

Enhanced Multi-decade Long Earthquake Catalogs at Multiple Scales: Insights into
Volcanic and Tectonic Environments in the Pacific Northwest

by

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DISSERTATION ABSTRACT

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Doctor of Philosophy in Earth Sciences

Title: Enhanced Multi-decade Long Earthquake Catalogs at Multiple Scales: Insights into Volcanic and Tectonic Environments in the Pacific Northwest

Earthquake catalogs are useful for characterizing deformation within active volcanoes and faults. In the Pacific Northwest, USA, earthquakes are generated near the plate boundary, within the upper crust, and within magmatic plumbing systems in the Cascade Volcanic Arc. In all three cases, earthquakes pose significant hazard to densely populated regions. Yet, the mechanisms driving seismogenesis in these environments remain poorly understood. Here, the ability to characterize and understand the driving processes of active faulting and volcanism is limited by traditional earthquake catalogs for their limited quantity of small-amplitude earthquakes and lack of location precision.

In this dissertation, I aim to fill the observational gap by applying advanced earthquake detection and location procedures to curate a series of high-resolution enhanced earthquake catalogs—each spanning several decades long. I then analyze these catalogs to improve our understanding of volcanic and tectonic earthquakes within and near the Mount St. Helens magmatic plumbing system. Lastly, I apply recent advances in machine learning, high throughput association, and 3D velocity model to create an improved earthquake catalog for entire Pacific Northwest—which will eventually become the official seismicity catalog of the Cascadia Region Earthquake Science Center (CRESCENT).

This dissertation includes previously published and unpublished co-authored material.

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DEDICATION

Dedicated to my family and all those who have made this journey possible.

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CHAPTER I

INTRODUCTION

Earthquake catalogs are records that contain information about when and where an earthquake has occurred. In the Pacific Northwest, United States, seismicity is monitored and catalogued in real time by the Pacific Northwest Seismic Network (PNSN). To detect and locate earthquakes, the PNSN applies traditional phase detection and location procedures—including amplitude-based phase detection (i.e. short-term averaged / long-term average ratios (STA/LTA)), and location solutions that are performed using region-specific 1-dimensional velocity models. These methods are fast, robust, and are therefore well-suited to real time monitoring. However, because seismic data is noisy, STA/LTA requires high amplitude thresholds to prevent false detections. Thus, small-amplitude events are often missed and are therefore underrepresented in the PNSN catalog. Moreover, 1-dimensional velocity models do not capture widespread velocity heterogeneities of the Earth's structure, limiting the precision of the resulting earthquake location solutions. Both the underrepresentation of small earthquakes and low quality of location solutions limit the ability to characterize active faulting and volcanoes, both of which are of high value for understanding their underlying mechanisms and hazard potential. In this dissertation, we develop a series of enhanced earthquake catalogs of varying spatial scales to better understand the evolution of faulting and volcanism over several decade-long timescales. In all cases, we apply modern detection and location techniques—including template-matching and machine learning, with 3-dimensional

velocity models to detect and locate higher quantities of small-amplitude earthquakes at a precision that exceeds that of the PNSN catalog.

In Chapter 2, we create and analyze an improved catalog of volcanic seismicity at Mount St. Helens, Washington, USA. This work is accepted for publication in *Journal of Geophysical Research: Solid Earth*. It has passed several rounds of peer-review from the journal and internal review from the United States Geological Survey. The research design and conceptualization were conceived by myself, my advisor Dr. Amanda M. Thomas (University of California, Davis; previously University of Oregon), and Dr. Weston A. Thelen (Cascade Volcano Observatory). Dr. David R. Shelly (National Earthquake Information Center) provided detailed instructions to perform the template matching and location procedure. I conducted all numerical analyses and serve as corresponding author. All coauthors provided useful feedback which contributed to the result I present.

