand, *Geochemistry, Geophysics, Geosystems*, 24(10), e2022GC010,836, doi:[10.1029/2022GC010836](https://doi.org/10.1029/2022GC010836), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2022GC010836>.

Bell, R., R. Sutherland, D. H. N. Barker, S. Henrys, S. Bannister, L. Wallace, and J. Beavan (2010), Seismic reflection character of the Hikurangi subduction interface, New Zealand, in the region of repeated Gisborne slow slip events, *Geophysical Journal International*, 180(1), 34–48, doi:[10.1111/j.1365-246X.2009.04401.x](https://doi.org/10.1111/j.1365-246X.2009.04401.x), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1111/j.1365-246X.2009.04401.x>.

Bell, R., C. Holden, W. Power, X. Wang, and G. Downes (2014), Hikurangi margin tsunami earthquake generated by slow seismic rupture over a subducted seamount, *Earth and Planetary Science Letters*, 397, 1–9, doi:[10.1016/j.epsl.2014.04.005](https://doi.org/10.1016/j.epsl.2014.04.005).

- Bell, R., J. Morgan, M. Warner, A. Fagereng, and L. McNeill (2017), NZ3D-FWI, doi:[10.7914/SN/3C_2017](https://doi.org/10.7914/SN/3C_2017).
- Bell, R., M. Gray, J. Morgan, M. Warner, A. Fagereng, L. McNeill, K. Jacobs, S. A. Henrys, B. Fry, S. Watkins, H. Lacey, J. Black, V. Lane, D. Daly, D. Lindsay, N. Bangs, R. Arai, and S. Kodaira (2019), New Zealand 3D Full Waveform Inversion (NZ3D-FWI) 2017-2018 field acquisition report, *GNS Science*, doi:[10.21420/ZZ8R-QR04](https://doi.org/10.21420/ZZ8R-QR04).
- Bilek, S. L., and T. Lay (2002), Tsunami earthquakes possibly widespread manifestations of frictional conditional stability, *Geophysical Research Letters*, 29(14), 18–1–18–4, doi:[10.1029/2002GL015215](https://doi.org/10.1029/2002GL015215), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2002GL015215>.
- Bird, P. (2003), An updated digital model of plate boundaries, *Geochemistry, Geophysics, Geosystems*, 4(3), doi:[10.1029/2001GC000252](https://doi.org/10.1029/2001GC000252), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2001GC000252>.
- Bormann, P., B. Engdahl, and R. Kind (2012), *New manual of seismological observatory practice 2: Seismic wave propagation and Earth models*, Deutsches GeoForschungZentrum GFZ.
- Brantut, N., J. Sulem, and A. Schubnel (2011), Effect of dehydration reactions on earthquake nucleation: Stable sliding, slow transients, and unstable slip, *Journal of Geophysical Research*, 116(B5), B05,304, doi:[10.1029/2010JB007876](https://doi.org/10.1029/2010JB007876).
- Brune, J. N. (1970), Tectonic stress and the spectra of seismic shear waves from earthquakes, *Journal of Geophysical Research* (1896-1977), 75(26), 4997–5009, doi:[10.1029/JB075i026p04997](https://doi.org/10.1029/JB075i026p04997), eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/JB075i026p04997>.
- Böttner, C., F. Gross, J. Geersen, G. J. Crutchley, J. J. Mountjoy, and S. Krastel (2018), Marine Forearc Extension in the Hikurangi Margin: New Insights From High-Resolution 3-D Seismic Data, *Tectonics*, 37(5), 1472–1491, doi:[10.1029/2017TC004906](https://doi.org/10.1029/2017TC004906).
- Bürgmann, R. (2018), The geophysics, geology and mechanics of slow fault slip, *Earth and Planetary Science Letters*, 495, 112–134, doi:[10.1016/j.epsl.2018.04.062](https://doi.org/10.1016/j.epsl.2018.04.062).

- Carlson, M., L. Wallace, D. Saffer, and C. Williams (2026), Slow Slip Accommodates the Full Plate Convergence Budget at the Northern Hikurangi Subduction Zone, doi:[10.5194/egusphere-egu26-8339](https://doi.org/10.5194/egusphere-egu26-8339).
- Chalumeau, C., H. Agurto-Detzel, A. Rietbrock, M. Frietsch, O. Oncken, M. Segovia, and A. Galve (2024), Seismological evidence for a multifault network at the subduction interface, *Nature*, 628(8008), 558–562, doi:[10.1038/s41586-024-07245-y](https://doi.org/10.1038/s41586-024-07245-y).
- Chamberlain, C. J., C. J. Hopp, C. M. Boese, E. Warren-Smith, D. Chambers, S. X. Chu, K. Michailos, and J. Townend (2018), EQcorrscan: Repeating and Near-Repeating Earthquake Detection and Analysis in Python, *Seismological Research Letters*, 89(1), 173–181, doi:[10.1785/0220170151](https://doi.org/10.1785/0220170151).
- Chen, S. K., Y.-M. Wu, and Y.-C. Chan (2018), Episodic Slow Slip Events and Overlying Plate Seismicity at the Southernmost Ryukyu Trench, *Geophysical Research Letters*, 45(19), 10,369–10,377, doi:[10.1029/2018GL079740](https://doi.org/10.1029/2018GL079740), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2018GL079740>.
- Chesley, C., S. Naif, K. Key, and D. Bassett (2021), Fluid-rich subducting topography generates anomalous forearc porosity, *Nature*, 595(7866), 255–260, doi:[10.1038/s41586-021-03619-8](https://doi.org/10.1038/s41586-021-03619-8).
- Chesley, C., S. Naif, and K. Key (2023), Characterizing the porosity structure and gas hydrate distribution at the southern Hikurangi Margin, New Zealand from offshore electromagnetic data, *Geophysical Journal International*, 234(3), 2411–2428, doi:[10.1093/gji/ggad243](https://doi.org/10.1093/gji/ggad243).
- Chow, B., Y. Kaneko, C. Tape, R. Modrak, and J. Townend (2020), An automated workflow for adjoint tomography—waveform misfits and synthetic inversions for the North Island, New Zealand, *Geophysical Journal International*, 223(3), 1461–1480, doi:[10.1093/gji/ggaa381](https://doi.org/10.1093/gji/ggaa381).
- Chow, B., Y. Kaneko, and J. Townend (2022a), Evidence for Deeply Subducted Lower-Plate Seamounts at the Hikurangi Subduction Margin: Implications for Seismic and Aseismic Behavior, *Journal of Geophysical Research*:

- Solid Earth*, 127(1), e2021JB022866, doi:[10.1029/2021JB022866](https://doi.org/10.1029/2021JB022866), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021JB022866>.
- Chow, B., Y. Kaneko, C. Tape, R. Modrak, N. Mortimer, S. Bannister, and J. Townend (2022b), Strong Upper-Plate Heterogeneity at the Hikurangi Subduction Margin (North Island, New Zealand) Imaged by Adjoint Tomography, *Journal of Geophysical Research: Solid Earth*, 127(1), e2021JB022865, doi:[10.1029/2021JB022865](https://doi.org/10.1029/2021JB022865), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021JB022865>.
- Coffin, M. F., and O. Eldholm (1994), Large igneous provinces: Crustal structure, dimensions, and external consequences, *Reviews of Geophysics*, 32(1), 1–36, doi:[10.1029/93RG02508](https://doi.org/10.1029/93RG02508).
- Collot, J.-Y., J. Delteil, K. B. Lewis, B. Davy, G. Lamarche, J.-C. Audru, P. Barnes, F. Chancier, E. Chaumillon, S. Lallemand, B. M. De Lepinay, A. Orpin, B. Pelletier, M. Sosson, B. Toussaint, and C. Uruski (1996), From oblique subduction to intra-continental transpression: Structures of the southern Kermadec-Hikurangi margin from multibeam bathymetry, side-scan sonar and seismic reflection, *Marine Geophysical Researches*, 18(2-4), 357–381, doi:[10.1007/BF00286085](https://doi.org/10.1007/BF00286085).
- Collot, J.-Y., K. Lewis, G. Lamarche, and S. Lallemand (2001), The giant Rutoria debris avalanche on the northern Hikurangi margin, New Zealand: Result of oblique seamount subduction, *Journal of Geophysical Research: Solid Earth*, 106(B9), 19,271–19,297, doi:[10.1029/2001JB900004](https://doi.org/10.1029/2001JB900004), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2001JB900004>.
- Crotwell, H. P., T. J. Owens, and J. Ritsema (1999), The TauP Toolkit: Flexible Seismic Travel-time and Ray-path Utilities, *Seismological Research Letters*, 70(2), 154–160, doi:[10.1785/gssrl.70.2.154](https://doi.org/10.1785/gssrl.70.2.154).
- Crutchley, G. J., I. A. Pecher, A. R. Gorman, S. A. Henrys, and J. Greinert (2010), Seismic imaging of gas conduits beneath seafloor seep sites in a shallow marine gas hydrate province, Hikurangi Margin, New Zealand, *Marine Geology*, 272(1-4), 114–126, doi:[10.1016/j.margeo.2009.03.007](https://doi.org/10.1016/j.margeo.2009.03.007).

- Darby, D., R. Funnell, C. I. Uruski, and B. D. Field (2000), Patterns of fluid flow in the East Coast Basin, *New Zealand Petroleum Conference Proceedings*.
- Dascher-Cousineau, K., and R. Bürgmann (2024), Global subduction slow slip events and associated earthquakes, *Science Advances*, 10(35), doi:[10.1126/sciadv.ado2191](https://doi.org/10.1126/sciadv.ado2191).
- Davy, B., K. Hoernle, and R. Werner (2008), Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history, *Geochemistry, Geophysics, Geosystems*, 9(7), doi:[10.1029/2007GC001855](https://doi.org/10.1029/2007GC001855), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2007GC001855>.
- Davy, R. G., L. Frahm, R. Bell, R. Arai, D. H. N. Barker, S. Henrys, N. Bangs, J. Morgan, and M. Warner (2021), Generating High-Fidelity Reflection Images Directly From Full-Waveform Inversion: Hikurangi Subduction Zone Case Study, *Geophysical Research Letters*, 48(19), e2021GL094981, doi:[10.1029/2021GL094981](https://doi.org/10.1029/2021GL094981).
- Delahaye, E. J., J. Townend, M. E. Reyners, and G. Rogers (2009), Microseismicity but no tremor accompanying slow slip in the Hikurangi subduction zone, New Zealand, *Earth and Planetary Science Letters*, 277(1), 21–28, doi:[10.1016/j.epsl.2008.09.038](https://doi.org/10.1016/j.epsl.2008.09.038).
- Diehl, T., and E. Kissling (2007), Users Guide for Consistent Phase Picking at Local to Regional Scales, *ETH Zurich*.
- Dixon, T. H., Y. Jiang, R. Malservisi, R. McCaffrey, N. Voss, M. Protti, and V. Gonzalez (2014), Earthquake and tsunami forecasts: Relation of slow slip events to subsequent earthquake rupture, *Proceedings of the National Academy of Sciences*, 111(48), 17,039–17,044, doi:[10.1073/pnas.1412299111](https://doi.org/10.1073/pnas.1412299111).
- Doser, D. I., and T. H. Webb (2003), Source parameters of large historical (1917-1961) earthquakes, North Island, New Zealand, *Geophysical Journal International*, 152(3), 795–832, doi:[10.1046/j.1365-246X.2003.01895.x](https://doi.org/10.1046/j.1365-246X.2003.01895.x).
- Douglas, A., J. Beavan, L. Wallace, and J. Townend (2005), Slow slip on the northern Hikurangi subduction interface, New Zealand, *Geo-*

physical Research Letters, 32(16), doi:[10.1029/2005GL023607](https://onlinelibrary.wiley.com/doi/pdf/10.1029/2005GL023607), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2005GL023607>.

Dragert, H., K. Wang, and T. S. James (2001), A Silent Slip Event on the Deeper Cascadia Subduction Interface, *Science*, 292(5521), 1525–1528, doi:[10.1126/science.1060152](https://doi.org/10.1126/science.1060152).

Eberhart-Phillips, D. (1986), Three-dimensional velocity structure in northern California coast ranges from inversion of local earthquake arrival times, *Bulletin of the Seismological Society of America*, 76(4), 1025–1052.

Eberhart-Phillips, D., and S. Bannister (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, doi:[10.1093/gji/ggv057](https://doi.org/10.1093/gji/ggv057).

Eberhart-Phillips, D., and B. Fry (2017), A new scheme for joint surface wave and earthquake travel-time inversion and resulting 3-D velocity model for the western North Island, New Zealand, *Physics of the Earth and Planetary Interiors*, 269, 98–111, doi:[10.1016/j.pepi.2017.05.014](https://doi.org/10.1016/j.pepi.2017.05.014).

Eberhart-Phillips, D., M. Chadwick, and S. Bannister (2008), Three-dimensional attenuation structure of central and southern South Island, New Zealand, from local earthquakes, *Journal of Geophysical Research: Solid Earth*, 113(B5), doi:[10.1029/2007JB005359](https://doi.org/10.1029/2007JB005359), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2007JB005359>.

Eberhart-Phillips, D., S. Bannister, and M. Reyners (2017), Deciphering the 3-D distribution of fluid along the shallow Hikurangi subduction zone using P- and S-wave attenuation, *Geophysical Journal International*, 211(2), 1032–1045, doi:[10.1093/gji/ggx348](https://doi.org/10.1093/gji/ggx348).

Eberhart-Phillips, D., S. Bannister, M. Reyners, and S. Henrys (2020), New Zealand Wide model 2.2 seismic velocity and Qs and Qp models for New Zealand, doi:[10.5281/zenodo.3779523](https://doi.org/10.5281/zenodo.3779523).

Eberhart-Phillips, D., C. H. Thurber, A. Rietbrock, B. Fry, M. Reyners, and F. Lanza (2024), Simul2023: a flexible program for inversion of earthquake data for 3-D velocity and hypocenters or 3-D Q.

- Ekström, G., M. Nettles, and A. Dziewoński (2012), The global CMT project 2004–2010: Centroid-moment tensors for 13,017 earthquakes, *Physics of the Earth and Planetary Interiors*, 200–201, 1–9, doi:[10.1016/j.pepi.2012.04.002](https://doi.org/10.1016/j.pepi.2012.04.002).
- Ellis, S., A. Fagereng, D. Barker, S. Henrys, D. Saffer, L. Wallace, C. Williams, and R. Harris (2015), Fluid budgets along the northern Hikurangi subduction margin, New Zealand: the effect of a subducting seamount on fluid pressure, *Geophysical Journal International*, 202(1), 277–297, doi:[10.1093/gji/ggv127](https://doi.org/10.1093/gji/ggv127).
- Ester, M., H.-P. Kriegel, and X. Xu (1996), A Density-Based Algorithm for Discovering Clusters in Large Spatial Databases with Noise, *KDD-96: The Second International Conference on Knowledge Discovery and Data Mining*, 240(6).
- Evans, J. R., D. Eberhart-Phillips, and C. H. Thurber (1994), User’s manual for SIMULPS12 for imaging Vp and Vp/Vs: a derivative of the “Thurber” tomographic inversion SIMUL3 for local earthquakes and explosions, *Open-File Report 94*, USGS, USGS, Menlo Park, California, series: Open-File Report.
- Fagereng, A., and J. F. A. Diener (2011), Non-volcanic tremor and discontinuous slab dehydration, *Geophysical Research Letters*, 38(15), 2011GL048214, doi:[10.1029/2011GL048214](https://doi.org/10.1029/2011GL048214).
- Ferrari, E., A. L. Rizzo, G. Capelli Ghioldi, A. Sciarra, G. Tamburello, F. Viveiros, S. Lovati, and M. Massa (2026), Multidisciplinary monitoring of mud volcanoes: a new perspective on fluids and seismicity link from the Salse di Nirano (Italy), *Natural Hazards*, 122(7), 311, doi:[10.1007/s11069-026-08071-3](https://doi.org/10.1007/s11069-026-08071-3).
- Flores, P. C. M., S. Kodaira, G. Kimura, K. Shiraishi, Y. Nakamura, G. Fujie, T. No, and Y. Kaiho (2024), Multiple seamount subduction and lithological variability possibly control pore fluid pressure and shallow slow earthquake activity in Nankai Trough off Muroto, doi:[10.22541/essoar.172927233.37434117/v1](https://doi.org/10.22541/essoar.172927233.37434117/v1).
- Frahm, L. (2025), Investigating elastic properties with full-waveform inversion at the Hikurangi margin, Thesis, Imperial College London, doi:[10.25560/117041](https://doi.org/10.25560/117041).

- François-Holden, C., S. Bannister, J. Beavan, J. Cousins, B. Field, R. McCaffrey, G. McVerry, M. Reyners, J. Ristau, S. Samsonov, and L. Wallace (2008), The Mw 6.6 Gisborne Earthquake of 2007: Preliminary Records and General Source Characterisation, *Bulletin of the New Zealand Society for Earthquake Engineering*, 41(4), 12.
- Fry, B., K. Chao, S. Bannister, Z. Peng, and L. Wallace (2011), Deep tremor in New Zealand triggered by the 2010 Mw8.8 Chile earthquake, *Geophysical Research Letters*, 38(15), doi:[10.1029/2011GL048319](https://doi.org/10.1029/2011GL048319), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2011GL048319>.
- Gale, N., K. Gledhill, M. Chadwick, and L. Wallace (2015), The Hikurangi Margin Continuous GNSS and Seismograph Network of New Zealand, *Seismological Research Letters*, 86(1), 101–108, doi:[10.1785/0220130181](https://doi.org/10.1785/0220130181).
- Gase, A., N. Bangs, H. Van Avendonk, D. Bassett, S. Henrys, R. Arai, G. Fujie, P. Barnes, S. Kodaira, D. Barker, and D. Okaya (2024), Volcanic crustal structure of the western Hikurangi Plateau (New Zealand) from marine seismic reflection imaging, *Geosphere*, doi:[10.1130/GES02744.1](https://doi.org/10.1130/GES02744.1).
- Gase, A. C., N. L. Bangs, D. M. Saffer, S. Han, P. K. Miller, R. E. Bell, R. Arai, S. A. Henrys, S. Kodaira, R. Davy, L. Frahm, and D. H. Barker (2023), Subducting volcanoclastic-rich upper crust supplies fluids for shallow megathrust and slow slip, *Science Advances*, 9(33), eadh0150, doi:[10.1126/sciadv.adh0150](https://doi.org/10.1126/sciadv.adh0150).
- Gauthier, A., N. Cubas, and L. Le Pourhiet (2026), Tectonic structures and stress state associated with seamount subduction in accretionary prism. Implications for slip behavior, *Earth and Planetary Science Letters*, 675, 119,767, doi:[10.1016/j.epsl.2025.119767](https://doi.org/10.1016/j.epsl.2025.119767).
- Ghisetti, F. C., P. M. Barnes, S. Ellis, A. A. Plaza-Faverola, and D. H. N. Barker (2016), The last 2 Myr of accretionary wedge construction in the central Hikurangi margin (North Island, New Zealand): Insights from structural modeling, *Geochemistry, Geophysics, Geosystems*, 17(7), 2661–2686, doi:[10.1002/2016GC006341](https://doi.org/10.1002/2016GC006341), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2016GC006341>.

- Ghosh, A., J. E. Vidale, and K. C. Creager (2012), Tremor asperities in the transition zone control evolution of slow earthquakes, *Journal of Geophysical Research: Solid Earth*, 117(B10), 2012JB009249, doi:[10.1029/2012JB009249](https://doi.org/10.1029/2012JB009249).
- Giggenbach, W., Y. Sano, and H. Wakita (1993), Isotopic composition of helium, and CO₂ and CH₄ contents in gases produced along the New Zealand part of a convergent plate boundary, *Geochimica et Cosmochimica Acta*, 57(14), 3427–3455, doi:[10.1016/0016-7037\(93\)90549-C](https://doi.org/10.1016/0016-7037(93)90549-C).
- GNS Science (2019), GeoNet Aotearoa New Zealand Stations Metadata Repository [Dataset], doi:[10.21420/0VY2-C144](https://doi.org/10.21420/0VY2-C144).
- GNS Science (2021), GeoNet Aotearoa New Zealand Seismic Digital Waveform Dataset [Dataset], doi:[10.21420/G19Y-9D40](https://doi.org/10.21420/G19Y-9D40).
- Goerzen, C. (2021), Mud volcano imaging using ambient seismic noise, Thesis, University of Victoria.
- Gomberg, J. (2010), Slow-slip phenomena in Cascadia from 2007 and beyond: A review, *Geological Society of America Bulletin*, 122(7-8), 963–978, doi:[10.1130/B30287.1](https://doi.org/10.1130/B30287.1).
- Gosselin, J. M., P. Audet, C. Estève, M. McLellan, S. G. Mosher, and A. J. Schaeffer (2020), Seismic evidence for megathrust fault-valve behavior during episodic tremor and slip, *Science Advances*, 6(4), eaay5174, doi:[10.1126/sciadv.aay5174](https://doi.org/10.1126/sciadv.aay5174).
- Gutenberg, B., and C. F. Richter (1944), Frequency of earthquakes in California, *Bulletin of the Seismological Society of America*, 34(4), 185–188.
- Haberland, C., A. Rietbrock, D. Lange, K. Bataille, and T. Dahm (2009), Structure of the seismogenic zone of the southcentral Chilean margin revealed by local earthquake traveltime tomography, *Journal of Geophysical Research: Solid Earth*, 114(B1), doi:[10.1029/2008JB005802](https://doi.org/10.1029/2008JB005802), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2008JB005802>.
- Heise, W., T. G. Caldwell, E. A. Bertrand, G. J. Hill, S. L. Bennie, and Y. Ogawa (2013), Changes in electrical resistivity track changes in tectonic plate coupling,

- Geophysical Research Letters*, 40(19), 5029–5033, doi:[10.1002/grl.50959](https://doi.org/10.1002/grl.50959), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/grl.50959>.
- Heise, W., E. A. Bertrand, T. G. Caldwell, Y. Ogawa, S. Bannister, S. L. Bennie, R. Hart, N. Palmer, K. H. Tseng, M. Fukai, M. Ishikawa, K. Seki, T. Nishizawa, and J. McGrath (2023), An electrical resistivity image of the Hikurangi subduction margin, *Geophysical Journal International*, 235(2), 1552–1564, doi:[10.1093/gji/ggad313](https://doi.org/10.1093/gji/ggad313).
- Helffrich, G., J. Wookey, and I. D. Bastow (2013), *The Seismic Analysis Code: A Primer and User's Guide*, 1st ed., Cambridge University Press.
- Henrys, S., D. Eberhart-Phillips, D. Bassett, R. Sutherland, D. Okaya, M. Savage, D. Evanzia, T. Stern, H. Sato, K. Mochizuki, T. Iwasaki, E. Kurashimo, A. Seward, and A. Wech (2020), Upper Plate Heterogeneity Along the Southern Hikurangi Margin, New Zealand, *Geophysical Research Letters*, 47(4), e2019GL085511, doi:[10.1029/2019GL085511](https://doi.org/10.1029/2019GL085511), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2019GL085511>.
- Henrys, S., D. Bassett, S. Ellis, L. Wallace, P. M. Barnes, D. Eberhart-Phillips, D. Saffer, and C. Boulton (2025), How Subduction Margin Processes and Properties Influence the Hikurangi Subduction Zone, *Annual Review of Earth and Planetary Sciences*, 53(1), 539–566, doi:[10.1146/annurev-earth-040523-115520](https://doi.org/10.1146/annurev-earth-040523-115520).
- Heslot, F., T. Baumberger, B. Perrin, B. Caroli, and C. Caroli (1994), Creep, stick-slip, and dry-friction dynamics: Experiments and a heuristic model, *Physical Review E*, 49(6), 4973–4988, doi:[10.1103/PhysRevE.49.4973](https://doi.org/10.1103/PhysRevE.49.4973).
- Hicks, S. P. (2015), Seismic properties and processes along the subduction plate interface: the February 2010 Mw 8.8 Maule, Chile earthquake, Ph.D. thesis, University of Liverpool, UK.
- Hicks, S. P., A. Rietbrock, I. M. Ryder, C.-S. Lee, and M. Miller (2014), Anatomy of a megathrust: The 2010 M8.8 Maule, Chile earthquake rupture zone imaged using seismic tomography, *Earth and Planetary Science Letters*, 405, 142–155, doi:[10.1016/j.epsl.2014.08.028](https://doi.org/10.1016/j.epsl.2014.08.028).
- Hirose, H., and K. Obara (2005), Repeating short- and long-term slow slip events with deep