This analysis spans a sixteen-year period in a small region of roughly 5 by 5 km. Mount St. Helens is the most active and well-instrumented volcano in the Cascade Volcanic Arc. In Chapter 2, we apply a high-precision template-matching and location procedure to detect small earthquakes between the within the Mount St. Helens magmatic transport system beginning after the end of the last eruption in 2008 until 2024. Here, we find that most earthquakes occur beneath the historical lava dome complex. In addition, high amounts of deep earthquake activity after 2016 suggest that the Mount St. Helens plumbing system is repressurizing 8 years after the end of the last eruption. Here, we suggest that earthquakes during the 2008 to 2024 non-eruptive period at Mount St. Helens are driven by volatiles from the upper-crustal magmatic reservoir, which then interact

with structural variations in the shallow conduit system emplaced during previous eruptions.

In Chapter 3, I create and analyze an improved 16 yearlong catalog of non-volcanic seismicity around Mount St. Helens, Washington. Compared with Chapter 2, this study covers a larger study region (roughly 50 by 50 km) and is intended to draw connection volcanic plumbing systems and faulting in the upper crust of the Pacific Northwest.

Upper-crustal faults in the Pacific Northwest can host damaging medium to large-magnitude earthquakes. In addition, they may be partially responsible for the emplacement of modern-day volcanic systems, such as Mount St. Helens. In Chapter 2, we apply a similar detection and location procedure to Chapter 2 to detect, locate, and characterize earthquakes in faults adjacent to the Mount St. Helens volcano. Here, I find that non-volcanic earthquakes cluster around Mount St. Helens in four distinct groups. Analysis of their hypocenters, fault-plane solutions, and review of mapped geologic features suggest that the Mount St. Helens volcano may be underlain by a sub-orthogonally oriented fracture zone, providing a high-permeability pathway to transport volatiles and melt from midcrustal depths to the surface.

In Chapter 4, I create and analyze an improved 22 yearlong catalog of earthquakes recorded across the states of Oregon and Washington. Compared with Chapters 2 and 3, this study covers a longer timeframe and spans a much greater region. Here, I collaborate with my advisor, Dr. Amanda M. Thomas, Yiyu Ni (University of Washington), Dr. Marine Denolle (University of Washington), Loic Bachelot (University of Oregon), and Dr. Daniel Trugman (University of Nevada, Reno). All coauthors contribute to the

research design. I conduct all numerical analysis. The manuscript for this work is in progress and will be submitted to a journal that is yet to be decided.

The Pacific Northwest is positioned above the Cascadia Megathrust and is host to a wide variety of earthquake sources—including the subducting oceanic plate, the overriding plate, and volcanic sources. Here, I develop an automated workflow to catalog previously hidden events using phase picks from an advanced machine-learning phase detector, associator, and CRESCENT 3D Community Velocity Model. The result is an improved earthquake catalog with nearly four times as many earthquakes as contained within the PNSN catalog. Our catalog should give useful observations for future studies into the nature of seismogenesis across the Pacific Northwest.

CHAPTER II

**MAGMATIC VOLATILE FLUX DRIVES NON-ERUPTIVE VOLCANO-
TECTONIC SEISMICITY AT MOUNT ST. HELENS, USA FROM 2008-2024**

From Hirao, B. W., Thomas, A. M., Shelly, D. R., Thelen, W. A., & Journeau, C. (2025). Magmatic Volatile Flux Drives Non-Eruptive Volcano-Tectonic Seismicity at Mount St. Helens, USA From 2008–2024. *Journal of Geophysical Research: Solid Earth*, 130(12), e2025JB031278. <https://doi.org/10.1029/2025JB031278>

This work is published in Journal of Geophysical Research: Solid Earth as of December, 2025. The writing, methodology, computation, and conceptualization were done by my advisor, Amanda Thomas and I, with assistance from coauthors. David Shelly (USGS) assisted with the template-matching code. Weston Thelen (USGS) provided expertise on the Mount St. Helens volcano. Cyril Journeau (University of California, Davis) provided editorial assistance and broader perspectives on volcanoes.

2.1) Introduction

Seismicity beneath active volcanoes reflects localized deformation that may be attributed to the movement of magma and its exsolved volatiles within the volcanic plumbing system (e.g. Eaton and Murata, 1960; Minakami, 1974; Scandone and Malone, 1985; Swanson et al., 1985; Klein et al., 1987; Chouet, 1996; Roman and Cashman, 2006; McNutt and Roman, 2015). Mount St. Helens, USA, is the youngest and the most active volcano in the Cascade Range volcanic arc (Hoblitt et al., 1980; Decker and Decker, 1981; Topinka, 2004; Gleeson, 2020). Since the catastrophic flank collapse and eruption of May 18, 1980, Mount St. Helens has been consistently monitored with a regional- scale seismic instrument network (University of Washington, 1963; USGS Cascades Volcano Observatory) and has become one of the most thoroughly studied volcanoes in the world. In the last four and a half decades, Mount St. Helens has experienced two seismically recorded dome-erupting sequences: the first from 1980 to 1986 (e.g. Lipman and Mullineaux, 1981; Endo et al., 1981; Weaver et al., 1983; Swanson et al., 1985) and second, from 2004 to 2008 (e.g. Dzurisin et al., 2005; Moran et al., 2008; Iverson et al., 2005; Dzurisin et al., 2015). In addition, a series of phreatic eruptions were observed from 1989 to 1991 (Mastin, 1994). Sixteen years have passed since the end of the last eruption in January 2008, providing a unique opportunity to examine trends in seismicity and its evolution after a long-lived dome-building eruption (Cashman et al., 2008; Pallister et al., 2013) and how they reflect changes of a dacitic volcanic plumbing complex in the upper

crust (Cashman et al., 2017).

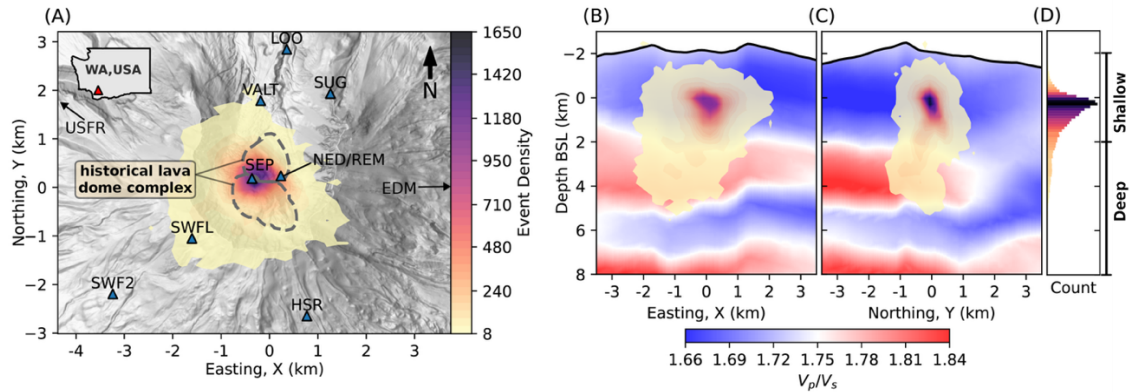


Figure 2.1: (A) Mount St. Helens, Washington, USA, study area projected along easting and northing in map view. Event density is depicted with contours, stations near the edifice are in triangles, and dashed outline shows boundaries of the historical dome complex. North of station SEP are lava domes from the 1980–1986 eruptions. South of station SEP are the 2004–2008 lava spines. Outlines of the lava dome complex are modified from Major et al. (2005) and Gabrielli et al. (2020). Stations USFR and EDM are nearby but outside of the map boundaries. Earthquake density contours are drawn from roughly 0.2 km by 0.2 km (0.04 km²) bins. Contours from bins containing at least 10 events each from the enhanced catalog in 2008–2024 in cross section along Easting (B), and Northing (C) superimposed on the Vp/Vs model from Kiser et al. (2019) velocity model, which we use for locating earthquakes in this study. Depth/X, Y aspect ratio in (B) and (C) is 0.7. Normalized earthquake depth histogram shown in (D). Full station configuration is shown in Fig. 2.S1. Topographic profiles in (B) and (C) are interpolated from SRTM (2013) at X = 0, and Y = 0, respectively.