- tremor activity around the Bungo channel region, southwest Japan, *Earth, Planets and Space*, 57(10), 961–972, doi:[10.1186/BF03351875](https://doi.org/10.1186/BF03351875).
- Hirose, H., K. Hirahara, F. Kimata, N. Fujii, and S. Miyazaki (1999), A slow thrust slip event following the two 1996 Hyuganada Earthquakes beneath the Bungo Channel, southwest Japan, *Geophysical Research Letters*, 26(21), 3237–3240, doi:[10.1029/1999GL010999](https://doi.org/10.1029/1999GL010999), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/1999GL010999>.
- Hoernle, K., F. Hauff, P. Van Den Bogaard, R. Werner, N. Mortimer, J. Geldmacher, D. Garbe-Schönberg, and B. Davy (2010), Age and geochemistry of volcanic rocks from the Hikurangi and Manihiki oceanic Plateaus, *Geochimica et Cosmochimica Acta*, 74(24), 7196–7219, doi:[10.1016/j.gca.2010.09.030](https://doi.org/10.1016/j.gca.2010.09.030).
- Hoernle, K., J. Gill, C. Timm, F. Hauff, R. Werner, D. Garbe-Schönberg, and M. Gutjahr (2020), Hikurangi Plateau subduction a trigger for Vitiaz arc splitting and Havre Trough opening (southwestern Pacific), *Geology*, 49(5), 536–540, doi:[10.1130/G48436.1](https://doi.org/10.1130/G48436.1).
- Hughes, L., C. J. Chamberlain, J. Townend, and A. M. Thomas (2021), A Repeating Earthquake Catalog From 2003 to 2020 for the Raukumara Peninsula, Northern Hikurangi Subduction Margin, New Zealand, *Geochemistry, Geophysics, Geosystems*, 22(5), e2021GC009670, doi:[10.1029/2021GC009670](https://doi.org/10.1029/2021GC009670), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021GC009670>.
- Hunt, T. M., and R. B. Glover (1995), Origin of mineral springs on the east coast, North Island, NZ, *17th NZ Geothermal Workshop*.
- Husen, S., E. Kissling, and E. R. Flueh (2000), Local earthquake tomography of shallow subduction in north Chile: A combined onshore and offshore study, *Journal of Geophysical Research: Solid Earth*, 105(B12), 28,183–28,198, doi:[10.1029/2000JB900229](https://doi.org/10.1029/2000JB900229).
- Hutton, L. K., and D. M. Boore (1987), The *ML* scale in Southern California, *Bulletin of the Seismological Society of America*, 77(6), 2074–2094, doi:[10.1785/BSSA0770062074](https://doi.org/10.1785/BSSA0770062074).
- Ide, S. (2012), Variety and spatial heterogeneity of tectonic tremor worldwide: Tec-

- tonic tremor worldwide, *Journal of Geophysical Research: Solid Earth*, 117(B3), doi:[10.1029/2011JB008840](https://doi.org/10.1029/2011JB008840).
- Ide, S., G. C. Beroza, D. R. Shelly, and T. Uchide (2007), A scaling law for slow earthquakes, *Nature*, 447(7140), 76–79, doi:[10.1038/nature05780](https://doi.org/10.1038/nature05780).
- Ito, Y., and K. Obara (2006), Very low frequency earthquakes within accretionary prisms are very low stress-drop earthquakes, *Geophysical Research Letters*, 33(9), 2006GL025,883, doi:[10.1029/2006GL025883](https://doi.org/10.1029/2006GL025883).
- Ito, Y., R. Hino, M. Kido, H. Fujimoto, Y. Osada, D. Inazu, Y. Ohta, T. Iinuma, M. Ohzono, S. Miura, M. Mishina, K. Suzuki, T. Tsuji, and J. Ashi (2013), Episodic slow slip events in the Japan subduction zone before the 2011 Tohoku-Oki earthquake, *Tectonophysics*, 600, 14–26, doi:[10.1016/j.tecto.2012.08.022](https://doi.org/10.1016/j.tecto.2012.08.022).
- Jacobs, K., M. Savage, and E. Smith (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, doi:[10.1080/00288306.2015.1127827](https://doi.org/10.1080/00288306.2015.1127827).
- Jacobs, K., S. A. Henrys, D. Okaya, H. Van Avendonk, J. Black, D. H. N. Barker, S. C. Karalilyadda, E. Kurashimo, W. R. Stratford, M. Savage, R. Sullivan, Z. R. Bruce, and L. Hughes (2020), Seismogenesis at Hikurangi Integrated Research Experiment (SHIRE) onshore seismic acquisition field report, doi:[10.21420/PEQZ-BR17](https://doi.org/10.21420/PEQZ-BR17).
- Kanamori, H. (1972), Mechanism of tsunami earthquakes, *Physics of the Earth and Planetary Interiors*, 6(5), 346–359, doi:[10.1016/0031-9201\(72\)90058-1](https://doi.org/10.1016/0031-9201(72)90058-1).
- Kaneko, Y., Y. Ito, B. Chow, L. M. Wallace, C. Tape, R. Grapenthin, E. D’Anastasio, S. Henrys, and R. Hino (2019), Ultra-long Duration of Seismic Ground Motion Arising From a Thick, Low-Velocity Sedimentary Wedge, *Journal of Geophysical Research: Solid Earth*, 124(10), 10,347–10,359, doi:[10.1029/2019JB017795](https://doi.org/10.1029/2019JB017795), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2019JB017795>.
- Katayama, I., T. Terada, K. Okazaki, and W. Tanikawa (2012), Episodic tremor and slow slip

- potentially linked to permeability contrasts at the Moho, *Nature Geoscience*, 5(10), 731–734, doi:[10.1038/ngeo1559](https://doi.org/10.1038/ngeo1559).
- Kato, A., K. Obara, T. Igarashi, H. Tsuruoka, S. Nakagawa, and N. Hirata (2012), Propagation of Slow Slip Leading Up to the 2011 M_w 9.0 Tohoku-Oki Earthquake, *Science*, 335(6069), 705–708, doi:[10.1126/science.1215141](https://doi.org/10.1126/science.1215141).
- Kennett, B. L. N., and E. R. Engdahl (1991), Traveltimes for global earthquake location and phase identification, *Geophysical Journal International*, 105(2), 429–465, doi:[10.1111/j.1365-246X.1991.tb06724.x](https://doi.org/10.1111/j.1365-246X.1991.tb06724.x).
- Kissling, E., W. L. Ellsworth, D. Eberhart-Phillips, and U. Kradolfer (1994), Initial reference models in local earthquake tomography, *Journal of Geophysical Research: Solid Earth*, 99(B10), 19,635–19,646, doi:[10.1029/93JB03138](https://doi.org/10.1029/93JB03138), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/93JB03138>.
- Kitajima, H., and D. M. Saffer (2012), Elevated pore pressure and anomalously low stress in regions of low frequency earthquakes along the Nankai Trough subduction megathrust, *Geophysical Research Letters*, 39(23), 2012GL053,793, doi:[10.1029/2012GL053793](https://doi.org/10.1029/2012GL053793).
- Kodaira, S., T. Iidaka, A. Kato, J.-O. Park, T. Iwasaki, and Y. Kaneda (2004), High Pore Fluid Pressure May Cause Silent Slip in the Nankai Trough, *Science*, 304(5675), 1295–1298, doi:[10.1126/science.1096535](https://doi.org/10.1126/science.1096535).
- Lay, T., and T. C. Wallace (1995), *Modern Global Seismology*, vol. 58, Elsevier.
- Leah, H., A. Fagereng, I. Bastow, R. Bell, V. Lane, S. Henrys, K. Jacobs, and B. Fry (2022), The northern Hikurangi margin three-dimensional plate interface in New Zealand remains rough 100 km from the trench, *Geology*, doi:[10.1130/G50272.1](https://doi.org/10.1130/G50272.1).
- Leighton, A. (2020), Engineering geological investigation of a recent mud volcano eruption, Waimata Valley, Ph.D. thesis, University of Auckland.
- Leighton, A., M. S. Brook, M. Cave, M. C. Rowe, A. Stanley, and J. F. Tunnicliffe (2022), Engineering geomorphological reconnaissance of the December 2018 Waimata Valley mud

- volcano eruption, Gisborne, New Zealand, *Quarterly Journal of Engineering Geology and Hydrogeology*, 55(4), qjagh2021–149, doi:[10.1144/qjagh2021-149](https://doi.org/10.1144/qjagh2021-149).
- Lentas, K. (2018), Towards routine determination of focal mechanisms obtained from first motion P-wave arrivals, *Geophysical Journal International*, 212(3), 1665–1686, doi:[10.1093/gji/ggx503](https://doi.org/10.1093/gji/ggx503).
- Lentas, K., D. Di Giacomo, J. Harris, and D. A. Storchak (2019), The ISC Bulletin as a comprehensive source of earthquake source mechanisms, *Earth System Science Data*, 11(2), 565–578, doi:[10.5194/essd-11-565-2019](https://doi.org/10.5194/essd-11-565-2019).
- Lewis, K. B., J. Collot, and S. E. Lallemand (1998), The dammed Hikurangi Trough: a channel-fed trench blocked by subducting seamounts and their wake avalanches (New Zealand–France GeodyNZ Project), *Basin Research*, 10(4), 441–468, doi:[10.1046/j.1365-2117.1998.00080.x](https://doi.org/10.1046/j.1365-2117.1998.00080.x).
- Lewis, K. B., S. E. Lallemand, and L. Carter (2004), Collapse in a Quaternary shelf basin off East Cape, New Zealand: Evidence for passage of a subducted seamount inboard of the Ruatoria giant avalanche, *New Zealand Journal of Geology and Geophysics*, 47(3), 415–429, doi:[10.1080/00288306.2004.9515067](https://doi.org/10.1080/00288306.2004.9515067).
- Li, H., M. Wei, D. Li, Y. Liu, Y. Kim, and S. Zhou (2018), Segmentation of Slow Slip Events in South Central Alaska Possibly Controlled by a Subducted Oceanic Plateau, *Journal of Geophysical Research: Solid Earth*, 123(1), 418–436, doi:[10.1002/2017JB014911](https://doi.org/10.1002/2017JB014911).
- Lin, G., P. M. Shearer, R. S. Matoza, P. G. Okubo, and F. Amelung (2014), Three-dimensional seismic velocity structure of Mauna Loa and Kilauea volcanoes in Hawaii from local seismic tomography, *Journal of Geophysical Research: Solid Earth*, 119(5), 4377–4392, doi:[10.1002/2013JB010820](https://doi.org/10.1002/2013JB010820).
- Litchfield, N., S. Ellis, K. Berryman, and A. Nicol (2007), Insights into subduction-related uplift along the Hikurangi Margin, New Zealand, using numerical modeling, *Journal of Geophysical Research: Earth Surface*, 112(F2), doi:[10.1029/2006JF000535](https://doi.org/10.1029/2006JF000535), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2006JF000535>.

- Litchfield, N., R. Van Dissen, R. Sutherland, P. Barnes, S. Cox, R. Norris, R. Beavan, R. Langridge, P. Villamor, K. Berryman, M. Stirling, A. Nicol, S. Nodder, G. Lamarche, D. Barrell, J. Pettinga, T. Little, N. Pondard, J. Mountjoy, and K. Clark (2014), A model of active faulting in New Zealand, *New Zealand Journal of Geology and Geophysics*, 57(1), 32–56, doi:[10.1080/00288306.2013.854256](https://doi.org/10.1080/00288306.2013.854256).
- Litchfield, N. J., D. Townsend, A. Howell, and R. Morgenstern (2022), Active fault mapping in Tairāwhiti Gisborne District using LiDAR data, *GNS Science*.
- Liu, Y., and J. R. Rice (2007), Spontaneous and triggered aseismic deformation transients in a subduction fault model, *Journal of Geophysical Research: Solid Earth*, 112(B9), doi:[10.1029/2007JB004930](https://doi.org/10.1029/2007JB004930), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2007JB004930>.
- Lomax, A., J. Virieux, P. Volant, and C. Berge-Thierry (2000), *Advances in Seismic Event Location*, 1st ed., Springer, Dordrecht.
- Lomax, A., A. Zollo, P. Capuano, and J. Virieux (2001), Precise, absolute earthquake location under Somma-Vesuvius volcano using a new three-dimensional velocity model, *Geophysical Journal International*, 146(2), 313–331, doi:[10.1046/j.0956-540x.2001.01444.x](https://doi.org/10.1046/j.0956-540x.2001.01444.x).
- Lomax, A., A. Michelini, and A. Curtis (2009), Earthquake Location, Direct, Global-Search Methods, in *Encyclopedia of Complexity and Systems Science*, Springer, New York, NY, doi:[10.1007/978-0-387-30440-3_150](https://doi.org/10.1007/978-0-387-30440-3_150).
- Lomax, A., T. Megies, S. Ho, J.-M. Saurel, and S. Luca (2023), alomax/NonLinLoc: Intermittent release to enable citing and reference of NonLinLoc through Zenodo [Software]. Zenodo., doi:[10.5281/zenodo.8046170](https://doi.org/10.5281/zenodo.8046170).
- Matsuzawa, T., Y. Asano, and K. Obara (2015), Very low frequency earthquakes off the Pacific coast of Tohoku, Japan, *Geophysical Research Letters*, 42(11), 4318–4325, doi:[10.1002/2015GL063959](https://doi.org/10.1002/2015GL063959).
- Mazengarb, C., and I. G. Speden (2000), *Geology of the Raukumara Area*, vol. 6. *Institute of Geological & Nuclear Sciences 1:250 000 geological map*, GNS Science, journal Abbreviation:

Institute of Geological and Nuclear Sciences 1:250,000 Geological Map Publication Title:
Institute of Geological and Nuclear Sciences 1:250,000 Geological Map.

McCaffrey, R. (2009), Time-dependent inversion of three-component continuous GPS for steady and transient sources in northern Cascadia, *Geophysical Research Letters*, 36(7), 2008GL036,784, doi:[10.1029/2008GL036784](https://doi.org/10.1029/2008GL036784).

McCaffrey, R., L. M. Wallace, and J. Beavan (2008), Slow slip and frictional transition at low temperature at the Hikurangi subduction zone, *Nature Geoscience*, 1(5), 316–320, doi:[10.1038/ngeo178](https://doi.org/10.1038/ngeo178).

McNamara, D. E. (2004), Ambient Noise Levels in the Continental United States, *Bulletin of the Seismological Society of America*, 94(4), 1517–1527, doi:[10.1785/012003001](https://doi.org/10.1785/012003001).

Merry, T. A. J. (2022), Upper mantle seismic anisotropy in the eastern Mediterranean and seismicity in Cyprus, PhD Thesis, Imperial College London, UK.

Michel, S., R. Jolivet, E. Klein, and L. Maubant (2025), 14 Years of Slip on the Hikurangi Subduction Zone, *Journal of Geophysical Research: Solid Earth*, 130(7), e2024JB030,865, doi:[10.1029/2024JB030865](https://doi.org/10.1029/2024JB030865), eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2024JB030865>.

Mochizuki, K., T. Yamada, M. Shinohara, Y. Yamanaka, and T. Kanazawa (2008), Weak Interplate Coupling by Seamounts and Repeating $M \sim 7$ Earthquakes, *Science*, 321(5893), 1194–1197, doi:[10.1126/science.1160250](https://doi.org/10.1126/science.1160250).

Mochizuki, K., R. Sutherland, S. Henrys, D. Bassett, H. Van Avendonk, R. Arai, S. Kodaira, G. Fujie, Y. Yamamoto, N. Bangs, and D. Barker (2019), Recycling of depleted continental mantle by subduction and plumes at the Hikurangi Plateau large igneous province, southwestern Pacific Ocean, *Geology*, 47(8), 795–798, doi:[10.1130/G46250.1](https://doi.org/10.1130/G46250.1).

Mochizuki, K., S. Henrys, D. Haijima, E. Warren-Smith, and B. Fry (2021), Seismicity and velocity structure in the vicinity of repeating slow slip earthquakes, northern Hikurangi subduction zone, New Zealand, *Earth and Planetary Science Letters*, 563, 116,887, doi:[10.1016/j.epsl.2021.116887](https://doi.org/10.1016/j.epsl.2021.116887).

- Montelli, R., G. Nolet, F. A. Dahlen, G. Masters, E. R. Engdahl, and S.-H. Hung (2004), Finite-Frequency Tomography Reveals a Variety of Plumes in the Mantle, *Science*, 303(5656), 338–343, doi:[10.1126/science.1092485](https://doi.org/10.1126/science.1092485).
- Mortimer, N., and D. Parkinson (1996), Hikurangi Plateau: A Cretaceous large igneous province in the southwest Pacific Ocean, *Journal of Geophysical Research: Solid Earth*, 101(B1), 687–696, doi:[10.1029/95JB03037](https://doi.org/10.1029/95JB03037), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/95JB03037>.
- Mousavi, S. M., Y. Sheng, W. Zhu, and G. C. Beroza (2019), STanford EArthquake Dataset (STEAD): A Global Data Set of Seismic Signals for AI, *IEEE Access*, 7, 179,464–179,476, doi:[10.1109/ACCESS.2019.2947848](https://doi.org/10.1109/ACCESS.2019.2947848), conference Name: IEEE Access.
- Mousavi, S. M., W. L. Ellsworth, W. Zhu, L. Y. Chuang, and G. C. Beroza (2020), Earthquake transformer—an attentive deep-learning model for simultaneous earthquake detection and phase picking, *Nature Communications*, 11(1), 3952, doi:[10.1038/s41467-020-17591-w](https://doi.org/10.1038/s41467-020-17591-w).
- Münchmeyer, J., J. Woollam, A. Rietbrock, F. Tilmann, D. Lange, T. Bornstein, T. Diehl, C. Giunchi, F. Haslinger, D. Jozinović, A. Michelini, J. Saul, and H. Soto (2022), Which Picker Fits My Data? A Quantitative Evaluation of Deep Learning Based Seismic Pickers, *Journal of Geophysical Research: Solid Earth*, 127(1), e2021JB023,499, doi:[10.1029/2021JB023499](https://doi.org/10.1029/2021JB023499), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021JB023499>.
- Nakajima, J., and N. Uchida (2018), Repeated drainage from megathrusts during episodic slow slip, *Nature Geoscience*, 11(5), 351–356, doi:[10.1038/s41561-018-0090-z](https://doi.org/10.1038/s41561-018-0090-z).
- Nicol, A., C. Mazengarb, F. Chanier, G. Rait, C. Uruski, and L. Wallace (2007), Tectonic evolution of the active Hikurangi subduction margin, New Zealand, since the Oligocene, *Tectonics*, 26(4), 2006TC002,090, doi:[10.1029/2006TC002090](https://doi.org/10.1029/2006TC002090).
- Nishikawa, T., T. Nishimura, and Y. Okada (2021), Earthquake Swarm Detection Along the Hikurangi Trench, New Zealand: Insights Into the Relationship Between Seismicity and Slow Slip Events, *Journal of Geophysical Research*

- search: *Solid Earth*, 126(4), e2020JB020,618, doi:[10.1029/2020JB020618](https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JB020618), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JB020618>.
- Nishikawa, T., S. Ide, and T. Nishimura (2023), A review on slow earthquakes in the Japan Trench, *Progress in Earth and Planetary Science*, 10(1), 1, doi:[10.1186/s40645-022-00528-w](https://doi.org/10.1186/s40645-022-00528-w).
- Nishimura, T. (2014), Short-term slow slip events along the Ryukyu Trench, southwestern Japan, observed by continuous GNSS, *Progress in Earth and Planetary Science*, 1(1), 22, doi:[10.1186/s40645-014-0022-5](https://doi.org/10.1186/s40645-014-0022-5).
- Obana, K., and S. Kodaira (2009), Low-frequency tremors associated with reverse faults in a shallow accretionary prism, *Earth and Planetary Science Letters*, 287(1-2), 168–174, doi:[10.1016/j.epsl.2009.08.005](https://doi.org/10.1016/j.epsl.2009.08.005).
- Obara, K. (2002), Nonvolcanic Deep Tremor Associated with Subduction in Southwest Japan, *Science*, 296(5573), 1679–1681, doi:[10.1126/science.1070378](https://doi.org/10.1126/science.1070378).
- Obara, K., H. Hirose, F. Yamamizu, and K. Kasahara (2004), Episodic slow slip events accompanied by non-volcanic tremors in southwest Japan subduction zone, *Geophysical Research Letters*, 31(23), doi:[10.1029/2004GL020848](https://doi.org/10.1029/2004GL020848), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2004GL020848>.
- Ozawa, S., H. Suito, and M. Tobita (2007), Occurrence of quasi-periodic slow-slip off the east coast of the Boso peninsula, Central Japan, *Earth, Planets and Space*, 59(12), 1241–1245, doi:[10.1186/BF03352072](https://doi.org/10.1186/BF03352072).
- Ozawa, S., Y. Yang, and E. M. Dunham (2024), Fault-Valve Instability: A Mechanism for Slow Slip Events, *Journal of Geophysical Research: Solid Earth*, 129(10), e2024JB029,165, doi:[10.1029/2024JB029165](https://doi.org/10.1029/2024JB029165), _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2024JB029165>.
- Paige, C., and M. Saunders (1982), LSQR: an algorithm for sparse linear equations and sparse least squares, *ACM Transactions on Mathematical Software*, 8(1), 43–71, doi:[10.1145/355994.355989](https://doi.org/10.1145/355994.355989).

- Peacock, S. M. (2009), Thermal and metamorphic environment of subduction zone episodic tremor and slip, *Journal of Geophysical Research: Solid Earth*, 114(B8), 2008JB005,978, doi:[10.1029/2008JB005978](https://doi.org/10.1029/2008JB005978).
- Peacock, S. M., N. I. Christensen, M. G. Bostock, and P. Audet (2011), High pore pressures and porosity at 35 km depth in the Cascadia subduction zone, *Geology*, 39(5), 471–474, doi:[10.1130/G31649.1](https://doi.org/10.1130/G31649.1).
- Pedley, K. L., P. M. Barnes, J. R. Pettinga, and K. B. Lewis (2010), Seafloor structural geomorphic evolution of the accretionary frontal wedge in response to seamount subduction, Poverty Indentation, New Zealand, *Marine Geology*, 270(1-4), 119–138, doi:[10.1016/j.margeo.2009.11.006](https://doi.org/10.1016/j.margeo.2009.11.006).
- Peng, Z., and J. Gomberg (2010), An integrated perspective of the continuum between earthquakes and slow-slip phenomena, *Nature Geoscience*, 3(9), 599–607, doi:[10.1038/ngeo940](https://doi.org/10.1038/ngeo940).
- Perez-Silva, A., Y. Kaneko, M. Savage, L. Wallace, D. Li, and C. Williams (2022), Segmentation of Shallow Slow Slip Events at the Hikurangi Subduction Zone Explained by Along-Strike Changes in Fault Geometry and Plate Convergence Rates, *Journal of Geophysical Research: Solid Earth*, 127(1), e2021JB022,913, doi:[10.1029/2021JB022913](https://doi.org/10.1029/2021JB022913), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021JB022913>.
- Perez-Silva, A., T. Wang, L. Wallace, M. Bebbington, and P. Denys (2025), Assessing Occurrence Patterns of Shallow Hikurangi Slow Slip Events Using Renewal Processes, *Geophysical Research Letters*, 52(13), e2025GL116,605, doi:[10.1029/2025GL116605](https://doi.org/10.1029/2025GL116605), eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2025GL116605>.
- Petersen, T., K. Gledhill, M. Chadwick, N. Gale, and J. Ristau (2011), The New Zealand National Seismograph Network, *Seismological Research Letters*, 82, 9–20, doi:[10.1785/gssrl.82.1.9](https://doi.org/10.1785/gssrl.82.1.9).
- Peterson, J. (1993), Observations and Modeling of Seismic Background Noise, *U.S. Geological Survey*, 93(322).

- Pettinga, J. R. (1982), Upper Cenozoic structural history, coastal Southern Hawke's Bay, New Zealand, *New Zealand Journal of Geology and Geophysics*, 25(2), 149–191, doi:[10.1080/00288306.1982.10421407](https://doi.org/10.1080/00288306.1982.10421407).
- Peña Castro, A. F., S. L. Dougherty, R. M. Harrington, and E. S. Cochran (2019), Delayed Dynamic Triggering of Disposal-Induced Earthquakes Observed by a Dense Array in Northern Oklahoma, *Journal of Geophysical Research: Solid Earth*, 124(4), 3766–3781, doi:[10.1029/2018JB017150](https://doi.org/10.1029/2018JB017150), eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2018JB017150>.
- Pita-Slim, O., C. J. Chamberlain, J. Townend, and E. Warren-Smith (2023), Parametric Testing of EQTransformer's Performance against a High-Quality, Manually Picked Catalog for Reliable and Accurate Seismic Phase Picking, *The Seismic Record*, 3(4), 332–341, doi:[10.1785/0320230024](https://doi.org/10.1785/0320230024).
- Podvin, P., and I. Lecomte (1991), Finite difference computation of traveltimes in very contrasted velocity models: a massively parallel approach and its associated tools, *Geophysical Journal International*, 105(1), 271–284, doi:[10.1111/j.1365-246X.1991.tb03461.x](https://doi.org/10.1111/j.1365-246X.1991.tb03461.x).
- Pohatu, P., T. Warmenhoven, A. Rae, and D. Bradshaw (2010), Low Enthalpy Geothermal Energy Resources for Rural Māori Communities–Te Puia Springs, East Coast, North Island New Zealand, *Proceedings World Geothermal Congress*.
- Radiguet, M., F. Cotton, M. Vergnolle, M. Campillo, B. Valette, V. Kostoglodov, and N. Cotte (2011), Spatial and temporal evolution of a long term slow slip event: the 2006 Guerrero Slow Slip Event: Evolution of the 2006 Guerrero SSE, *Geophysical Journal International*, 184(2), 816–828, doi:[10.1111/j.1365-246X.2010.04866.x](https://doi.org/10.1111/j.1365-246X.2010.04866.x).
- Radiguet, M., H. Perfettini, N. Cotte, A. Gualandi, B. Valette, V. Kostoglodov, T. Lhomme, A. Walpersdorf, E. Cabral Cano, and M. Campillo (2016), Triggering of the 2014 Mw7.3 Papanoa earthquake by a slow slip event in Guerrero, Mexico, *Nature Geoscience*, 9(11), 829–833, doi:[10.1038/ngeo2817](https://doi.org/10.1038/ngeo2817).

- Reyes, A. G., B. W. Christenson, and K. Faure (2010), Sources of solutes and heat in low-enthalpy mineral waters and their relation to tectonic setting, New Zealand, *Journal of Volcanology and Geothermal Research*, 192(3), 117–141, doi:[10.1016/j.jvolgeores.2010.02.015](https://doi.org/10.1016/j.jvolgeores.2010.02.015).
- Reyes, A. G., S. M. Ellis, B. W. Christenson, and S. Henrys (2022), Fluid flow rates and compositions and water–rock interaction in the Hikurangi margin forearc, New Zealand, *Chemical Geology*, 614, 121,169, doi:[10.1016/j.chemgeo.2022.121169](https://doi.org/10.1016/j.chemgeo.2022.121169).
- Reyners, M., and D. Eberhart-Phillips (2009), Small earthquakes provide insight into plate coupling and fluid distribution in the Hikurangi subduction zone, New Zealand, *Earth and Planetary Science Letters*, 282(1-4), 299–305, doi:[10.1016/j.epsl.2009.03.034](https://doi.org/10.1016/j.epsl.2009.03.034).
- Reyners, M., D. Eberhart-Phillips, and G. Stuart (1999), A three-dimensional image of shallow subduction: crustal structure of the Raukumara Peninsula, New Zealand, *Geophysical Journal International*, 137(3), 873–890, doi:[10.1046/j.1365-246x.1999.00842.x](https://doi.org/10.1046/j.1365-246x.1999.00842.x).
- Reyners, M., D. Eberhart-Phillips, and S. Bannister (2011), Tracking repeated subduction of the Hikurangi Plateau beneath New Zealand, *Earth and Planetary Science Letters*, 311(1-2), 165–171, doi:[10.1016/j.epsl.2011.09.011](https://doi.org/10.1016/j.epsl.2011.09.011).
- Rhoades, D. A., A. Christophersen, S. Bourguignon, J. Ristau, and J. Salichon (2020), A Depth-Dependent Local Magnitude Scale for New Zealand Earthquakes Consistent with Moment Magnitude, *Bulletin of the Seismological Society of America*, 111(2), 1056–1066, doi:[10.1785/0120200252](https://doi.org/10.1785/0120200252).
- Ridd, M. F. (1970), Mud Volcanoes in New Zealand, *AAPG Bulletin*, 54, doi:[10.1306/5D25CA19-16C1-11D7-8645000102C1865D](https://doi.org/10.1306/5D25CA19-16C1-11D7-8645000102C1865D).
- Rogers, G., and H. Dragert (2003), Episodic Tremor and Slip on the Cascadia Subduction Zone: The Chatter of Silent Slip, *Science*, 300(5627), 1942–1943, doi:[10.1126/science.1084783](https://doi.org/10.1126/science.1084783).
- Romanet, P., and S. Ide (2019), Ambient tectonic tremors in Manawatu, Cape Turnagain, Marlborough, and Puysegur, New Zealand, *Earth, Planets and Space*, 71(1), 59, doi:[10.1186/s40623-019-1039-1](https://doi.org/10.1186/s40623-019-1039-1).

- Ross, Z. E., M.-A. Meier, and E. Hauksson (2018), P Wave Arrival Picking and First-Motion Polarity Determination With Deep Learning, *Journal of Geophysical Research: Solid Earth*, 123(6), 5120–5129, doi:[10.1029/2017JB015251](https://doi.org/10.1029/2017JB015251), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2017JB015251>.
- Saffer, D. M., and L. M. Wallace (2015), The frictional, hydrologic, metamorphic and thermal habitat of shallow slow earthquakes, *Nature Geoscience*, 8(8), 594–600, doi:[10.1038/ngeo2490](https://doi.org/10.1038/ngeo2490).
- Scholz, C. H. (1998), Earthquakes and friction laws, *Nature*, 391(6662), 37–42, doi:[10.1038/34097](https://doi.org/10.1038/34097).
- Scholz, C. H. (2002), *The Mechanics of Earthquakes and Faulting*, 2 ed., Cambridge University Press, doi:[10.1017/CBO9780511818516](https://doi.org/10.1017/CBO9780511818516).
- Scholz, C. H., and C. Small (1997), The effect of seamount subduction on seismic coupling, *Geology*, 25(6), 487, doi:[10.1130/0091-7613\(1997\)025<0487:teosso>2.3.co;2](https://doi.org/10.1130/0091-7613(1997)025<0487:teosso>2.3.co;2).
- Schwartz, S. Y., and J. M. Rokosky (2007), Slow slip events and seismic tremor at circum-Pacific subduction zones, *Reviews of Geophysics*, 45(3), 2006RG000,208, doi:[10.1029/2006RG000208](https://doi.org/10.1029/2006RG000208).
- Seebeck, H., R. J. Van Dissen, N. J. Litchfield, P. M. Barnes, A. Nicol, R. M. Langridge, D. J. A. Barrell, P. Villamor, S. M. Ellis, and M. S. Rattenbury (2022), New Zealand Community Fault Model - version 1.0, doi:[10.21420/GA7S-BS61](https://doi.org/10.21420/GA7S-BS61).
- Segovia, M., Y. Font, M. Régnier, P. Charvis, A. Galve, J. Nocquet, P. Jarrín, Y. Hello, M. Ruiz, and A. Pazmiño (2018), Seismicity Distribution Near a Subducting Seamount in the Central Ecuadorian Subduction Zone, Space-Time Relation to a Slow-Slip Event, *Tectonics*, 37(7), 2106–2123, doi:[10.1029/2017TC004771](https://doi.org/10.1029/2017TC004771).
- Shaddox, H. R., and S. Y. Schwartz (2019), Subducted seamount diverts shallow slow slip to the forearc of the northern Hikurangi subduction zone, New Zealand, *Geology*, 47(5), 415–418, doi:[10.1130/G45810.1](https://doi.org/10.1130/G45810.1).

- Sibson, R. H. (1990), Conditions for fault-valve behaviour, *Geological Society, London, Special Publications*, 54(1), 15–28, doi:[10.1144/GSL.SP.1990.054.01.02](https://doi.org/10.1144/GSL.SP.1990.054.01.02).
- Snoke, J. A. (2003), 85.12 FOCMEC: FOCal MEChanism determinations, in *International Geophysics*, vol. 81, pp. 1629–1630, Elsevier, doi:[10.1016/S0074-6142\(03\)80291-7](https://doi.org/10.1016/S0074-6142(03)80291-7).
- Song, T.-R. A., D. V. Helmberger, M. R. Brudzinski, R. W. Clayton, P. Davis, X. Pérez-Campos, and S. K. Singh (2009), Subducting Slab Ultra-Slow Velocity Layer Coincident with Silent Earthquakes in Southern Mexico, *Science*, 324(5926), 502–506, doi:[10.1126/science.1167595](https://doi.org/10.1126/science.1167595).
- Sun, T., D. Saffer, and S. Ellis (2020), Mechanical and hydrological effects of seamount subduction on megathrust stress and slip, *Nature Geoscience*, 13(3), 249–255, doi:[10.1038/s41561-020-0542-0](https://doi.org/10.1038/s41561-020-0542-0).
- Thurber, C., and D. Eberhart-Phillips (1999), Local earthquake tomography with flexible gridding, *Computers & Geosciences*, 25, 809–818.
- Thurber, C. H. (1983), Earthquake locations and three-dimensional crustal structure in the Coyote Lake Area, central California, *Journal of Geophysical Research: Solid Earth*, 88(B10), 8226–8236, doi:[10.1029/JB088iB10p08226](https://doi.org/10.1029/JB088iB10p08226), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/JB088iB10p08226>.
- Thurber, C. H. (1993), Local earthquake tomography: velocities and Vp/Vs theory, *Seismic Tomography: Theory and Practice*, pp. 563–583.
- Timm, C., K. Hoernle, R. Werner, F. Hauff, P. V. Den Bogaard, J. White, N. Mortimer, and D. Garbe-Schönberg (2010), Temporal and geochemical evolution of the Cenozoic intraplate volcanism of Zealandia, *Earth-Science Reviews*, 98(1-2), 38–64, doi:[10.1016/j.earscirev.2009.10.002](https://doi.org/10.1016/j.earscirev.2009.10.002).
- Todd, E. K., and S. Y. Schwartz (2016), Tectonic tremor along the northern Hikurangi Margin, New Zealand, between 2010 and 2015, *Journal of Geophysical Research: Solid Earth*, 121(12), 8706–8719, doi:[10.1002/2016JB013480](https://doi.org/10.1002/2016JB013480), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2016JB013480>.

- Todd, E. K., S. Y. Schwartz, K. Mochizuki, L. M. Wallace, A. F. Sheehan, S. C. Webb, C. A. Williams, J. Nakai, J. Yancey, B. Fry, S. Henrys, and Y. Ito (2018), Earthquakes and Tremor Linked to Seamount Subduction During Shallow Slow Slip at the Hikurangi Margin, New Zealand, *Journal of Geophysical Research: Solid Earth*, 123(8), 6769–6783, doi:[10.1029/2018JB016136](https://doi.org/10.1029/2018JB016136), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2018JB016136>.
- Toomey, D. R., and G. R. Foulger (1989), Tomographic inversion of local earthquake data from the Hengill-Grensdalur Central Volcano Complex, Iceland, *Journal of Geophysical Research: Solid Earth*, 94(B12), 17,497–17,510, doi:[10.1029/JB094iB12p17497](https://doi.org/10.1029/JB094iB12p17497).
- Townend, J., S. Sherburn, R. Arnold, C. Boese, and L. Woods (2012), Three-dimensional variations in present-day tectonic stress along the Australia–Pacific plate boundary in New Zealand, *Earth and Planetary Science Letters*, 353–354, 47–59, doi:[10.1016/j.epsl.2012.08.003](https://doi.org/10.1016/j.epsl.2012.08.003).
- Tozer, B., D. T. Sandwell, W. H. F. Smith, C. Olson, J. R. Beale, and P. Wessel (2019), Global Bathymetry and Topography at 15 Arc Sec: SRTM15+, *Earth and Space Science*, 6(10), 1847–1864, doi:[10.1029/2019EA000658](https://doi.org/10.1029/2019EA000658).
- Um, J., and C. Thurber (1987), A fast algorithm for two-point seismic ray tracing, *Bulletin of the Seismological Society of America*, 77(3), 972–986.
- Vallée, M., J.-M. Nocquet, J. Battaglia, Y. Font, M. Segovia, M. Régnier, P. Mothes, P. Jarrin, D. Cisneros, S. Vaca, H. Yepes, X. Martin, N. Béthoux, and M. Chlieh (2013), Intense interface seismicity triggered by a shallow slow slip event in the Central Ecuador subduction zone, *Journal of Geophysical Research: Solid Earth*, 118(6), 2965–2981, doi:[10.1002/jgrb.50216](https://doi.org/10.1002/jgrb.50216), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/jgrb.50216>.
- van der Elst, N. J., and E. E. Brodsky (2010), Connecting near-field and far-field earthquake triggering to dynamic strain, *Journal of Geophysical Research: Solid Earth*, 115(B7), doi:[10.1029/2009JB006681](https://doi.org/10.1029/2009JB006681), eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2009JB006681>.

- Wadati, K., and S. Oki (1933), On the Travel Time of Earthquake Waves. (Part II), *Journal of the Meteorological Society of Japan. Ser. II*, 11(1), 14–28, doi:[10.2151/jmsj1923.11.1_14](https://doi.org/10.2151/jmsj1923.11.1_14).
- Waldhauser, F. (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, doi:[10.1785/0120000006](https://doi.org/10.1785/0120000006).
- Wallace, L., A. Sheehan, Susan Schwartz, and S. Webb (2014), Hikurangi Ocean Bottom Investigation of Tremor and Slow Slip [Data set], doi:[10.7914/SN/YH_2014](https://doi.org/10.7914/SN/YH_2014).
- Wallace, L., D. Saffer, P. Barnes, I. Pecher, K. Petronotis, L. LeVay, and Expedition 372/375 Scientists (2019), *Hikurangi Subduction Margin Coring, Logging, and Observatories, Proceedings of the International Ocean Discovery Program*, vol. 372B/375, International Ocean Discovery Program, doi:[10.14379/iodp.proc.372B375.2019](https://doi.org/10.14379/iodp.proc.372B375.2019).
- Wallace, L. M. (2020), Slow Slip Events in New Zealand, *Annual Review of Earth and Planetary Sciences*, 48(1), 175–203, doi:[10.1146/annurev-earth-071719-055104](https://doi.org/10.1146/annurev-earth-071719-055104).
- Wallace, L. M., and J. Beavan (2006), A large slow slip event on the central Hikurangi subduction interface beneath the Manawatu region, North Island, New Zealand, *Geophysical Research Letters*, 33(11), doi:[10.1029/2006GL026009](https://doi.org/10.1029/2006GL026009), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2006GL026009>.
- Wallace, L. M., and J. Beavan (2010), Diverse slow slip behavior at the Hikurangi subduction margin, New Zealand, *Journal of Geophysical Research: Solid Earth*, 115(B12), doi:[10.1029/2010JB007717](https://doi.org/10.1029/2010JB007717), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2010JB007717>.
- Wallace, L. M., and D. Eberhart-Phillips (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, doi:[10.1002/2013GL057682](https://doi.org/10.1002/2013GL057682), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2013GL057682>.

- Wallace, L. M., J. Beavan, R. McCaffrey, and D. Darby (2004), Subduction zone coupling and tectonic block rotations in the North Island, New Zealand, *Journal of Geophysical Research: Solid Earth*, 109(B12), doi:[10.1029/2004JB003241](https://doi.org/10.1029/2004JB003241), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2004JB003241>.
- Wallace, L. M., M. Reyners, U. Cochran, S. Bannister, P. M. Barnes, K. Berryman, G. Downes, D. Eberhart-Phillips, A. Fagereng, S. Ellis, A. Nicol, R. McCaffrey, R. J. Beavan, S. Henrys, R. Sutherland, D. H. N. Barker, N. Litchfield, J. Townend, R. Robinson, R. Bell, K. Wilson, and W. Power (2009), Characterizing the seismogenic zone of a major plate boundary subduction thrust: Hikurangi Margin, New Zealand, *Geochemistry, Geophysics, Geosystems*, 10(10), doi:[10.1029/2009GC002610](https://doi.org/10.1029/2009GC002610), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2009GC002610>.
- Wallace, L. M., J. Beavan, S. Bannister, and C. Williams (2012a), 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: Solid Earth*, 117(B11), doi:[10.1029/2012JB009489](https://doi.org/10.1029/2012JB009489), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2012JB009489>.
- Wallace, L. M., P. Barnes, J. Beavan, R. Van Dissen, N. Litchfield, J. Mountjoy, R. Langridge, G. Lamarche, and N. Pondard (2012b), 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(B2), doi:[10.1029/2011JB008640](https://doi.org/10.1029/2011JB008640), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2011JB008640>.
- Wallace, L. M., S. C. Webb, Y. Ito, K. Mochizuki, R. Hino, S. Henrys, S. Y. Schwartz, and A. F. Sheehan (2016), Slow slip near the trench at the Hikurangi subduction zone, New Zealand, *Science*, 352(6286), 701–704, doi:[10.1126/science.aaf2349](https://doi.org/10.1126/science.aaf2349).
- Wallace, L. M., Y. Kaneko, S. Hreinsdóttir, I. Hamling, Z. Peng, N. Bartlow, E. D’Anastasio, and B. Fry (2017), Large-scale dynamic triggering of shallow slow slip enhanced by overlying sedimentary wedge, *Nature Geoscience*, 10(10), 765–770, doi:[10.1038/ngeo3021](https://doi.org/10.1038/ngeo3021).

- Wallace, L. M., S. Hreinsdóttir, S. Ellis, I. Hamling, E. D’Anastasio, and P. Denys (2018), Triggered Slow Slip and Afterslip on the Southern Hikurangi Subduction Zone Following the Kaikōura Earthquake, *Geophysical Research Letters*, 45(10), 4710–4718, doi:[10.1002/2018GL077385](https://doi.org/10.1002/2018GL077385), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2018GL077385>.
- Wang, K., and S. L. Bilek (2011), Do subducting seamounts generate or stop large earthquakes?, *Geology*, 39(9), 819–822, doi:[10.1130/G31856.1](https://doi.org/10.1130/G31856.1).
- Warren-Smith, E., B. Fry, L. Wallace, E. Chon, S. Henrys, A. Sheehan, K. Mochizuki, S. Schwartz, S. Webb, and S. Lebedev (2019), Episodic stress and fluid pressure cycling in subducting oceanic crust during slow slip, *Nature Geoscience*, 12(6), 475–481, doi:[10.1038/s41561-019-0367-x](https://doi.org/10.1038/s41561-019-0367-x).
- Warren-Smith, E., K. Jacobs, C. Rollins, C. J. Chamberlain, D. Eberhart-Phillips, and C. Williams (2024), A quantitative assessment of GeoNet earthquake location quality in Aotearoa New Zealand, *New Zealand Journal of Geology and Geophysics*, pp. 1–14, doi:[10.1080/00288306.2024.2421309](https://doi.org/10.1080/00288306.2024.2421309).
- Watson, S. J., J. J. Mountjoy, P. M. Barnes, G. J. Crutchley, G. Lamarche, B. Higgs, J. Hillman, A. R. Orpin, A. Micallef, H. Neil, J. Mitchell, A. Pallentin, T. Kane, S. Woelz, D. Bowden, A. A. Rowden, and I. A. Pecher (2020), Focused fluid seepage related to variations in accretionary wedge structure, Hikurangi margin, New Zealand, *Geology*, 48(1), 56–61, doi:[10.1130/G46666.1](https://doi.org/10.1130/G46666.1).
- Webb, T. H., and H. Anderson (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, doi:[10.1046/j.1365-246x.1998.00531.x](https://doi.org/10.1046/j.1365-246x.1998.00531.x).
- Wiemer, S. (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, doi:[10.1785/0119990114](https://doi.org/10.1785/0119990114).

- Williams, C. A., and L. M. Wallace (2015), Effects of material property variations on slip estimates for subduction interface slow-slip events, *Geophysical Research Letters*, 42(4), 1113–1121, doi:[10.1002/2014GL062505](https://doi.org/10.1002/2014GL062505).
- Williams, C. A., D. Eberhart-Phillips, S. Bannister, D. H. N. Barker, S. Henrys, M. Reyners, and R. Sutherland (2013), Revised Interface Geometry for the Hikurangi Subduction Zone, New Zealand, *Seismological Research Letters*, 84(6), 1066–1073, doi:[10.1785/0220130035](https://doi.org/10.1785/0220130035).
- Wood, J., A. C. Whittaker, R. E. Bell, S. M. Coveney, and H. D. Kranis (2025), Resolving Coseismic and Aseismic Normal Fault Slip Behaviors From InSAR Time Series of the European Ground Motion Service (EGMS), *Tectonics*, 44(9), e2025TC008,904, doi:[10.1029/2025TC008904](https://doi.org/10.1029/2025TC008904).
- Woods, K., L. M. Wallace, C. A. Williams, I. J. Hamling, S. C. Webb, Y. Ito, N. Palmer, R. Hino, S. Suzuki, M. K. Savage, E. Warren-Smith, and K. Mochizuki (2024), Spatiotemporal Evolution of Slow Slip Events at the Offshore Hikurangi Subduction Zone in 2019 Using GNSS, InSAR, and Seafloor Geodetic Data, *Journal of Geophysical Research: Solid Earth*, 129(8), e2024JB029,068, doi:[10.1029/2024JB029068](https://doi.org/10.1029/2024JB029068), eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2024JB029068>.
- Woodward, A., I. Bastow, and R. Bell (2025), Local Earthquake and Focal Mechanism Catalogue for the Northern Hikurangi Margin December 2017-October 2018 [Dataset], doi:[10.5281/ZENODO.15058318](https://doi.org/10.5281/ZENODO.15058318).
- Woodward, A., I. D. Bastow, R. Bell, L. Wallace, K. Jacobs, S. Henrys, B. Fry, T. A. J. Merry, V. Lane, L. Ville, P. Houldsworth, and L. Broadley (2026a), Seismological Characterization of Northern Hikurangi Margin Slow Slip Regions Associated With Normal Faults, Seamounts, and Seeps, *Journal of Geophysical Research*, 131.
- Woodward, A., I. Bastow, R. Bell, M. Paulatto, S. Henrys, and K. Jacobs (2026b), Local earthquake tomography of the northern Hikurangi margin: Vp and Vp/Vs models centred on Gisborne/Raukūmara Peninsula [Dataset], doi:[10.5281/zenodo.19552451](https://doi.org/10.5281/zenodo.19552451).

- Woollam, J., J. Münchmeyer, F. Tilmann, A. Rietbrock, D. Lange, T. Bornstein, T. Diehl, C. Giunchi, F. Haslinger, D. Jozinović, A. Michelini, J. Saul, and H. Soto (2022), SeisBench—A Toolbox for Machine Learning in Seismology, *Seismological Research Letters*, 93(3).
- Worthington, T. J., R. Hekinian, P. Stoffers, T. Kuhn, and F. Hauff (2006), Osbourn Trough: Structure, geochemistry and implications of a mid-Cretaceous paleosspreading ridge in the South Pacific, *Earth and Planetary Science Letters*, 245(3-4), 685–701, doi:[10.1016/j.epsl.2006.03.018](https://doi.org/10.1016/j.epsl.2006.03.018).
- Yamashita, Y., H. Yakiwara, Y. Asano, H. Shimizu, K. Uchida, S. Hirano, K. Umakoshi, H. Miyamachi, M. Nakamoto, M. Fukui, M. Kamizono, H. Kanehara, T. Yamada, M. Shinohara, and K. Obara (2015), Migrating tremor off southern Kyushu as evidence for slow slip of a shallow subduction interface, *Science*, 348(6235), 676–679, doi:[10.1126/science.aaa4242](https://doi.org/10.1126/science.aaa4242).
- Yarce, J., A. F. Sheehan, J. S. Nakai, S. Y. Schwartz, K. Mochizuki, M. K. Savage, L. M. Wallace, S. A. Henrys, S. C. Webb, Y. Ito, R. E. Abercrombie, B. Fry, H. Shaddox, and E. K. Todd (2019), Seismicity at the Northern Hikurangi Margin, New Zealand, and Investigation of the Potential Spatial and Temporal Relationships With a Shallow Slow Slip Event, *Journal of Geophysical Research: Solid Earth*, 124(5), 4751–4766, doi:[10.1029/2018JB017211](https://doi.org/10.1029/2018JB017211), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2018JB017211>.
- Yarce, J., A. Sheehan, S. Roecker, and K. Mochizuki (2021), Seismic Velocity Heterogeneity of the Hikurangi Subduction Margin, New Zealand: Elevated Pore Pressures in a Region With Repeating Slow Slip Events, *Journal of Geophysical Research: Solid Earth*, 126(5), e2020JB021605, doi:[10.1029/2020JB021605](https://doi.org/10.1029/2020JB021605), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JB021605>.
- Zal, H. J., K. Jacobs, M. K. Savage, J. Yarce, S. Mroczek, K. Graham, E. K. Todd, J. Nakai, Y. Iwasaki, A. Sheehan, K. Mochizuki, L. Wallace, S. Schwartz, S. Webb, and S. Henrys (2020), Temporal and spatial variations in seismic anisotropy and Vp/Vs ratios in a region of slow slip, *Earth and Planetary Science Letters*, 532, 115,970, doi:[10.1016/j.epsl.2019.115970](https://doi.org/10.1016/j.epsl.2019.115970).

Zhu, W. (2022), wayneweiqiang/GaMMA. [Software]. Zenodo., doi:[10.5281/zenodo.6271310](https://doi.org/10.5281/zenodo.6271310).

Zhu, W., I. W. McBrearty, S. M. Mousavi, W. L. Ellsworth, and G. C. Beroza (2022), Earthquake Phase Association Using a Bayesian Gaussian Mixture Model, *Journal of Geophysical Research: Solid Earth*, 127(5), e2021JB023,249, doi:[10.1029/2021JB023249](https://doi.org/10.1029/2021JB023249), eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021JB023249>.