2.1.1) Upper-crustal volcanic structure

Melt within the Cascade volcanic arc is sourced from below the Moho discontinuity at 35 to 40 km below sea level (BSL) (e.g. Stanley et al., 1990; Kiser et al., 2016, 2021). It is then transported towards the upper-crustal volcanic complex at Mount St. Helens through a vertically extensive transcrustal magmatic system (Smith and Leeman, 1993; Kiser et al., 2016; Cashman et al., 2017; Bedrosian et al., 2018; Leeman and Smith, 2018; Wanke et al., 2019; Kiser et al., 2020, etc.). Mount St. Helens is positioned above the St. Helens Seismic Zone (SHZ, Fig. 2.S1A), a NNW–SSE oriented strike-slip fault zone that facilitates the transfer of magma towards the upper-crustal reservoir (Weaver et al., 1987; Musumeci et al., 2002; Lehto et al., 2013). Numerous constraints from petrologic (Pallister et al., 1992; Pallister et al., 2008; Rutherford and

Hill, 1993; Cashman and Blundy, 2000; Cashman and McConnell, 2005), geodetic (Scandone and Malone; 1985; Anderson et al., 2010; Anderson and Segall, 2011, 2013; Wong et al., 2017; Wong and Segall, 2019; 2020), and tomographic studies (Lees 1992; Waite and Moran, 2009; De Siena et al., 2014; Kiser, 2018; Kiser et al., 2019, 2020; Ulberg et al., 2020) agree that the upper extent of the shallow crustal magma reservoir lies between 5 to 8 km BSL, or between 7 to 10 km directly beneath the crater floor of Mount St. Helens. Magma from the source reservoir is transported to the crater floor during eruptions through a narrow, vertical conduit (Pallister et al., 2008, 2013). In addition, field observations conducted after the 2004 to 2008 eruption reveal that the upper 0.5 to 1 km of the conduit is filled with a solidified dacite plug with low porosity and volatile content (Cashman et al., 2008; Pallister et al., 2013; Gaunt et al., 2014, 2016; Heap et al., 2016).

2.1.2) Seismicity during non-eruptive periods

The location, timing, and character of seismicity are useful for observing magmatic and volatile transport processes that occur during non-eruptive periods. Seismic events at Mount St. Helens are routinely identified, located, and published in an event catalog by the Pacific Northwest Seismic Network (PNSN) (University of Washington; 1963). Multiple types of volcanic seismicity have been identified at Mount St. Helens, including volcano-tectonic (VT) earthquakes (Moran 1994; Moran et al, 2008; Thelen et al, 2008), long-period (LP) earthquakes (Fehler and Chouet, 1982; Waite et al., 2008, Hansen and Schmandt, 2015; Matoza et al., 2010, 2015; Vidale et al., 2014; Glasgow et al., 2018; Nichols et al., 2011; Han et al., 2018), and tremor (Fehler et al, 1983; Moran et al., 2008; Denlinger and Moran, 2014; Malone, 2020; Köpfli et al., 2024). VT earthquakes, which are characterized by a waveform character and frequency spectra that resemble an

earthquake from a tectonic source (e.g. Endo et al., 1981; Malone et al., 1983; Power et al., 1994; Stephens et al., 1994), are the most commonly recorded seismic event type during non-eruptive periods (Moran 2008; Lehto et al., 2013; Glasgow et al., 2018).

VT seismicity during non-eruptive periods at Mount St. Helens is typically attributed to slip on pre-existing faults triggered by pore-pressure variations (Goto, 1999; Benson et al., 2008; Tuffen et al., 2008; Scholz et al. Incl, 2019) in crystallized magma or wallrock. These variations, thought to be due to magmatic volatiles and fluids that originate from the upper crustal magma chamber, are then channeled to shallower depths through the conduit (Moran, 1994; Musumeci et al., 2002; Moran et al., 2008; Zhang et al., 2022). Shallow VT activity between -2 to 2 km BSL is common and persistent during the non-eruptive periods from 1992 to 2004 and 2008 to 2023. By contrast, VT earthquakes deeper than 2 km tend to cluster within quasi-periodic swarms that last from several months to several years and are regarded as evidence for repressurization induced by magmatic recharge (Moran, 1994; Musumeci et al., 2002; Moran et al., 2008; Zhang et al., 2022). Lastly, VT earthquakes can occur adjacent to an upper crustal magma reservoir below roughly 8 km BSL, but rarely occur within it (Moran, 1994; Musumeci et al., 2002; Moran et al., 2008; Lehto et al., 2013; Zhang et al. 2022). In this study, we use an augmented seismicity catalog to observe spatial and temporal patterns in VT seismicity, from which we infer the locations of brittle structures, mechanisms of volatile transfer, as well as the evolving state of pressurization during the current non-eruptive period.

2.2) Methods and Data

We create an enhanced, high-resolution seismicity catalog with the goal of observing the evolution of non-eruptive seismicity between January 2008 and the end of

December 2023 (Video 2.4, Supporting Data 2.S1). We detect and locate both previously catalogued and newly discovered events by applying a network-based template-matching and location workflow similar to Shelly et al., (2013; 2016) and Ross et al., (2019), using multi-station P- and S-waveform templates from a total of 4,044 analyst-identified events recorded between January 2008 and December 2023 (University of Washington, 1963). We apply this workflow to a total of 57 short period broadband seismometers within approximately 50 km of the Mount St. Helens edifice. Station metadata may be viewed in Supporting Text 2.S5 and Supporting Table 2.S1. In this section, we summarize the key components to the detection and location workflow and provide additional details concerning data quality selection and location uncertainty estimation in Supporting Text 2.S1 and 2.S2.

2.2.1) Event Location: Three-dimensional velocity model

Accurately locating earthquakes at Mount St. Helens is challenging owing to its complex velocity structure and high topographic relief, which can influence seismic ray paths and travel times over typical source to receiver length scales (Neuberg and Pointer, 2000; Ripperger et al., 2003; Burjáněk et al., 2014; Zhang et al., 2022). To mitigate these effects, we use the three-dimensional (3D) P- and S-wave velocity models of Kiser et al. (2019). These models were derived from the recordings of active source borehole shots on over 3000 geophones during the Imaging Under Mount St. Helens (iMUSH) dense array experiment in 2014 (Kiser et al., 2016; 2019). We then resolve 3D P- and S-wave travel time grids using ‘Stingray’, a seismic ray tracing algorithm from Moser (1991) and Toomey et al., (1994). In addition, we input into Stingray the 15 second topographic model from the NASA Shuttle Radar Topography Mission (SRTM, 2013). The resulting

grids are centered at the middle of the Mount St. Helens crater at -122.19° longitude and 46.20° latitude, span 100 km along longitude (x), 95 km along latitude (y), and 40 km below the topographic surface (z), and have a spacing of 0.25 km. We use these grids to solve for absolute and relative locations in NonLinLoc (Lomax et al., 2000; 2014) and GrowClust3D (Trugman and Shearer, 2017; Trugman et al. 2023), respectively.

2.2.2) Event detection: template matching

Network-based template matching is well suited for detecting low amplitude events in volcanic environments (Shelly et al., 2013, 2016; Matoza et al., 2015; Hotovec-Ellis et al., 2018; Garza-Giron et al., 2023). Each template is associated with a starting location that is solved using analyst-confirmed phase arrivals with the NonLinLoc OCT-TREE nested grid search algorithm (Lomax et al., 2014). We then generate separate 2.5-second-long waveform templates for P- and S-wave arrivals for each station used to record the template, including all available channels. Each waveform template begins 0.5 seconds before its respective arrival time. We predict the arrival time of missing S-wave arrival picks using a V_p/V_s ratio of 1.65. If the difference between S- and P-wave arrivals is less than 2.5 seconds, the P-wave template is truncated to exclude the S-wave arrival. Continuous waveform streams are loaded and pre-processed using tools included in Obspy, a Python library for processing seismological data (Beyreuther et al., 2010). Prior to cross-correlation, we detrend, taper, interpolate to 100 Hz, and filter each template and day-long waveform trace with a zero-phase Butterworth filter between 2 and 20 Hz.