CHAPTER A

Seismograph networks

Table A.1: Temporary seismograph network.

Net	Stat.	Lat.	Lon.	Elev. (m)	Start	End
3C	BC12	-38.55108	178.19759	18	2017:353:00:00:00	2018:042:23:06:01
3C	BC30	-38.47982	178.08534	136	2017:353:00:00:00	2018:041:21:37:54
3C	BS01	-38.58737	178.18833	15	2017:351:00:00:00	2018:043:04:11:21
3C	BS03	-38.56257	178.15494	139	2017:351:00:00:00	2018:042:02:14:10
3C	BS04	-38.55215	178.14247	181	2017:354:00:00:00	2018:043:01:59:10
3C	BS05	-38.54805	178.11816	242	2017:353:00:00:00	2018:042:21:37:55
3C	BS06	-38.53010	178.09573	138	2017:351:00:00:00	2018:038:01:05:26
3C	BS07	-38.52748	178.07580	208	2017:351:00:00:00	2018:037:23:49:59
3C	BS08	-38.52083	178.04708	244	2017:352:00:00:00	2018:037:21:11:45
3C	BS09	-38.50833	178.03052	205	2017:349:00:00:00	2018:044:02:48:48
3C	BS10	-38.49756	178.01841	340	2017:351:00:00:00	2018:044:00:59:50
3C	BS11	-38.48998	177.99483	320	2017:351:00:00:00	2018:043:22:33:20
3C	BS12	-38.47780	177.97403	211	2017:351:00:00:00	2018:043:20:41:55
3C	BS13	-38.56394	178.19807	88	2017:351:00:00:00	2018:043:03:00:51
3C	BS14	-38.55419	178.18089	182	2017:351:00:00:00	2018:043:00:30:16
3C	BS15	-38.55647	178.17201	153	2017:356:00:00:00	2018:042:22:50:00
3C	BS16	-38.54392	178.13487	213	2017:349:00:00:00	2018:012:23:57:15
3C	BS17	-38.53753	178.12382	220	2017:353:00:00:00	2018:042:23:32:24
3C	BS18	-38.51613	178.10000	271	2017:352:00:00:00	2018:038:02:00:10

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Table A.1 – continued from previous page

Net	Stat.	Lat.	Lon.	Elev. (m)	Start	End
3C	BS19	-38.51527	178.08644	178	2017:354:00:00:00	2018:038:02:50:46
3C	BS20	-38.49480	178.06113	199	2017:355:00:00:00	2018:038:21:34:43
3C	BS21	-38.50428	178.05014	116	2017:354:00:00:00	2018:043:02:13:12
3C	BS25	-38.54947	178.20995	3	2017:351:00:00:00	2018:043:01:07:03
3C	BS26	-38.53780	178.19572	39	2017:354:00:00:00	2018:042:21:59:05
3C	BS27	-38.52998	178.17114	234	2017:353:00:00:00	2018:044:02:25:46
3C	BS28	-38.52214	178.16051	247	2017:353:00:00:00	2018:043:23:00:20
3C	BS30	-38.50770	178.12067	240	2018:002:00:00:00	2018:041:22:30:59
3C	BS31	-38.49900	178.09589	168	2018:044:01:39:46	2018:281:23:59:59
3C	BS32	-38.49172	178.07451	183	2018:044:03:30:38	2018:279:23:59:59
3C	BS34	-38.46960	178.03052	380	2018:038:20:09:14	2018:280:23:59:59
3C	BS37	-38.53733	178.22762	51	2018:041:02:18:40	2018:279:23:59:59
3C	BS38	-38.52661	178.20808	151	2018:040:22:42:44	2018:279:23:59:59
3C	BS39	-38.51959	178.18650	80	2018:043:04:22:22	2018:281:23:59:59
3C	BS40	-38.51168	178.17030	186	2018:043:21:32:09	2018:282:23:59:59
3C	BS41	-38.51057	178.15138	212	2018:043:22:12:43	2018:281:23:59:59
3C	BS42	-38.48893	178.12537	220	2018:043:00:00:00	2018:280:23:59:59
3C	BS43	-38.47933	178.10652	201	2018:041:22:29:40	2018:215:23:59:59
3C	BS44	-38.47351	178.09042	173	2018:043:00:00:00	2018:284:23:59:59
3C	BS46	-38.45290	178.04890	411	2017:009:00:00:00	2018:043:23:49:59
3C	BS47	-38.43790	178.03766	449	2017:351:00:00:00	2018:042:23:47:06
3C	BS49	-38.51994	178.23625	100	2017:352:00:00:00	2018:040:21:34:03
3C	BS50	-38.51625	178.22719	100	2017:351:00:00:00	2018:040:22:53:20
3C	BS51	-38.50270	178.19779	132	2017:354:00:00:00	2018:041:02:46:02
3C	BS52	-38.49687	178.17213	133	2017:356:00:00:00	2018:042:23:48:13
3C	BS53	-38.48383	178.15887	100	2017:353:00:00:00	2018:042:21:06:49
3C	BS54	-38.48363	178.14136	178	2017:352:00:00:00	2018:042:02:36:18

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Table A.1 – continued from previous page

Net	Stat.	Lat.	Lon.	Elev. (m)	Start	End
3C	BS55	-38.46680	178.11298	324	2017:353:00:00:00	2018:043:22:37:31
3C	BS59	-38.43191	178.04747	539	2017:352:00:00:00	2018:039:00:18:03
3C	BS60	-38.42132	178.02083	434	2018:043:00:00:00	2018:280:23:59:59

Table A.2: Permanent seismograph network.

Net	Stat.	Lat.	Lon.	Elev. (m)	Start	Instrument
NZ	CNGZ	-38.48372	178.20747	179	2006-07-21	Short period
NZ	HAZ	-37.75610	177.78260	81	2010-01-14	Broadband
NZ	KNZ	-39.02176	177.67367	32	1998-10-19	Broadband
NZ	MHGZ	-39.15256	177.90700	302	2006-05-04	Short period
NZ	MWZ	-38.33400	177.52778	573	2004-02-29	Broadband
NZ	MXZ	-37.56226	178.30663	106	2004-02-29	Broadband
NZ	PRGZ	-38.92259	177.88328	503	2006-07-23	Short period
NZ	PUZ	-38.07155	178.25721	430	1989-03-08	Broadband
NZ	RAGZ	-38.49560	177.41540	947	2009-06-10	Short period
NZ	RIGZ	-38.70580	177.76220	621	2009-06-08	Short period
NZ	RUGZ	-37.96513	177.67748	1175	2011-05-19	Short period
NZ	TKGZ	-38.43920	177.84850	157	2010-03-13	Short period
NZ	TWGZ	-38.17670	177.98040	687	2009-12-09	Short period

CHAPTER B

Station magnitude corrections

Table B.1: Station code and local magnitude corrections. Stations from the temporary network (those with codes beginning with BC or BS) are calculated as in Chapter 3. Stations from the permanent network are from *Rhoades et al. (2020)*.

Station	Correction (M_L)	Station	Correction (M_L)	Station	Correction (M_L)
BC12	0.09	BS39	0.09	BS27	-0.23
BC30	0.02	BS40	0.11	BS28	0.1
BS01	0.13	BS41	0.21	BS30	0.06
BS03	-0.01	BS42	-0.38	BS31	0.06
BS04	-0.12	BS43	0.04	BS32	-0.02
BS05	0.02	MXZ	-0.109	BS34	0.04
BS06	0.16	BS46	-0.11	BS37	-0.13
BS07	0.11	BS47	0.05	BS38	0.08
BS08	0.0	BS49	0.18	PRGZ	0.036
BS09	0.26	BS50	0.04	PUZ	0.054
BS11	-0.07	BS51	0.12	RAGZ	0.132
BS12	0.35	BS52	0.05	RIGZ	0.164
BS13	0.25	BS53	0.13	RUGZ	-0.041
BS14	0.04	BS54	0.16	TKGZ	0.151
BS15	0.14	BS55	0.11	TWGZ	0.253
BS16	-0.13	BS59	0.21	BS20	0.24
BS17	0.06	BS60	-0.05	BS21	0.31
BS18	0.13	CNGZ	-0.116	BS25	0.03
BS19	-0.09	HAZ	0.0003	BS26	0.13
KNZ	0.0378	MHGZ	-0.073	MWZ	-0.053

CHAPTER C

1-D velocity model

Table C.1: 1-D velocity model derived in this study.

Depth (km)	V_P (km/s)
-2.0	2.44
0.0	3.47
1.0	4.24
2.0	4.33
3.0	4.49
4.0	4.50
5.0	4.64
6.0	4.85
7.0	4.88
8.0	5.02
9.0	5.10
10.0	5.04
11.0	5.06
12.0	5.00
14.0	5.33
16.0	5.58
20.0	6.16
24.0	6.96
26.0	7.13
Continued on next page	

Table C.1 – continued from previous page

Depth (km)	V_P (km/s)
28.0	7.47
30.0	7.54
32.0	7.90
36.0	8.04
40.0	8.14
44.0	8.34
48.0	8.55

CHAPTER D

Time-dependent slow slip inversions and seismicity

The time-dependent slow slip inversions were completed by Laura Wallace during a research visit to GEOMAR. The following text is part of [Woodward et al. \(2026a\)](#). These models are used for jointly analysing seismicity and SSEs in chapter 4.

To detect and characterize SSEs that occurred immediately prior to, during, and after our dense seismic deployment, we undertook a time-dependent inversion of GNSS time series for transient SSEs. The non-linear, time-dependent inversion software TDEFNODE ([McCaffrey, 2009](#)) was used to invert the GNSS time series for transient slip sources characterizing SSE sequences during the late 2017 period through to 2019. TDEFNODE uses an elastic block modelling approach, which subdivides the region into rotating, elastic tectonic blocks, bounded by known faults. Our TDEFNODE model set-up and inversion approach for the Hikurangi margin was identical to that of ([Woods et al., 2024](#)), where further details can be found. We fixed our Euler poles for each block to those derived from ([Wallace et al., 2012b](#)), and the relative motion between the rotating blocks was used to fix the rake of slip on the subduction plate boundary (to reduce the number of free parameters). The subduction interface geometry of [Williams et al. \(2013\)](#) was used to define the plate interface, on which the SSE transients were assumed to occur.

D.1 Ten-day time slices

The time-dependent slip measurements were binned into ten-day intervals. The following figures show the ten-day slip areas during the SSEs, and the seismicity that occurred during them.

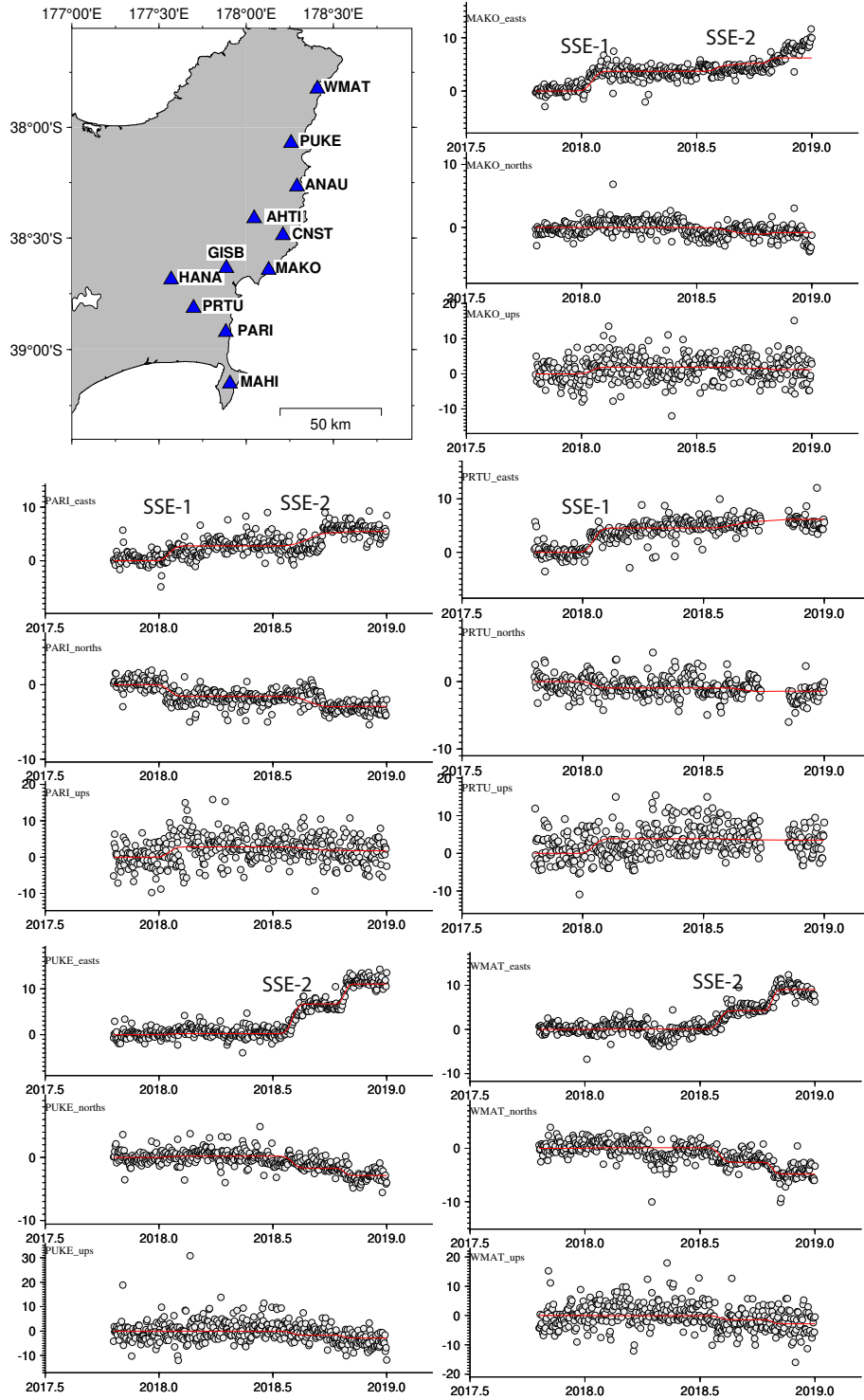


Figure D.1: Inset map: GNSS stations (blue triangles) shown in this figure and the next figure. For each station (labelled top right in each graph), the east, north, and vertical components of the GNSS stations are shown. Grey circles: daily displacement data. Red line: time-dependent slow slip model determined using TDEFNODE. Where there are clear examples, SSE-1 and SSE-2 are labelled.

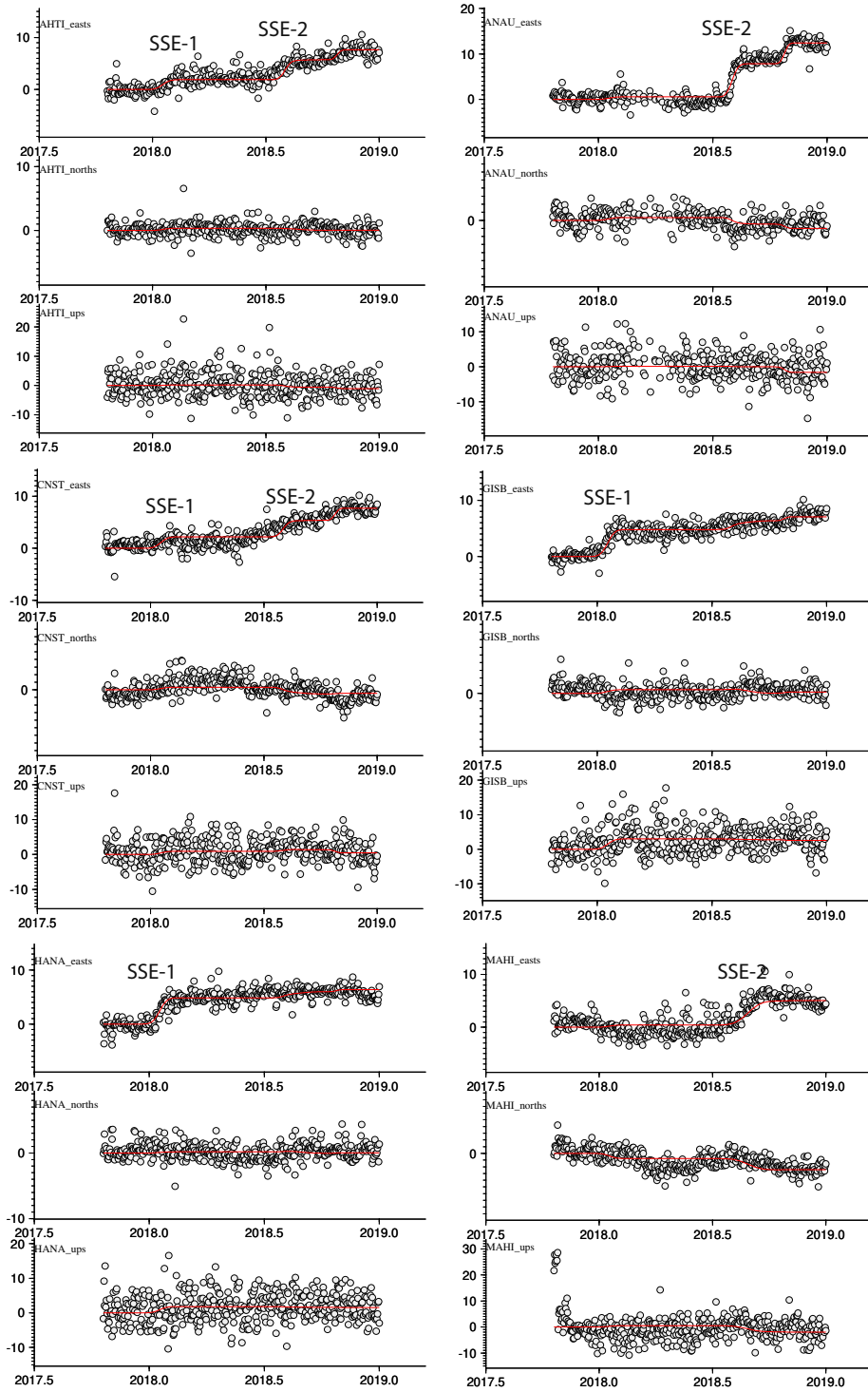


Figure D.2: For each station (locations shown in previous figure), the east, north, and vertical components of the GNSS stations are shown. Grey circles: daily displacement data. Red line: time-dependent slow slip model determined using TDEFNODE. Where there are clear examples, SSE-1 and SSE-2 are labelled.

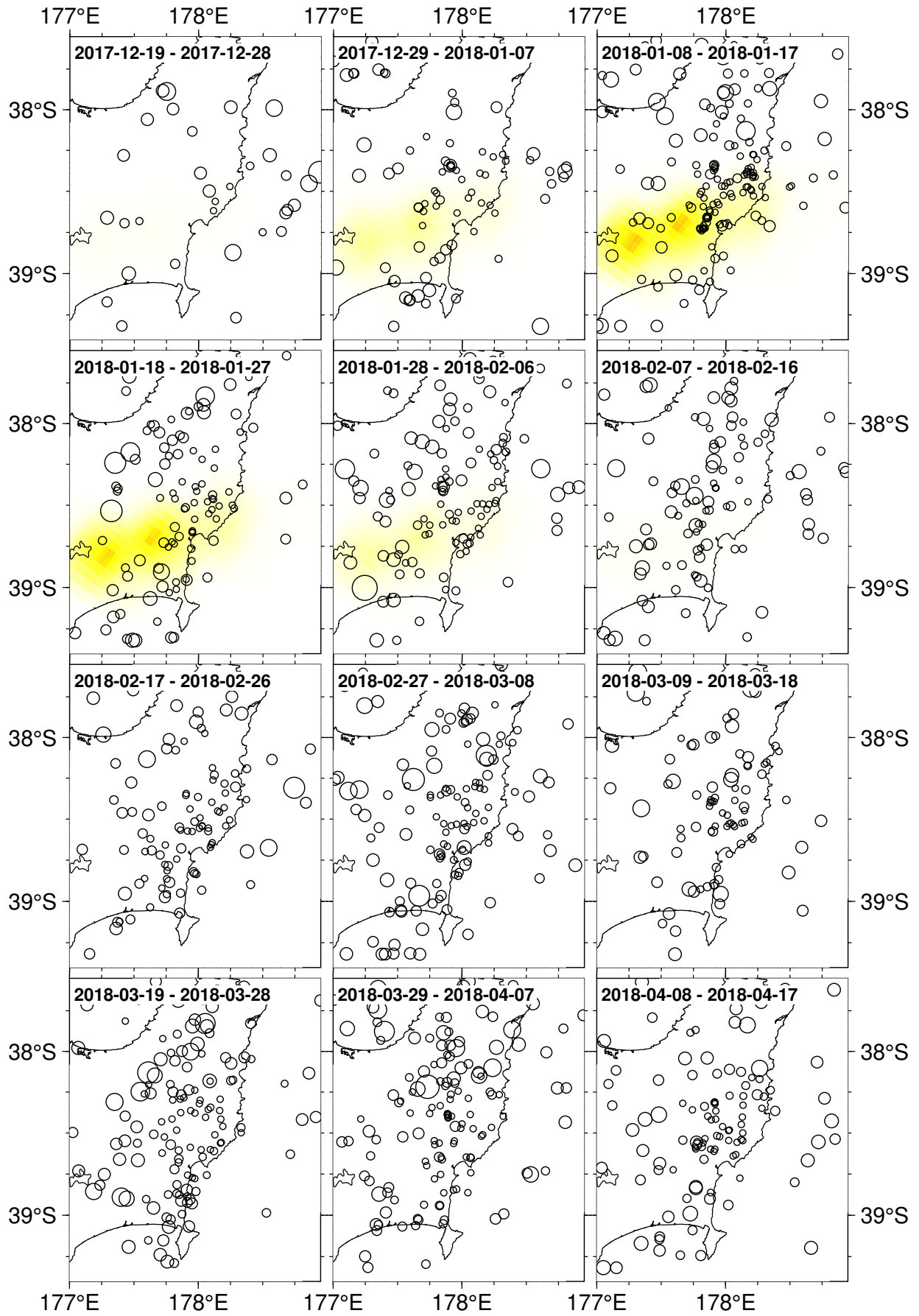


Figure D.3: Seismicity (circles) and slow slip distributions in ten-day windows, for dates as shown in each panel.

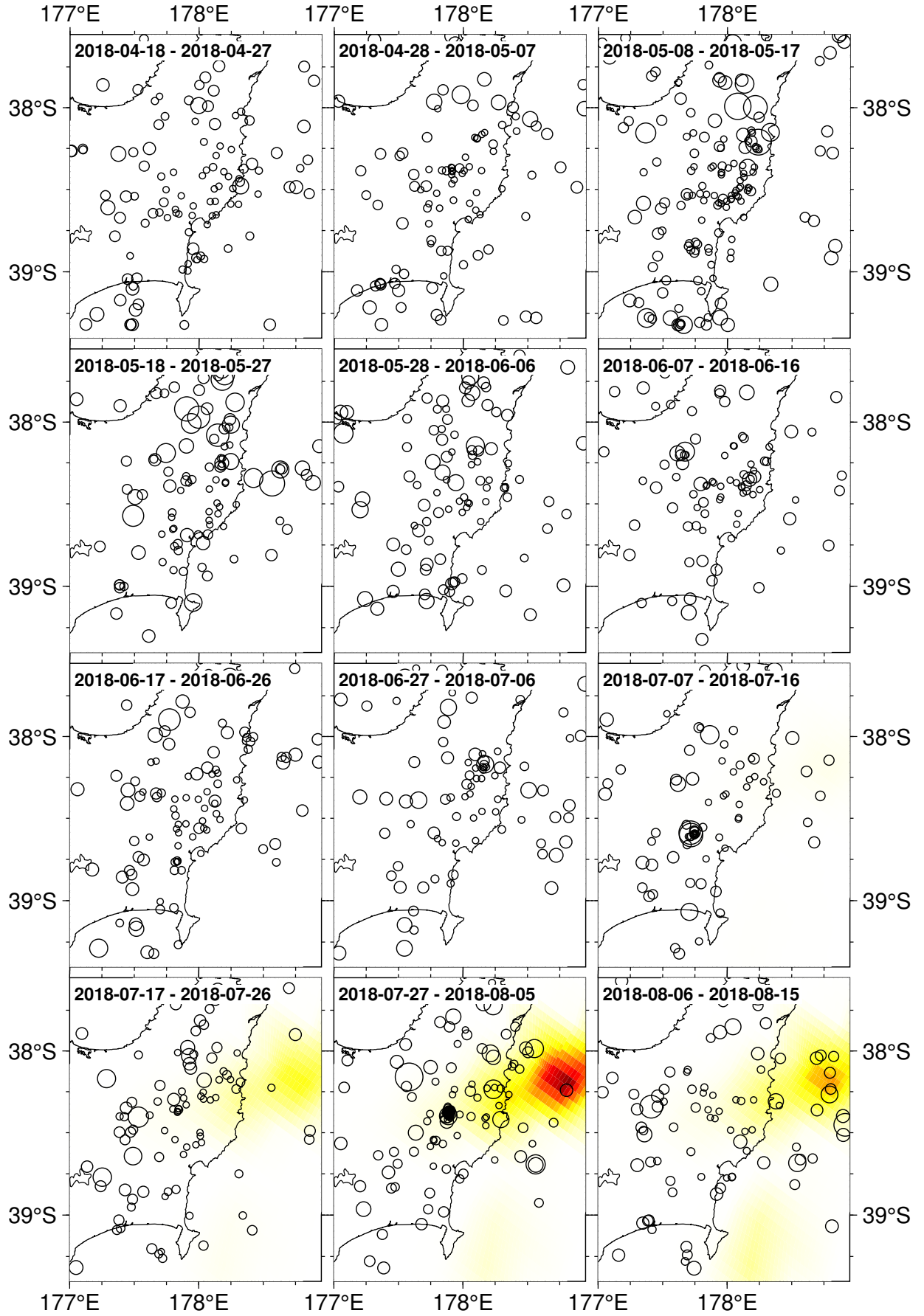


Figure D.4: Seismicity (circles) and slow slip distributions in ten-day windows, for dates as shown in each panel.

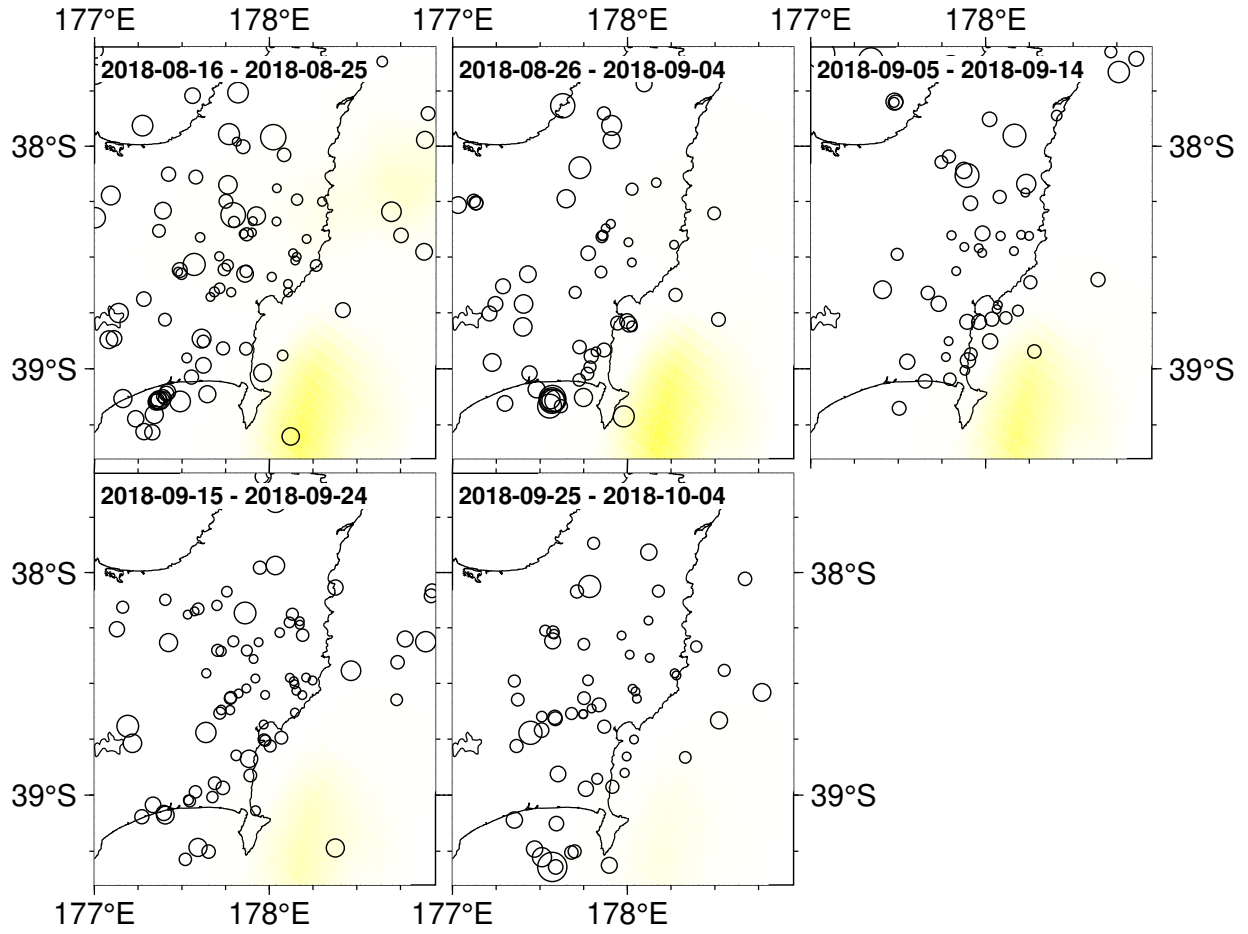


Figure D.5: Seismicity (circles) and slow slip distributions in ten-day windows, for dates as shown in each panel.

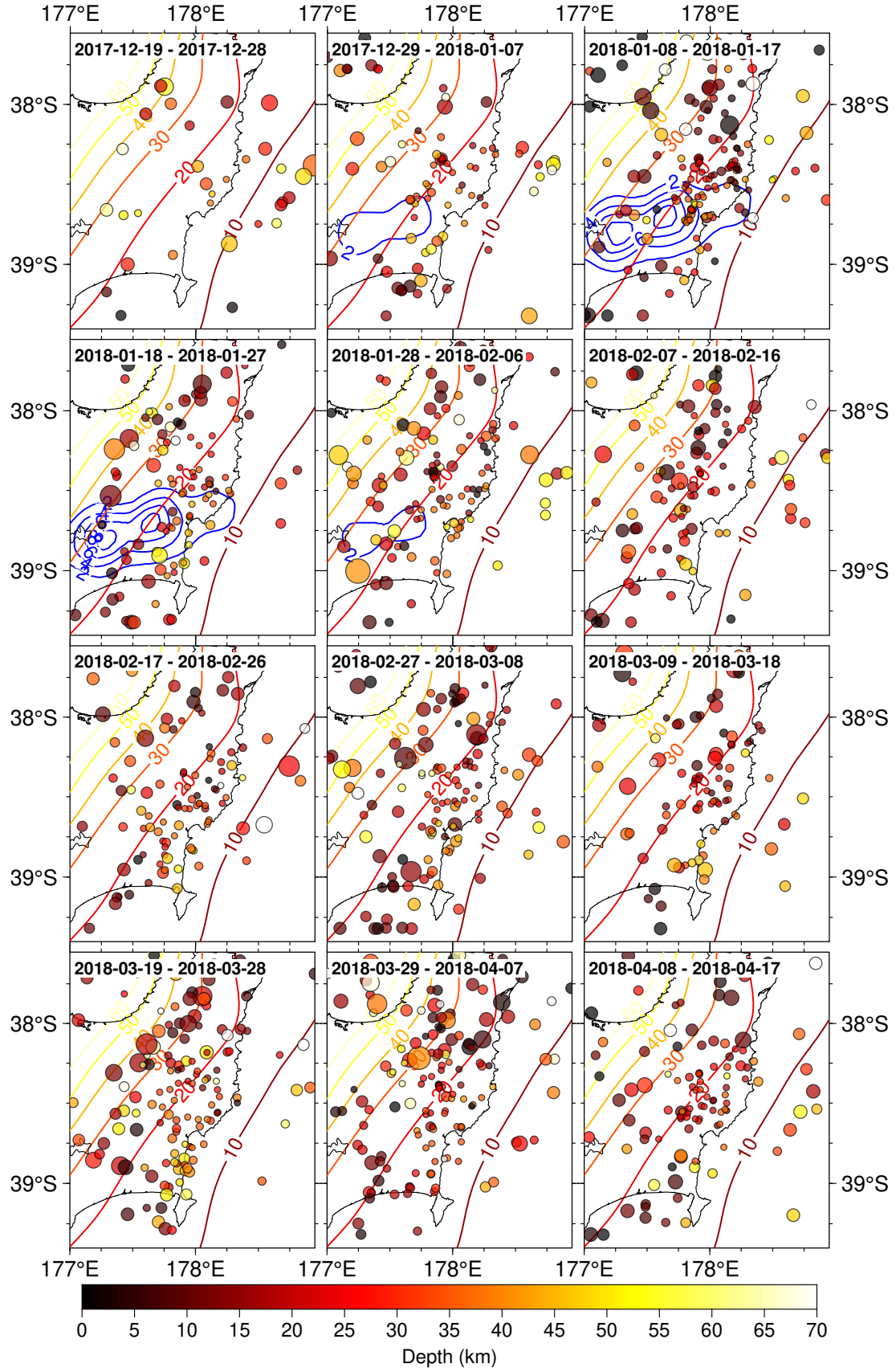


Figure D.6: Seismicity distributions in ten-day windows plotted by depth. Blue contours: SSE displacement (mm). Contours: plate interface depth (*Williams et al., 2013*).

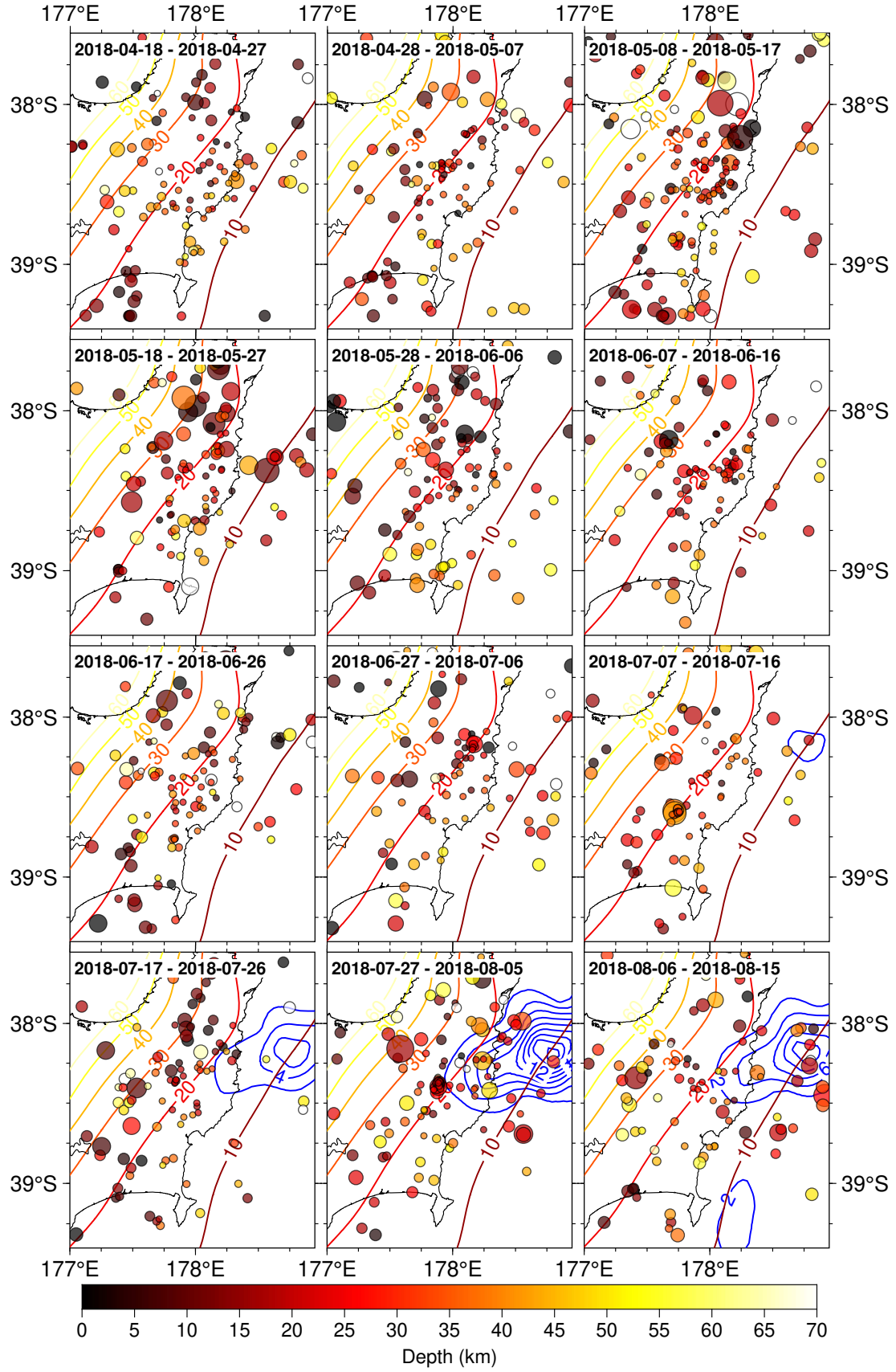


Figure D.7: Seismicity distributions in ten-day windows plotted by depth. Blue contours: SSE displacement (mm). Contours: plate interface depth (*Williams et al., 2013*).

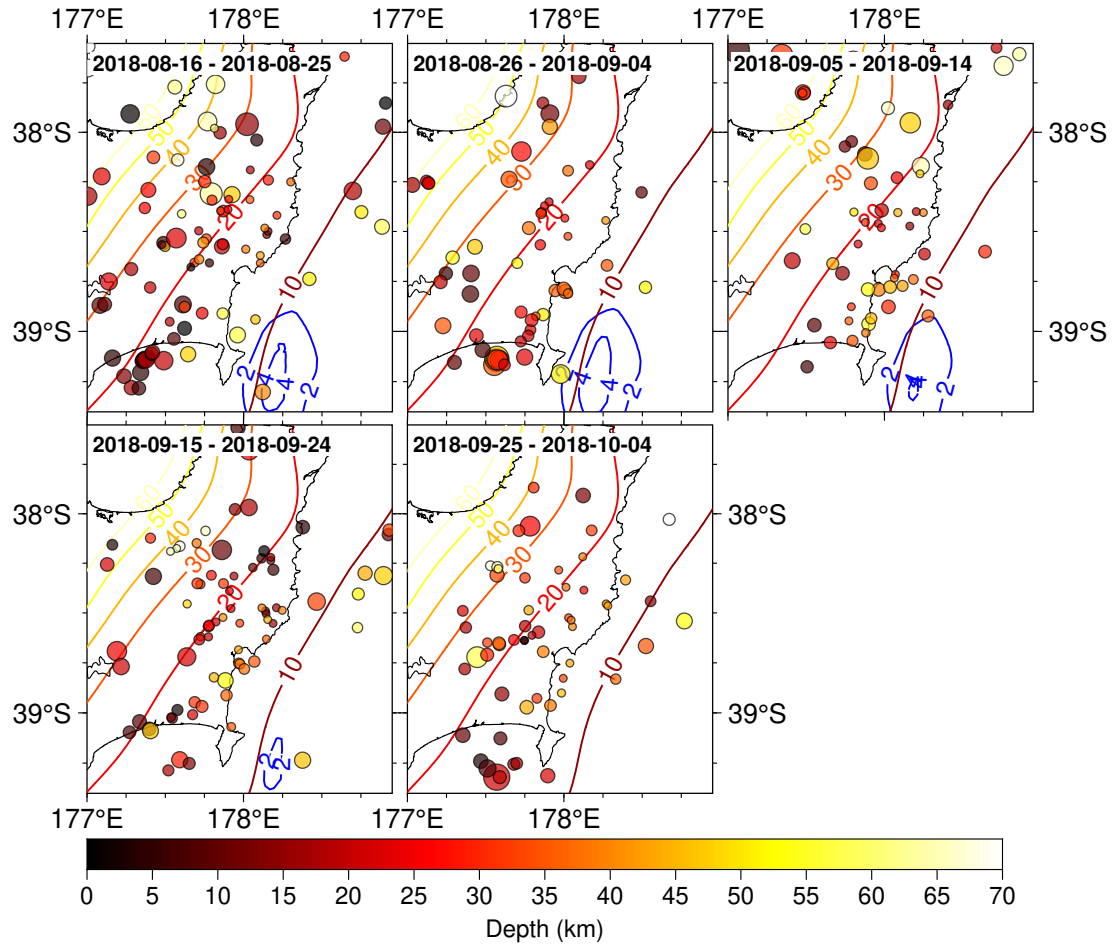


Figure D.8: Seismicity distributions in ten-day windows plotted by depth. Blue contours: SSE displacement (mm). Contours: plate interface depth (*Williams et al., 2013*).

CHAPTER E

Tomography results

Here we present further figures, showing tomograms, spread functions, and checkerboard recovery, as discussed in Chapter 5. The following figures are shown:

- Composite figures of final models, spread function results, and checkerboard recovery for additional depth slices.
- Final models in map view shown as absolute V_P , percentage variation in V_P relative to the mean V_P at each depth, and V_P/V_S .
- Spread function results in cross-section for V_P and V_P/V_S .
- Coarse checkerboards (15 km anomalies) in cross-section for V_P and V_P/V_S .
- Finer checkerboards (10 km anomalies) in cross-section for V_P and V_P/V_S .
- De-clustered earthquake catalogue tomography results.

For each composite depth slice, figure captions are as follows:

Caption for (Figures E.1 – E.3). **a)** Final V_P model at given depth. Black solid line: interpretable region defined by the spread function (contoured in geographic coordinates). Black dashed line: intersection of plate interface. **b)** Spread function for V_P model. Circles: spread function (SF) at each node. Polygons: 70 per cent contour of resolution kernel. Black solid line: interpretable region defined by SF = 2.0 (contoured in Cartesian coordinates). Crosses: fixed nodes. **c)** Final V_P/V_S model at given depth. Black solid line: interpretable region defined by the spread function (contoured in geographic coordinates). Black dashed line: intersection of plate interface. **d)** Spread function for V_P/V_S model. Circles: spread function (SF) at each node. Polygons: 70 per cent contour of resolution kernel. Black solid

line: interpretable region defined by $SF = 2.2$ (contoured in Cartesian coordinates). Crosses: fixed nodes. **e)** Input checkerboard to calculate synthetic travel-times. **f)** 15 km anomaly checkerboard recovery for V_P from a 1-D starting model. Open circles: node positions. **g)** 15 km anomaly checkerboard recovery for V_P/V_S from a 1-D starting model. Open circles: node positions.

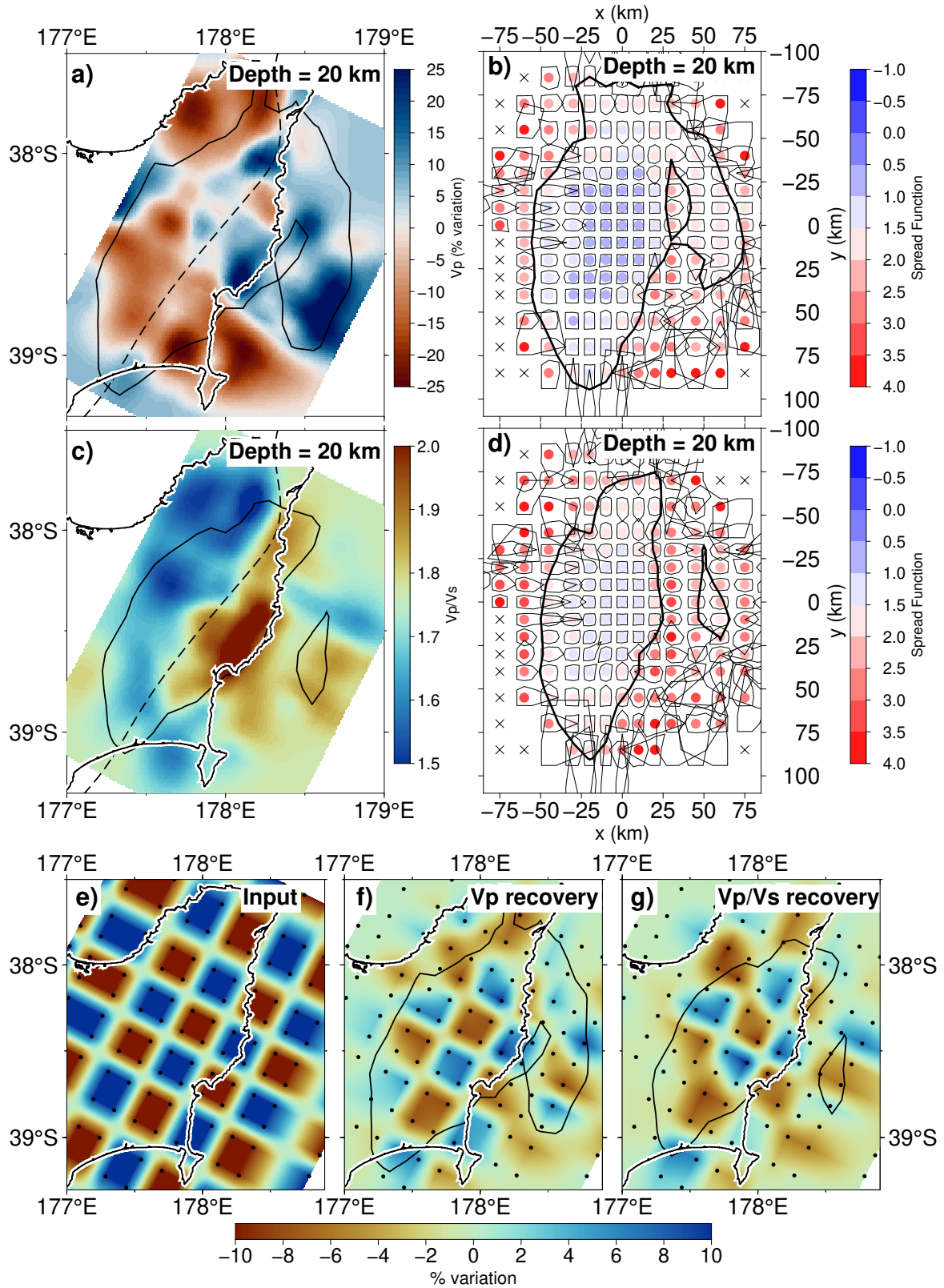


Figure E.1: Summary of tomograms, spread functions, and checkerboard recovery at 20 km depth. Caption as above.

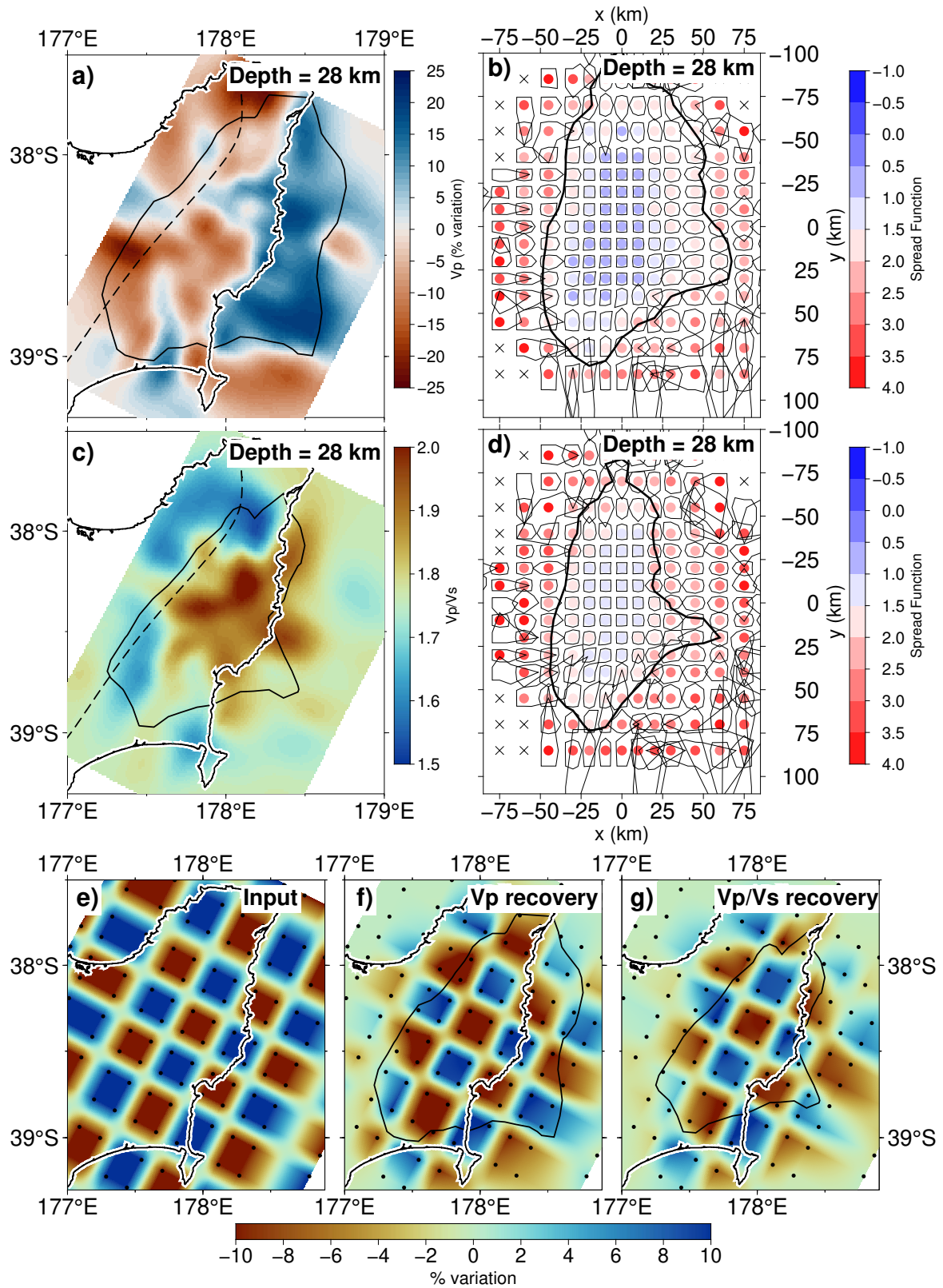


Figure E.2: Summary of tomograms, spread functions, and checkerboard recovery at 28 km depth. Caption as above.