Each template is cross-correlated with continuous seismic data on all days that have at least 4 available stations within the template between January 2008 to December

2023. Day-long normalized cross-correlation (NCC) functions generated with the Super Efficient Cross-Correlation package (Senobari et al., 2019) are then summed across all channels and phases from the template. An event detection is considered when peaks of the summed NCC function exceed 12 times its median absolute deviation (MAD). We choose the $12 \times \text{MAD}$ threshold after Ross et al. (2019) to favor detection robustness over long time periods where station health and noise can vary significantly. The resulting cross-correlation and differential time measurements are input into the GrowClust3D algorithm (Trugman et al., 2023), which simultaneously locates newly detected events and relocates catalogued template locations. Our resulting catalog contains 31,133 events (about 7.7 times greater than the PNSN catalog) with magnitudes between -3 to 4.5 (Fig. 2.S3) within a roughly 4 km radius of the middle of the Mount St. Helens edifice.

2.2.3) Event magnitudes

We measure the size of both previously catalogued events and new events using the relative magnitude procedure described by Shelly et al. (2016). To reconcile changes in template event location and derive magnitudes for template events with a consistent scale, we compute the Richter magnitude (M_L) using the Obspy `estimate_magnitude` function (Beyreuther et al., 2010) and attenuation-correction terms of Bakun and Joyner (1984). The resulting M_L values are then extended to new events by computing the relative amplitude method of Shelly et al. (2016). Here, we refer to the extended local magnitudes as M_{L_ext} . Additional details for the magnitude computation procedure are described in Supporting Text 2.S4.

2.2.4) Changes to monitoring network and magnitude of completeness correction

Within 10 km of the modern-day crater floor, a total of 6 stations (CC.NFT, CC.HOA, CC.USFR, CC.REM, CC.SWF2, CC.LOO) were added and 3 stations (CC.SWFL, CC.NED, UW.JUN) were removed (Fig. 2.1A, Fig. 2.S1). Changes in the quantity of broadband stations near Mount St. Helens throughout the 2008 to 2023 study period can influence the ability of the network to detect and locate small magnitude events. Thus, changes in apparent event rate may reflect changes in the seismic network detection threshold, instead of true changes in seismic activity. To ensure a robust interpretation of long-term event rates, we apply a time-dependent magnitude of completeness correction within non-overlapping time windows. To test the effect of window length on completeness magnitudes (M_c), we compute M_c within 3-month, 6-month, 1-year, and 2-year long windows. For each window, we compute the frequency-magnitude distribution with 0.1 magnitude-wide bins and estimate the completeness magnitude of each window using the maximum curvature (MAXC) definition of Wiemer and Wyss (2000). Here, the M_c is equated to the magnitude bin with the highest frequency. We choose the MAXC definition because it is simple to compute and does not require assumed b-values (Lapins, 2022). To balance between temporal resolution and numerical stability, we apply the time-dependent M_c computed within 6-month long bins. We apply the M_c to the long-term event rates by removing events that are below the M_c at its respective time (Fig. 2.S2). Trends in both the uncorrected and M_c corrected seismicity rates closely resemble each other. Thus, analysis and discussion of event-rates presented in this manuscript are based on the uncorrected rates computed with all events in the catalog.