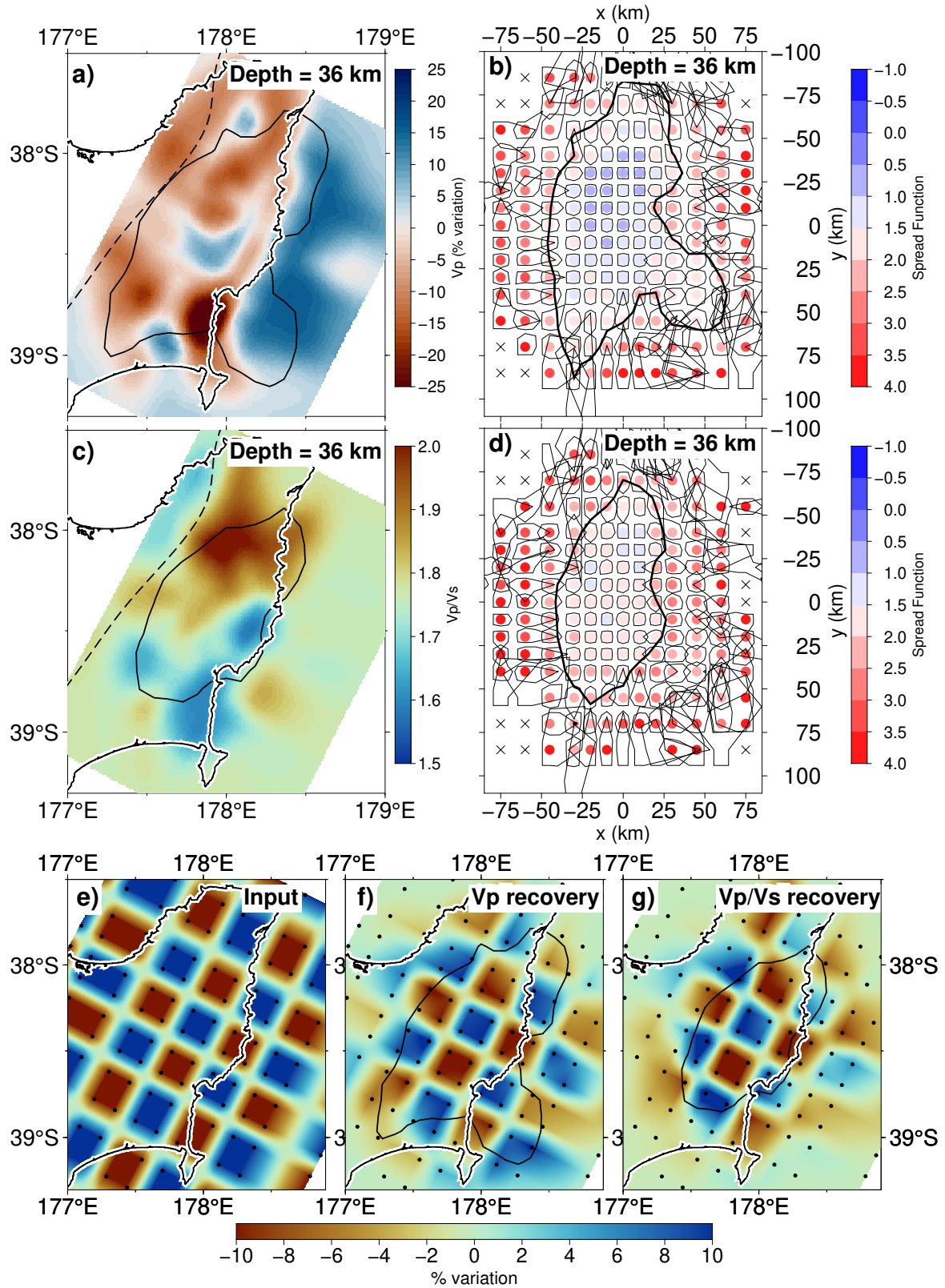


Figure E.3: Summary of tomograms, spread functions, and checkerboard recovery at 36 km depth. Caption as above.

Final models in map view

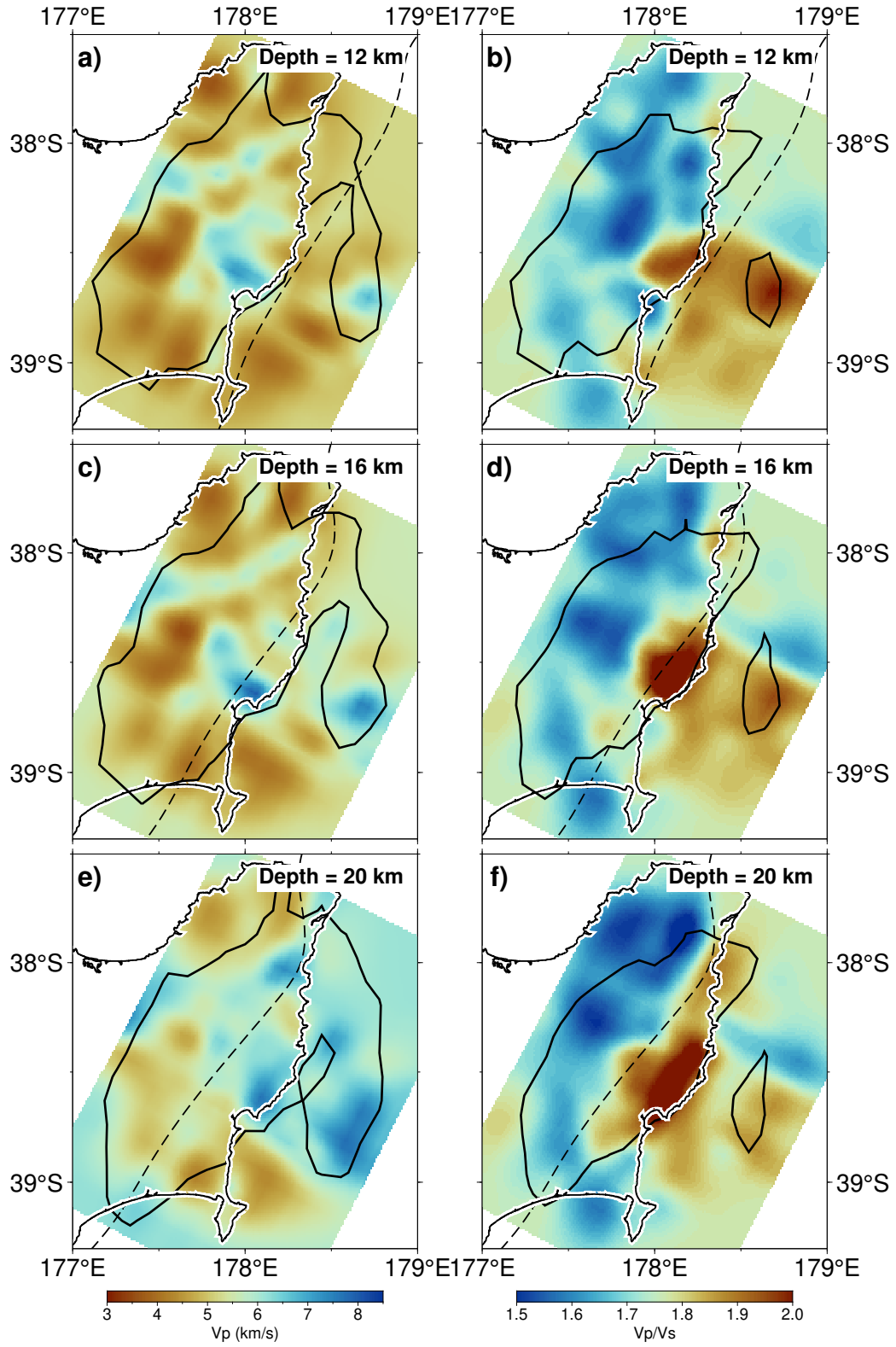


Figure E.4: Absolute V_P and V_P/V_S models for 12, 16, and 20 km depth.

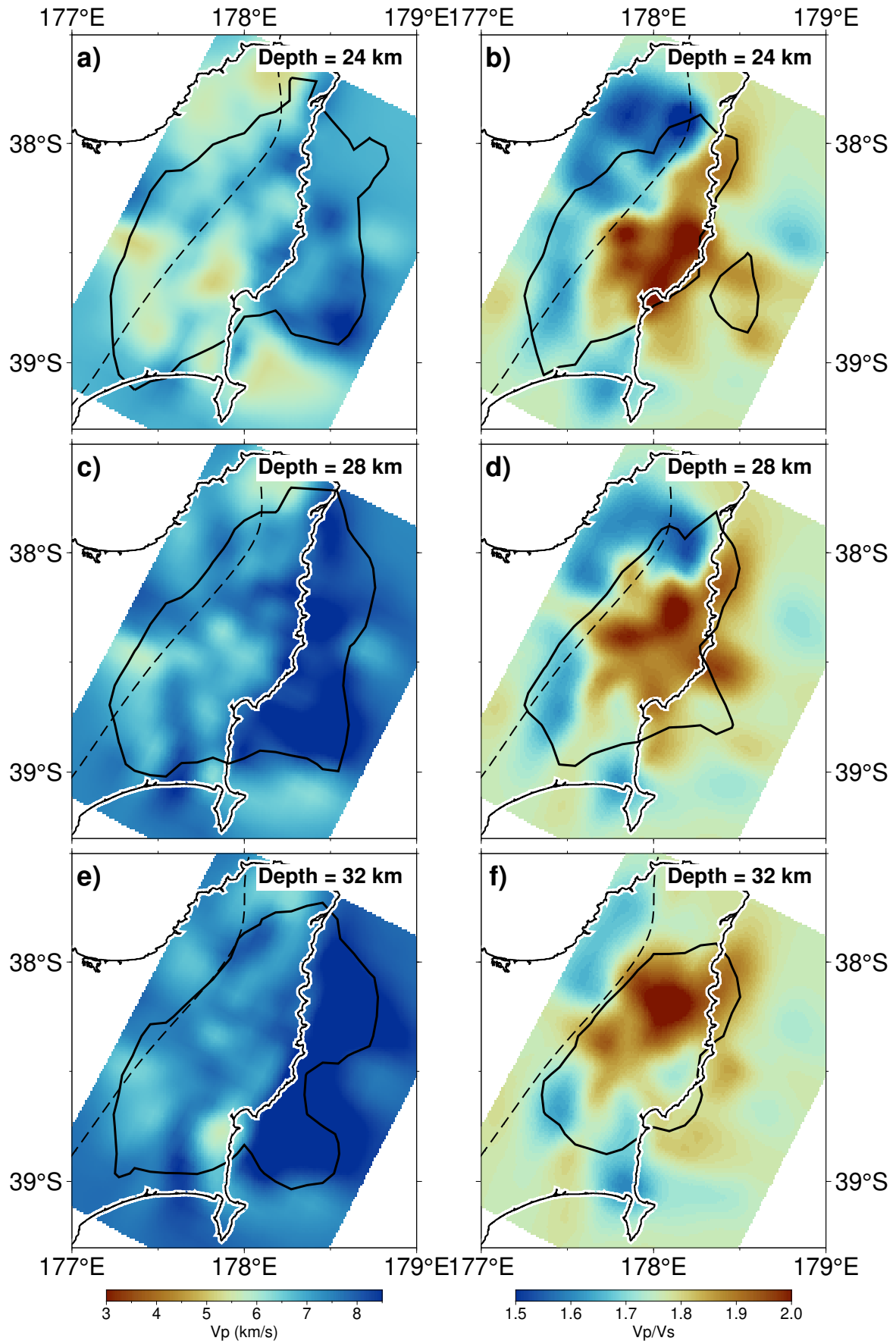


Figure E.5: Absolute V_P and V_P/V_S models for 24, 28, and 32 km depth.

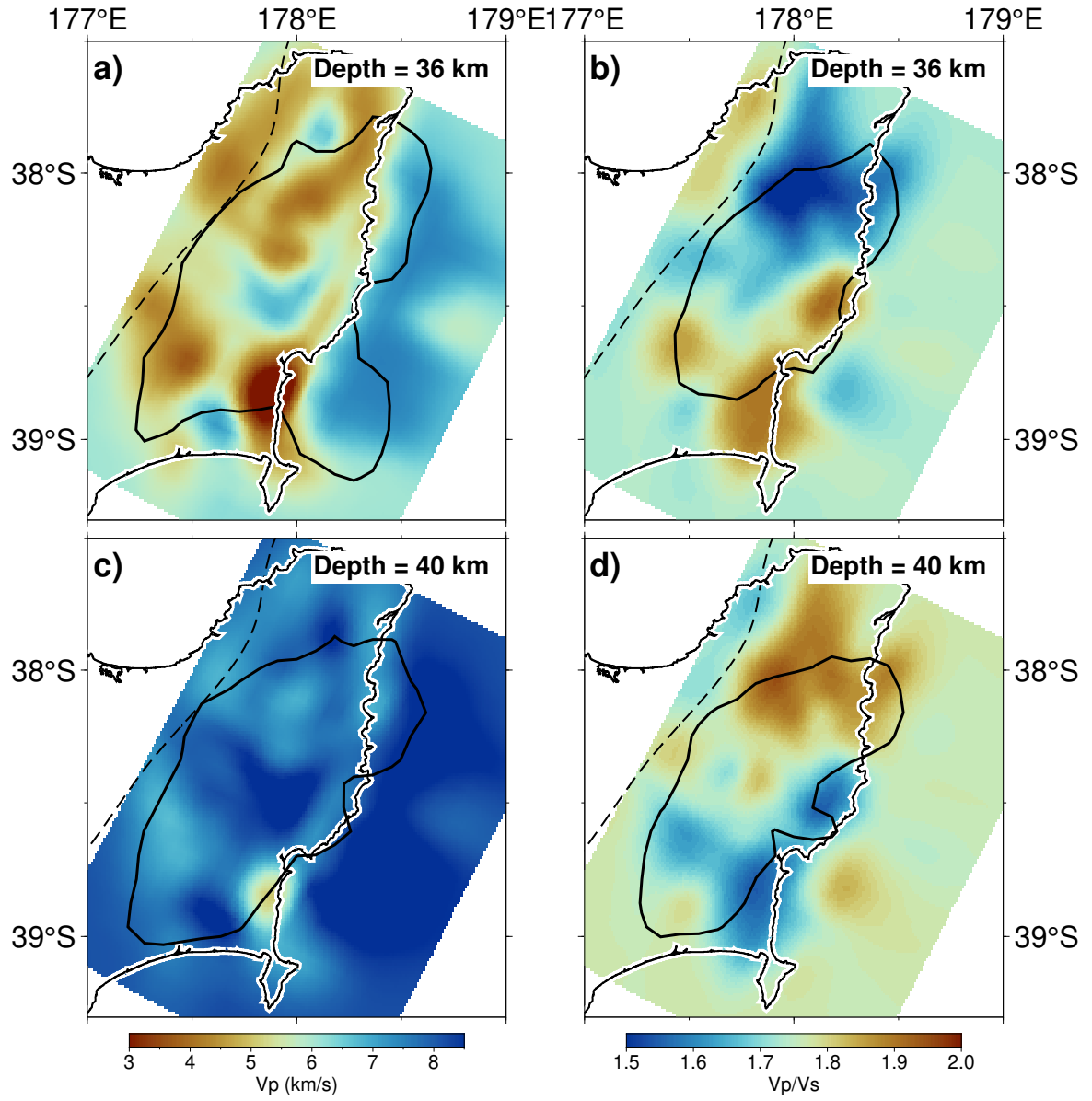


Figure E.6: Absolute V_P and V_P/V_S models for 36 and 40 km depth.

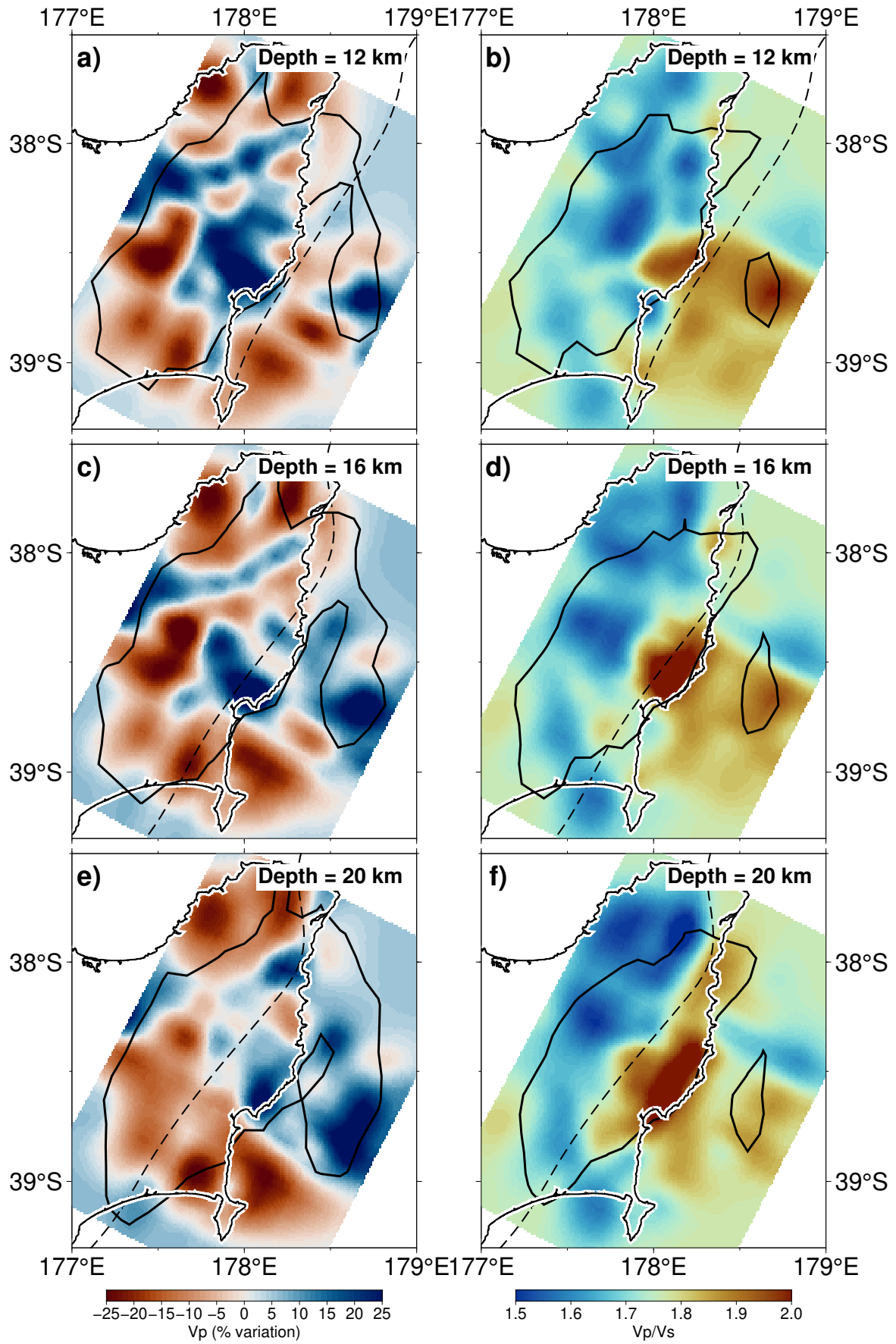


Figure E.7: Relative V_P and V_P/V_S models for 12, 16, and 20 km depth.

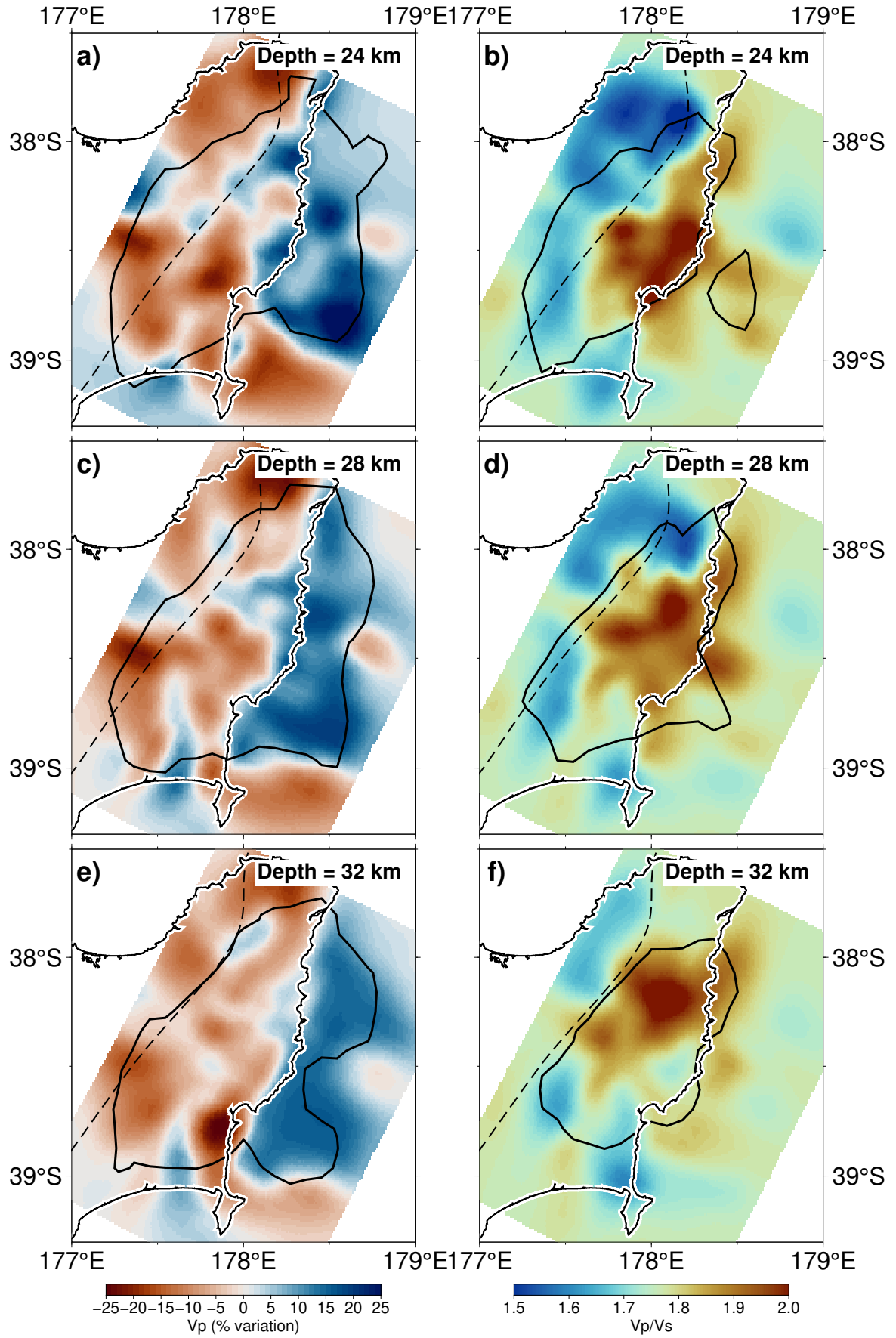


Figure E.8: Relative V_P and V_P/V_S models for 24, 28, and 32 km depth.