2.2.5) Moment tensor solutions

We attempt to constrain source mechanisms for individual earthquakes in the PNSN catalog using the MTfit package, which probabilistically solves for both full moment tensors and double-couple solutions using first-motion P-wave polarities (Pugh and White, 2018; Pugh et al., 2016). We analyze a selection of 31 well-recorded earthquakes with good focal sphere coverage and at least 15 analyst-picked polarities, which are accessed through Obspy (Beyreuther et al., 2010) from the IRIS database. Special care was taken to correct for inverted vertical channel orientations at some recording stations, according to the IRIS metadata. We choose not to use P-S amplitude ratios as we do not have empirical station correction terms to accommodate the heterogeneous site effects (e.g. Hardebeck and Shearer, 2003). We then solve for source-to-station azimuth and takeoff-angle grids using the NonLinLoc Time2Angles function, which are then interpolated for their precise location determined from the template match and relative location procedure. The polarities and maximum likelihood solutions for full moment tensor and double-couple inversions are shown in Fig. 2.S9, 2.S10, and 2.S11.

2.3) Results

2.3.1) Spatial Patterns

Most seismicity within our catalog is laterally confined within a 2-km radius centered near the middle of the 1980s crater floor (Fig. 2.1A, Fig. 2.2A). The distribution of earthquake epicenter density sharply concentrates towards the middle of this source region, which lies beneath remnants of the dacite lava domes and spines extruded during the 1980 to 1986 (e.g. Lipman et al., 1981) and 2004 to 2008 (e.g. Cashman et al., 2008; Pallister et al., 2013) eruptive periods, respectively. Nearly all seismicity within our catalog occurs at depths between the crater floor and around 7.5 km BSL, with the lower

limit of seismicity roughly consistent with the upper extent of the magmatic source reservoir. For convenience, we refer to depth regions between the crater floor (-2.2 km BSL) to 2 km depth BSL as ‘shallow’, and 2 km to 8 km depth BSL as ‘deep’. The boundaries we define here are not to be confused with those defined by other studies, which may include: 2-3 km below a datum at 1.1 km above sea level (ASL) in Moran (1994) and Moran et al., (2008), 0.5 km BSL in Zhang et al., (2022), 3 km below a 2.2 km ASL datum in Lehto et al., (2010; 2013), and 2 km below the surface in Waite et al., (2008). Approximately 87% of all hypocenters occur at depths shallower than 2 km BSL (Fig. 2.1D) (25,855 shallow and 5,278 deep earthquakes). Most shallow earthquakes localize within a distinct seismogenic volume of roughly 1 km radius area between -0.5 to 2 km depth BSL (Fig. 2.1B, 2.1C, 2.1D). In further discussion we refer to this shallow zone as the ‘seismogenic volume’ (Fig. 2.2A). Similar groups of persistent shallow earthquakes have also been previously identified between 1991 to 2004 and after 2008, which have been referred to as a shallow ‘seismogenic volume’, ‘seismicity band’, or ‘seismic lid’ (Moran et al., 2008; Blundy et al., 2008; Lehto et al., 2013; Zhang et al., 2022).