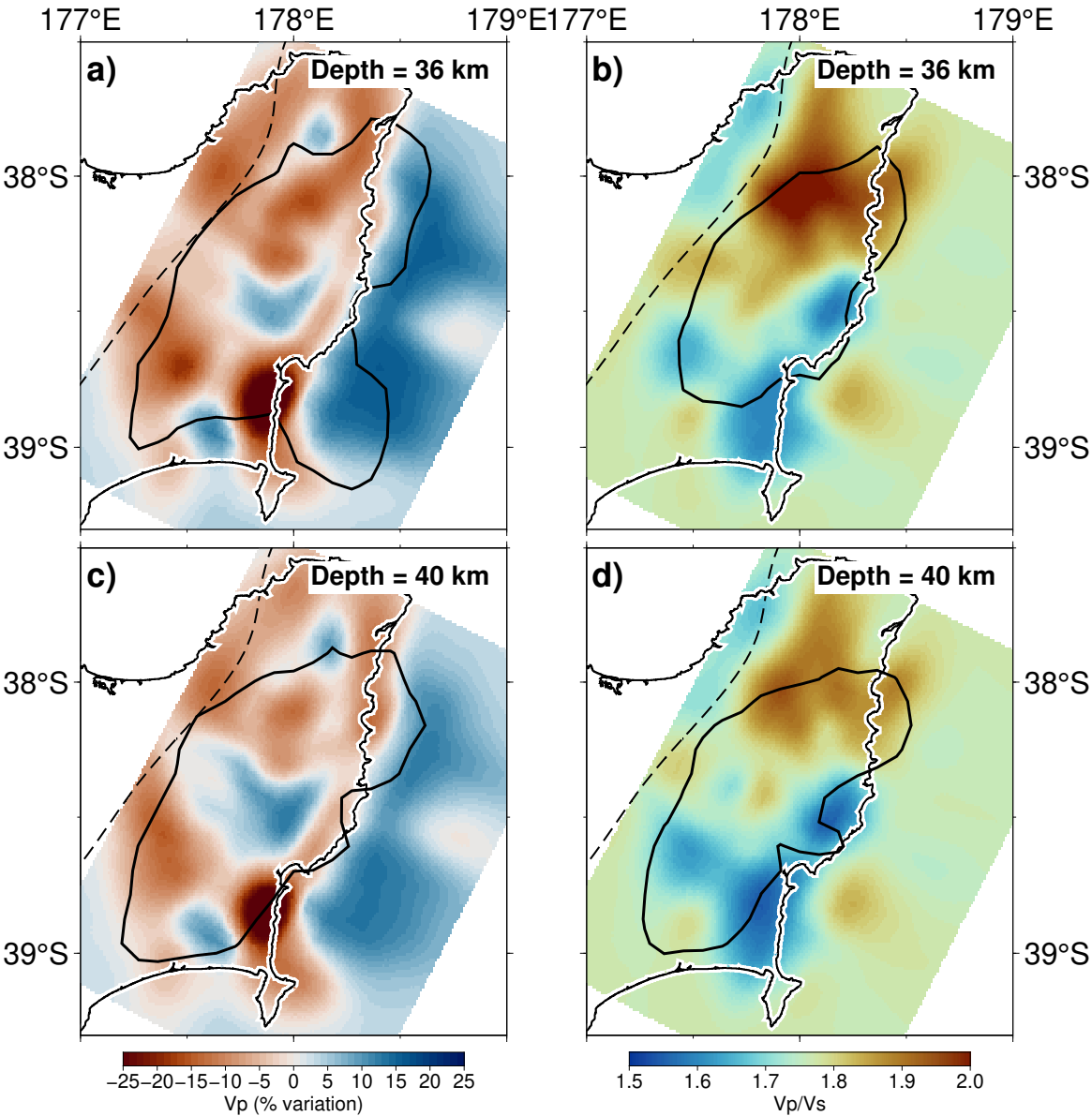


Figure E.9: Relative V_P and V_P/V_S models for 36 and 40 km depth.

Spread functions

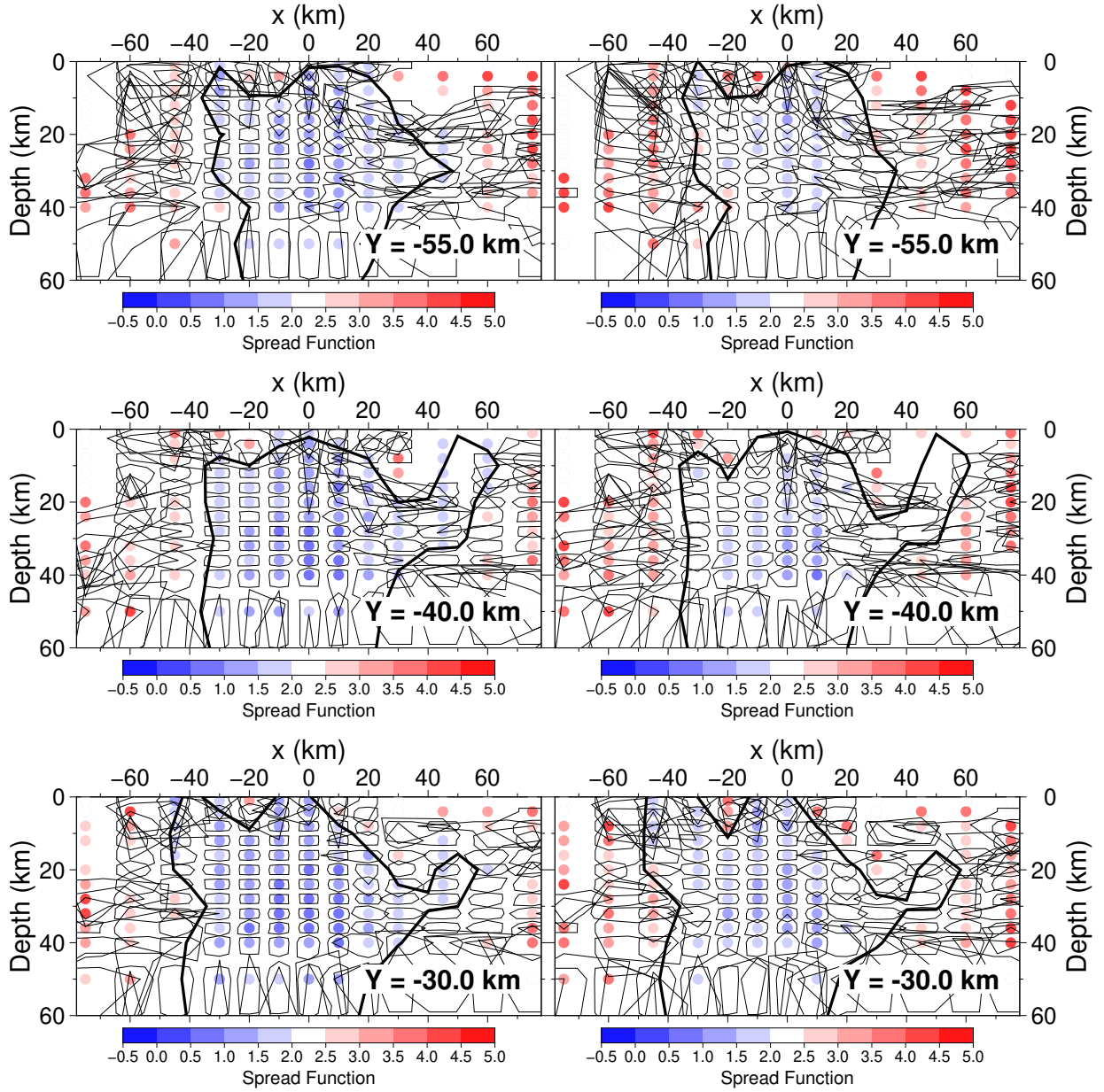


Figure E.10: Left) Spread function for V_P model in cross-section at $Y = -55$ km, $Y = -40$ km, and $Y = -30$ km. Circles: spread function (SF) at each node. Polygons: 70 per cent contour of resolution kernel. Black solid line: interpretable region defined by $SF = 2.0$ (contoured in Cartesian coordinates). Right) Spread function for V_P/V_S model. Circles: spread function (SF) at each node. Polygons: 70 per cent contour of resolution kernel. Black solid line: interpretable region defined by $SF = 2.2$ (contoured in Cartesian coordinates).

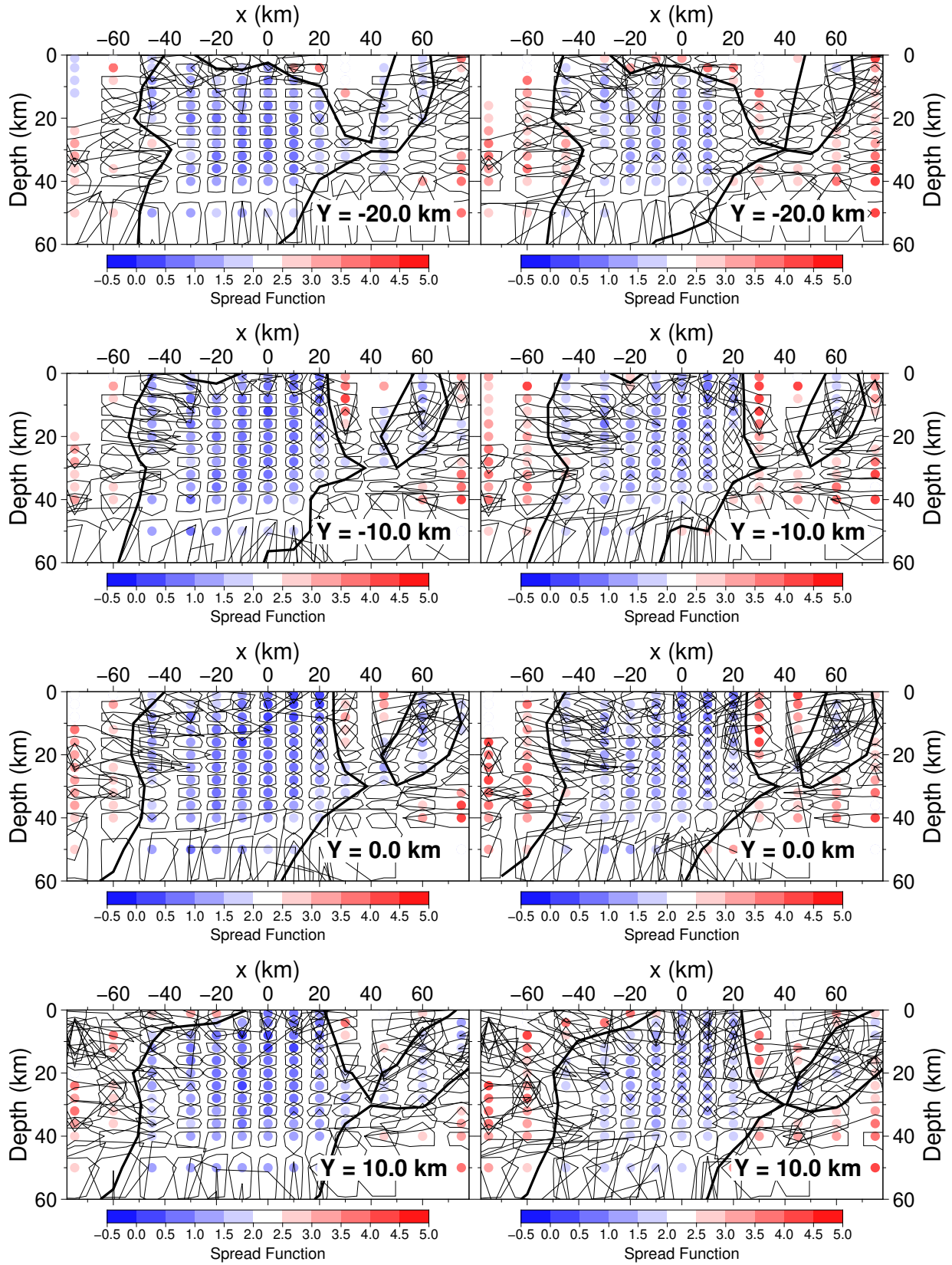


Figure E.11: As in Figure E.10 for $Y = -20$ km, $Y = -10$ km, $Y = 0$ km, and $Y = 10$ km.

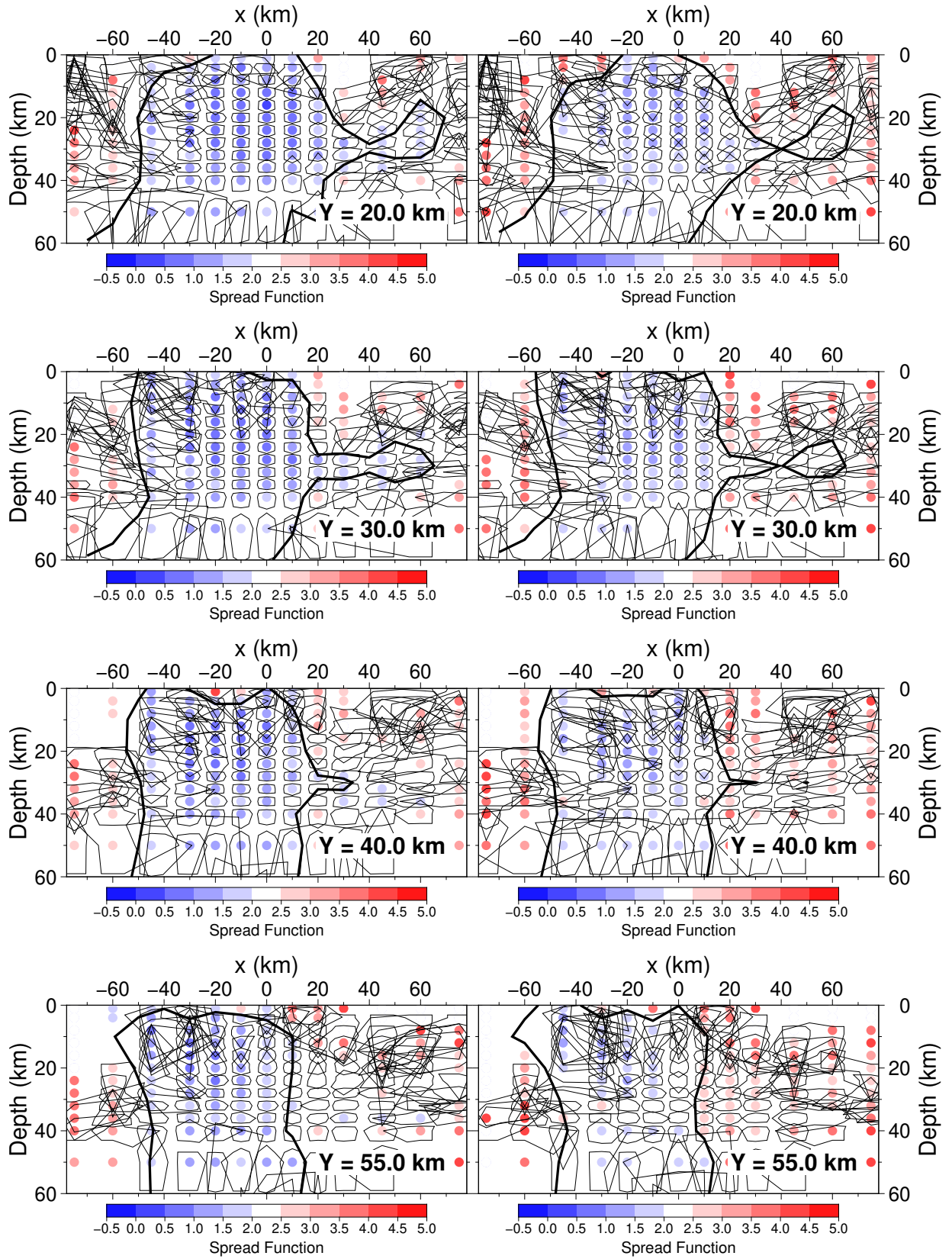


Figure E.12: As in Figure E.10 for $Y=20$ km, $Y=30$ km, $Y=40$ km and $Y=55$ km.

Coarse checkerboard tests

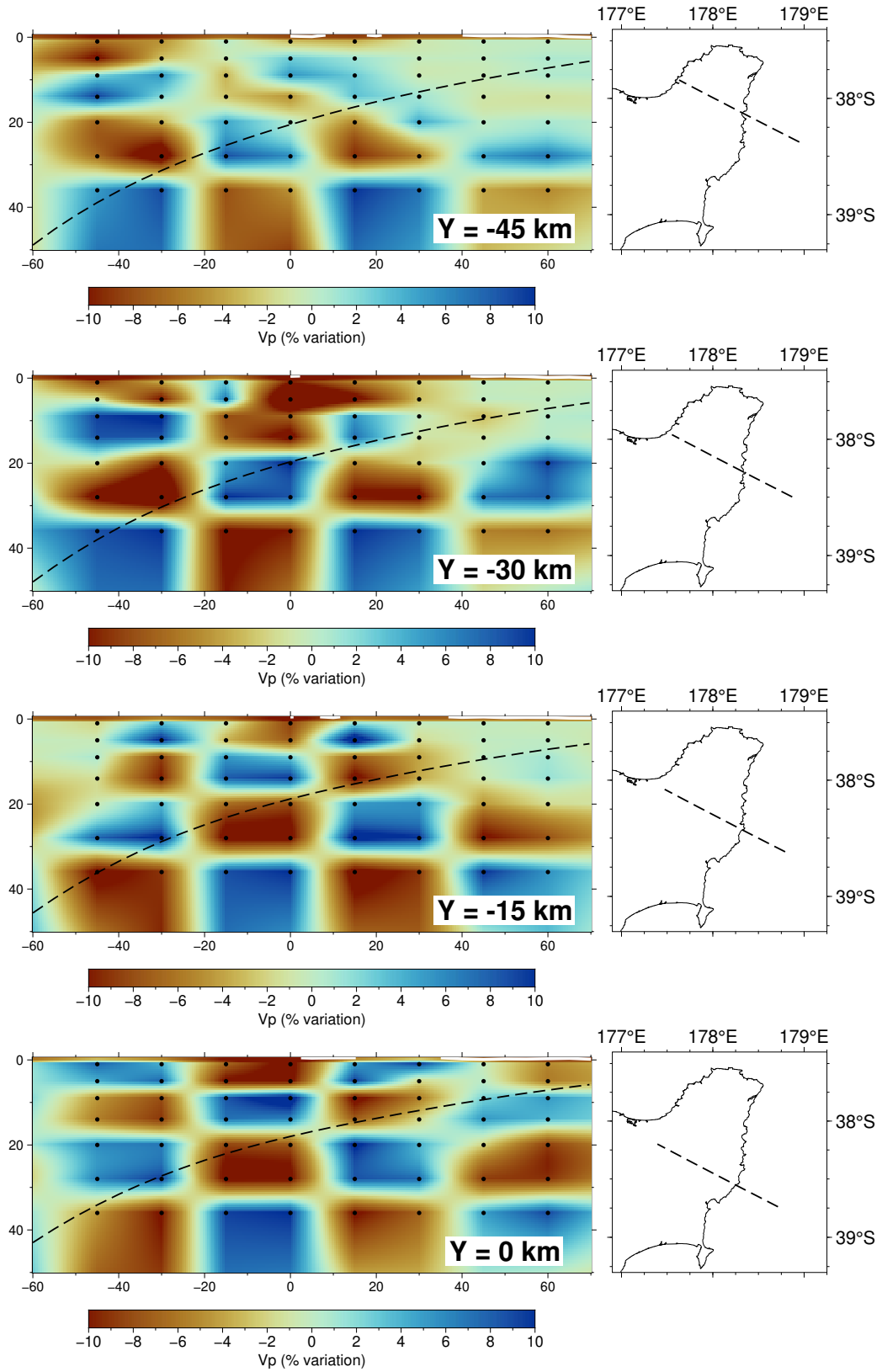


Figure E.13: V_p coarse checkerboard recovery at $Y = -45$ km, $Y = -30$ km, $Y = -15$ km, and $Y = 0$ km.

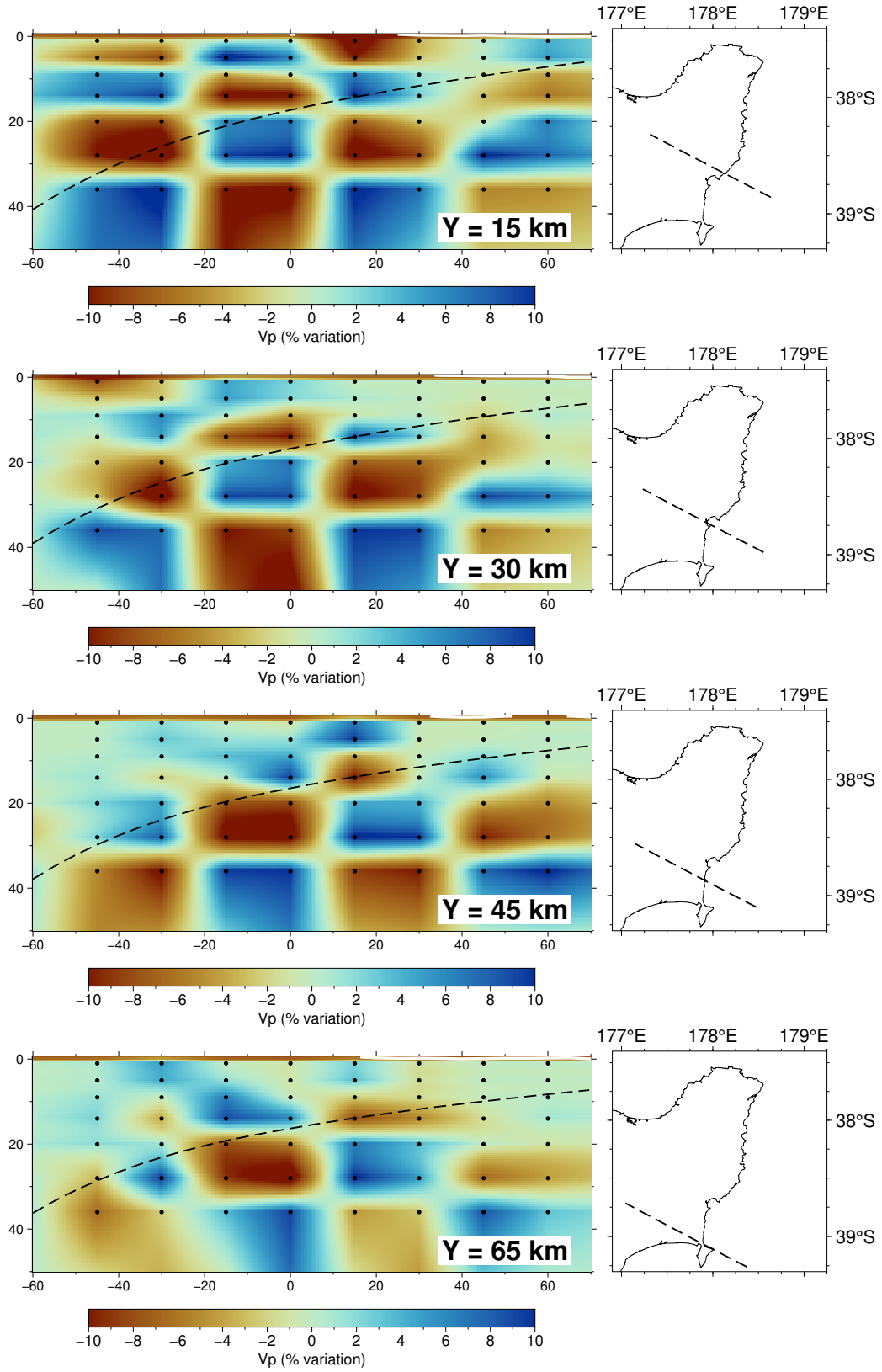


Figure E.14: V_P coarse checkerboard recovery at $Y=15$ km, $Y=30$ km, $Y=45$ km, and $Y=65$ km.

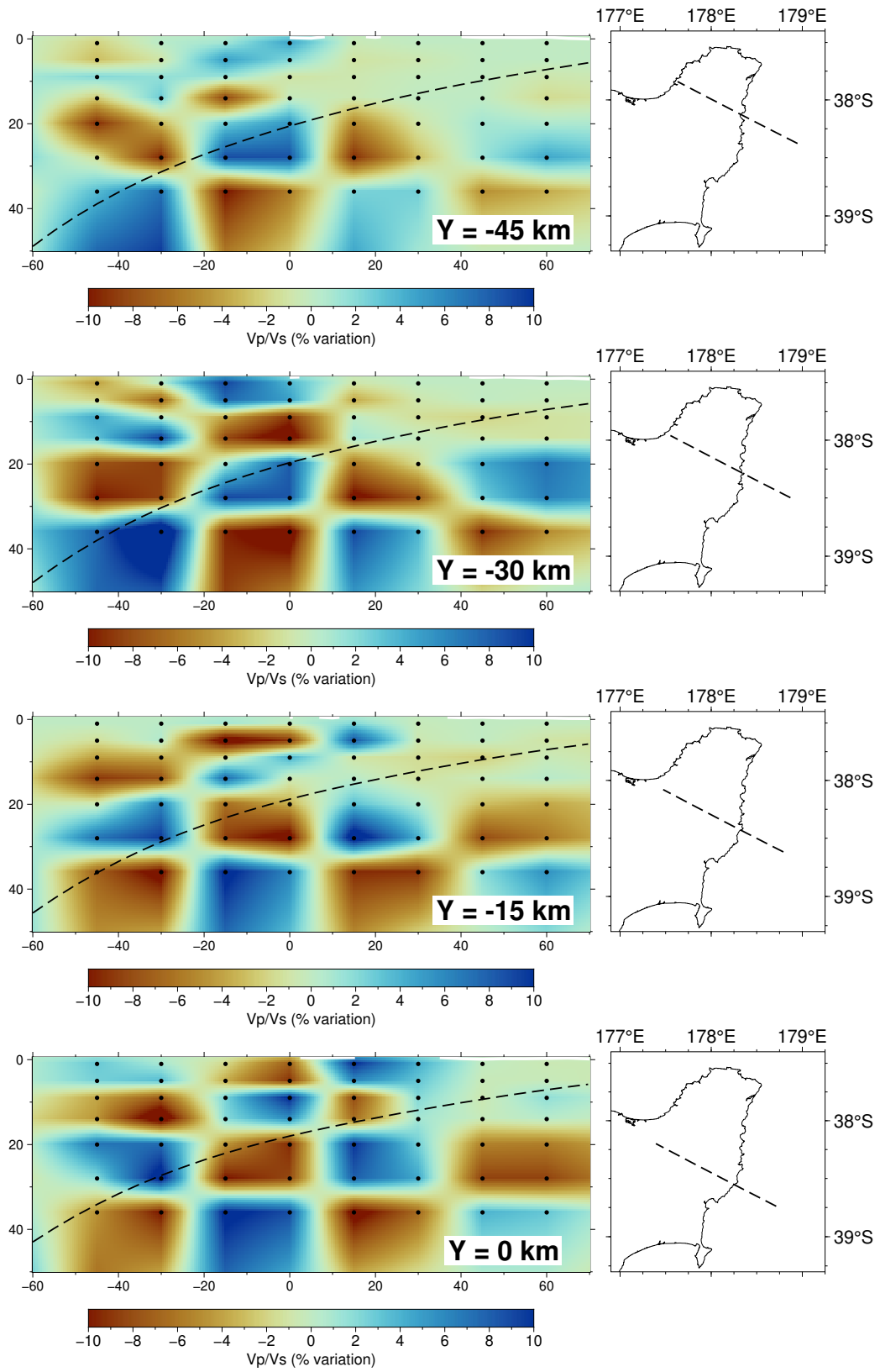


Figure E.15: V_P/V_S coarse checkerboard recovery at $Y=-45$ km, $Y=-30$ km, $Y=-15$ km, and $Y=0$ km.

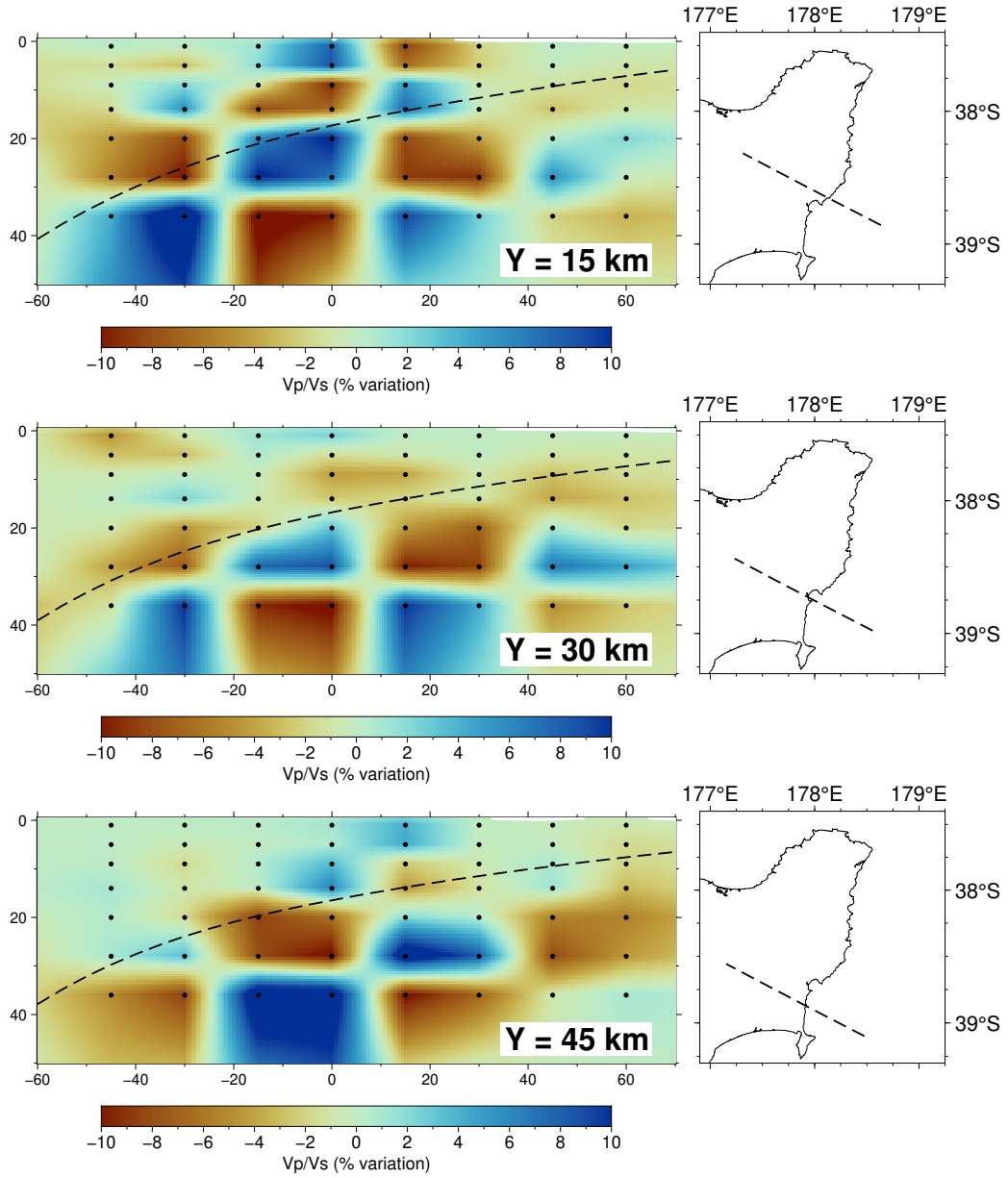


Figure E.16: V_P/V_S coarse checkerboard recovery at $Y=15$ km, $Y=30$ km, and $Y=45$ km.

Finer checkerboard tests

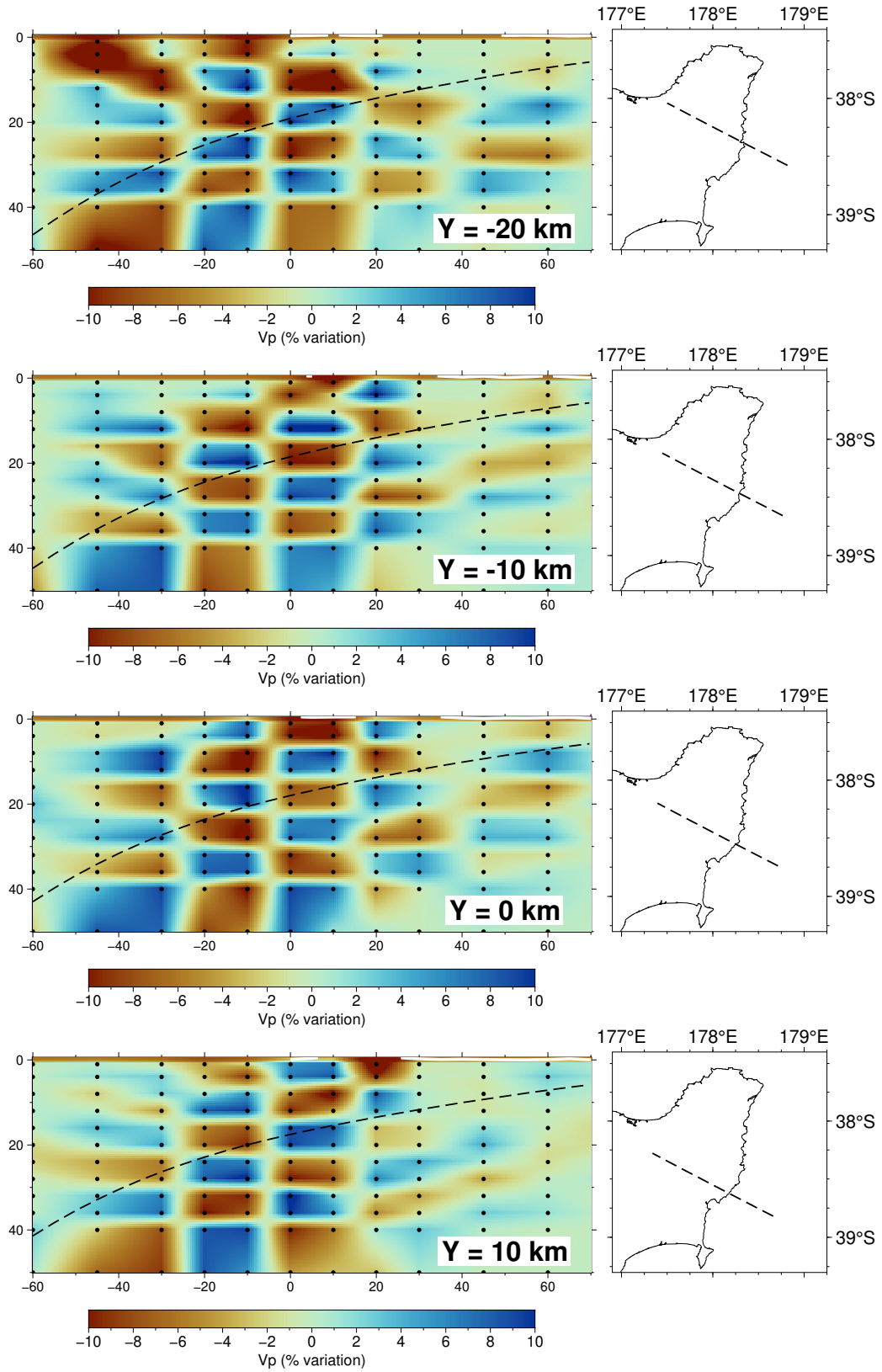


Figure E.17: V_p finer checkerboard recovery at $Y=-20$ km, $Y=-10$ km, $Y=0$ km, and $Y=10$ km.

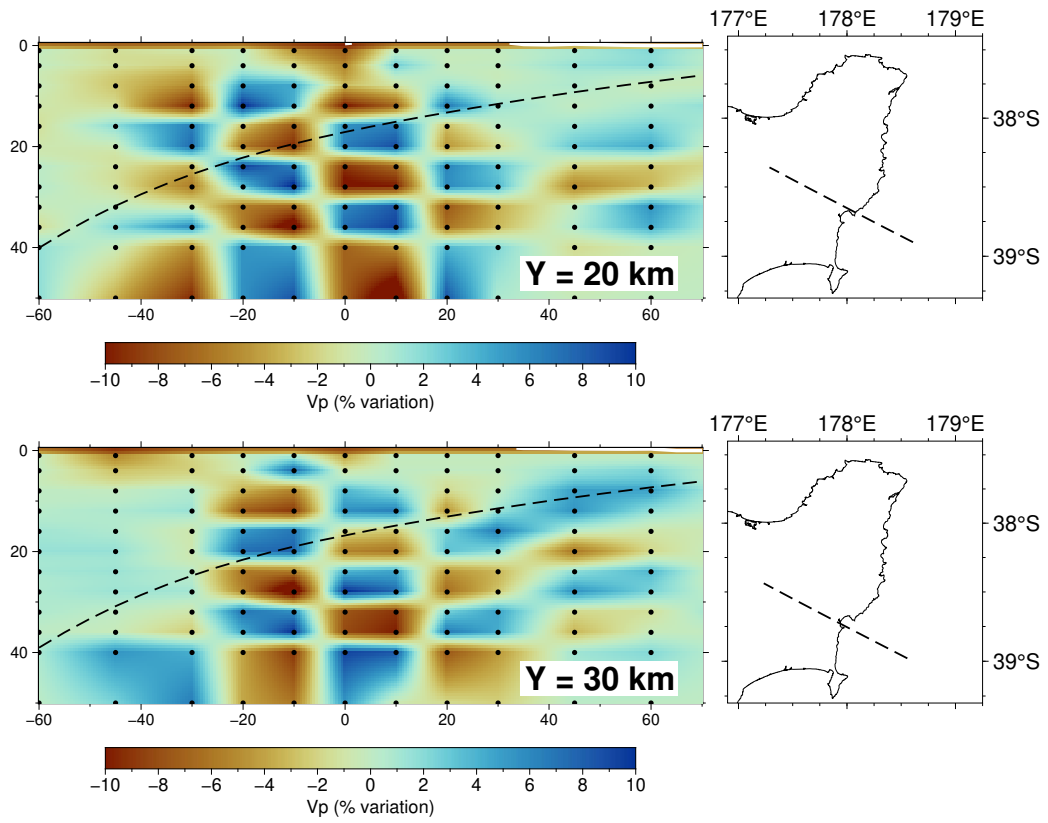


Figure E.18: V_P finer checkerboard recovery at $Y=20$ km and $Y=30$ km.

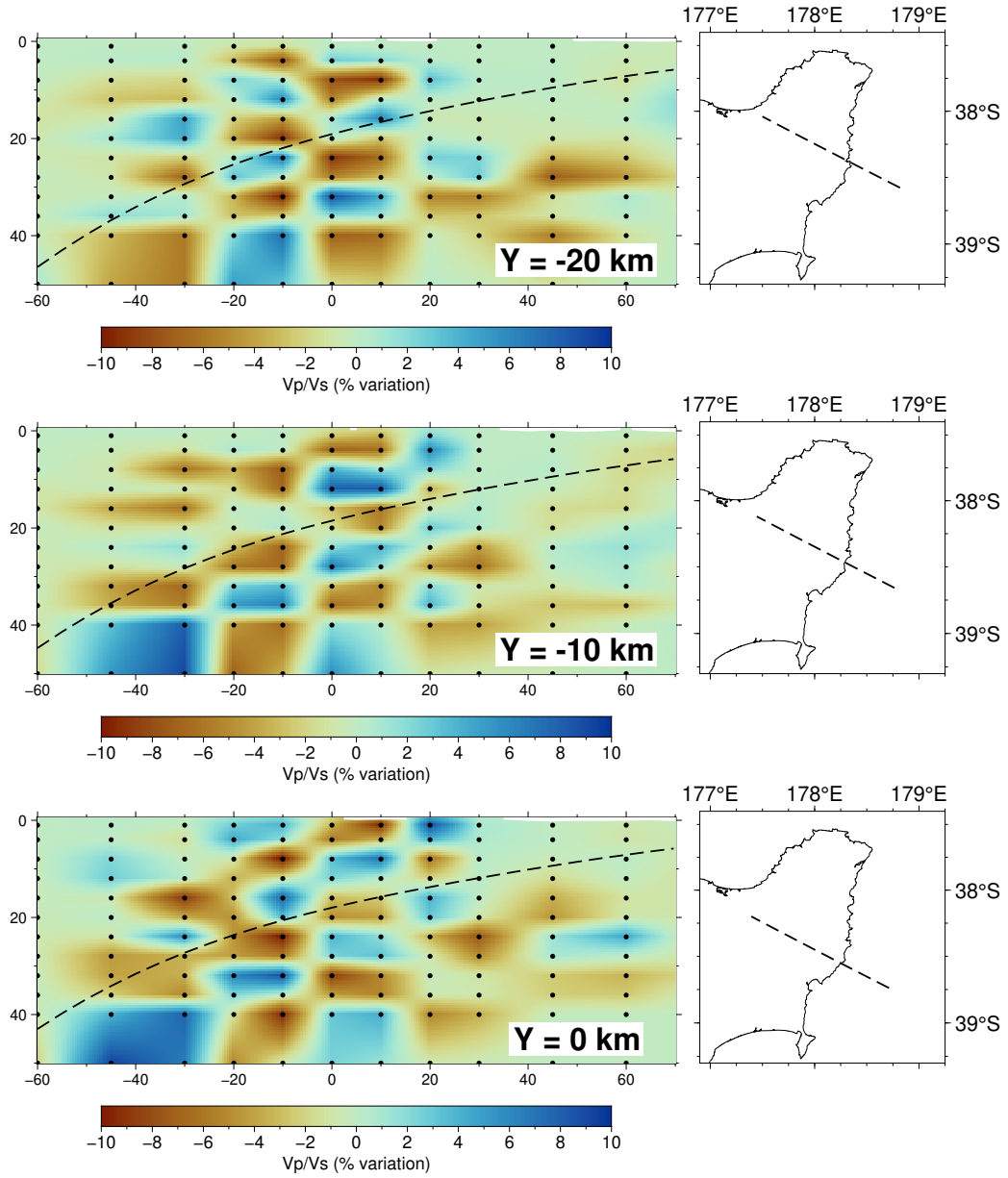


Figure E.19: V_p/V_s finer checkerboard recovery at $Y = -20$ km, $Y = -10$ km, and $Y = 0$ km.

E.1 De-clustered earthquake catalogue tomography results

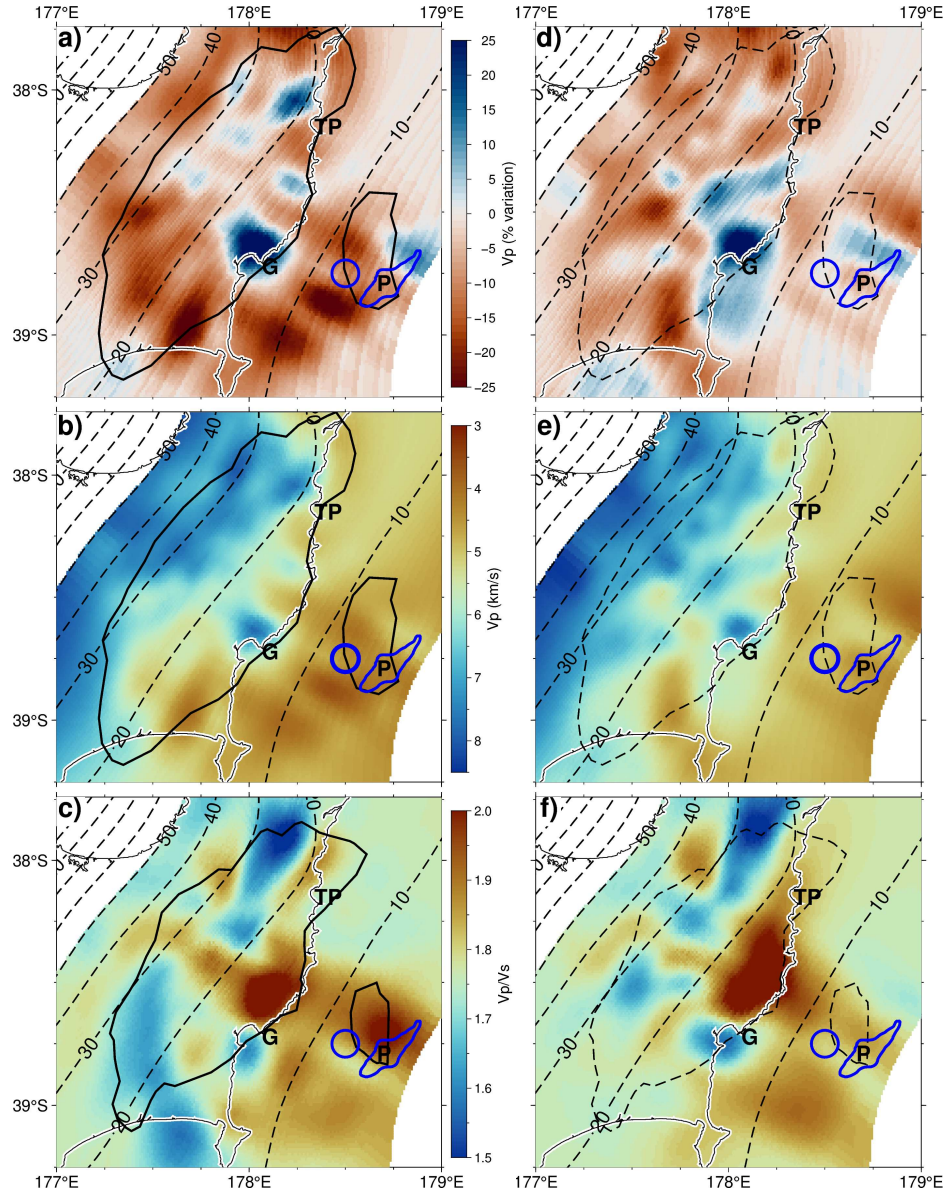


Figure E.20: a), b), and c) as in Figure 5.22. d), e), and f) show the plate interface V_P (percentage variation), V_P absolute values, and V_P/V_S models derived from the de-clustered catalogues. The same spread function contours are shown for both models for comparison, derived from the full earthquake catalogue inversion. Abbreviations: TP: Te Puia. G: Gisborne. P: Pāpaku seamount.

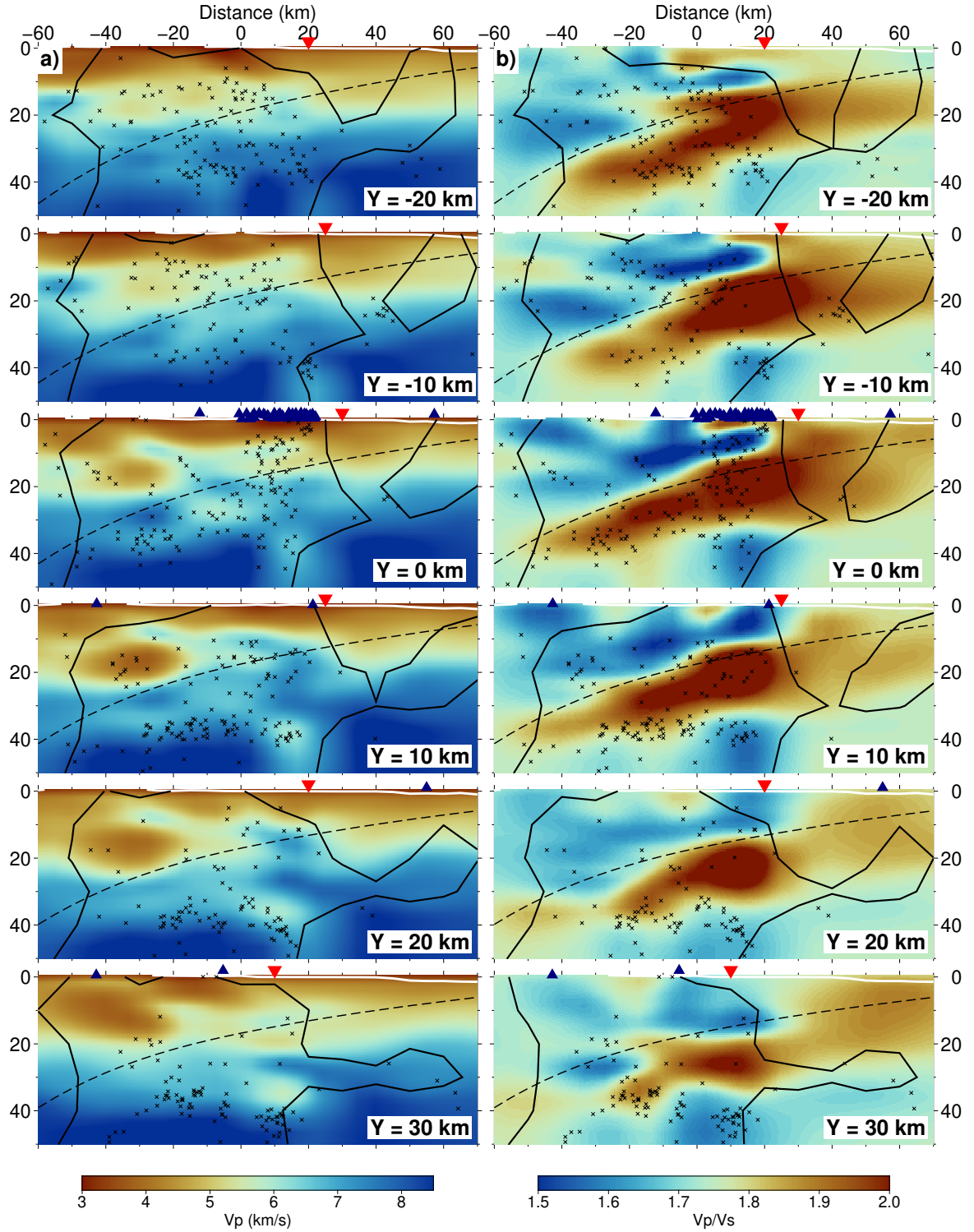


Figure E.21: V_P (left) and V_P/V_S models in cross section from north to south. Dashed black line: plate interface model (Williams et al., 2013). lines: spread function (SF) contours for V_P (SF = 2) and V_P/V_S (SF = 2.2). Navy triangles: seismometers. Inverted red triangle: coastline. Crosses: relocated earthquakes.