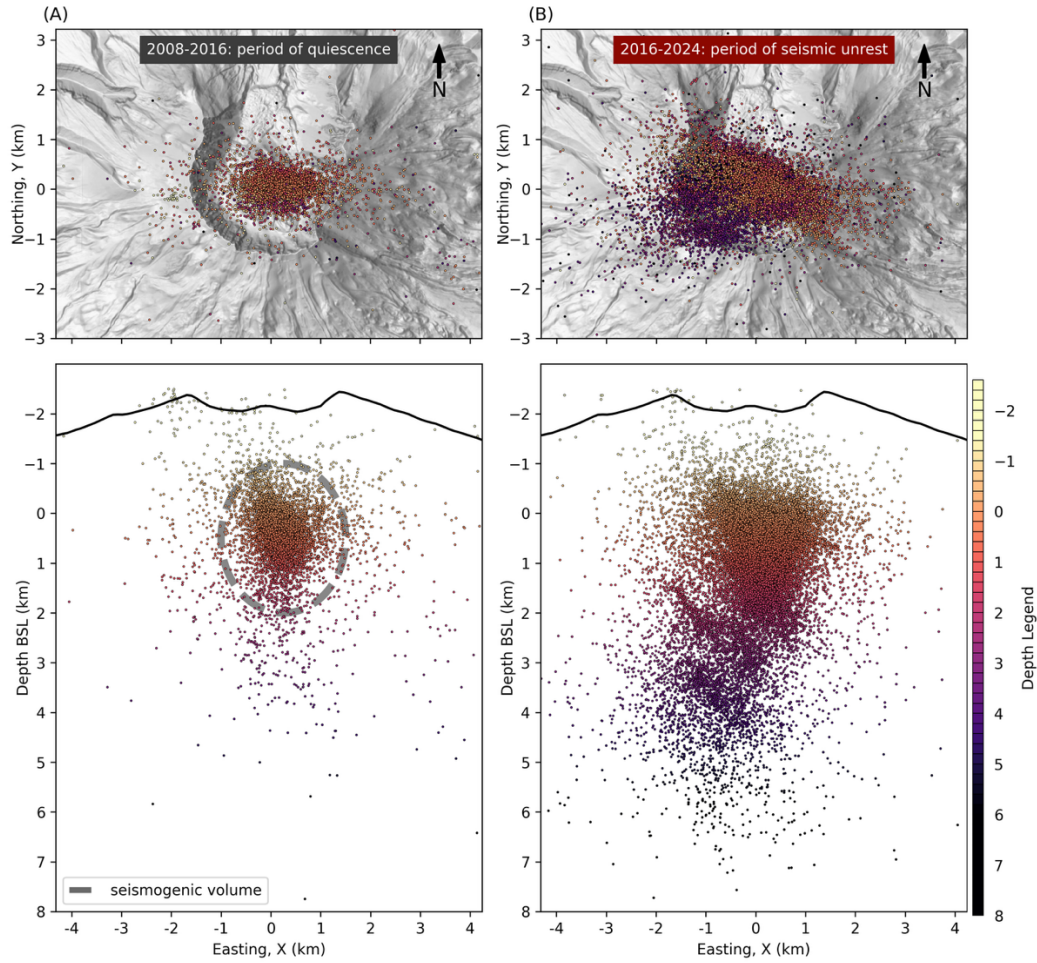


Figure 2.2: Map and depth sections of seismicity within the periods (A) 2008 to 2016 and (B) 2016 to 2024 (see Figure 3). Seismogenic volume is drawn in (A).

The uncertainties of deep earthquake locations are nearly twice as large as uncertainties for shallow earthquakes (Fig. 2.S5). In addition, all earthquakes have location uncertainties that are twice as great along the east-west direction as along the north-south direction, owing to the station geometry near the edifice (Fig. 2.1A, Fig. 2.S1). The median one standard deviation uncertainties along east-west (X), north-south (Y), and depth are 0.66, 0.32, and 0.59 km for shallow earthquakes and 1.01, 0.52, and 1.05 km for deep earthquakes. This may partly explain why deep earthquakes are more spatially

diffuse than shallow earthquakes, and why hypocenters occupy a wider footprint along the east-west direction compared to the north-south direction (Fig. 2.1, Fig. 2.2, Fig. 2.3).

2.3.2) Long timescale patterns

We divide the study period into two distinct eight-year-long periods from 2008 through the end of 2015 and from 2016 to 2024, which we refer to as the ‘period of quiescence’ and the ‘period of seismic unrest’, respectively (Fig. 2.2, Fig. 2.3). These two eight year-long periods are distinguished by the relative absence and abundance of deep seismicity, respectively. Shallow seismicity within and around the seismogenic volume (depth < 2 km BSL) is persistent throughout the entire 2008 to 2023 study period (Fig. 2.3D). Deep seismicity is rare between 2008 to 2016 but is frequent between 2016 to 2024, where it usually clusters within earthquake swarms that each last several months (Fig. 2.4). The number of broadband seismic recording stations within 10 km of the edifice increases from 8 to 15 between 2008 to 2023, with temporary decreases in station count due to poor weather in the winter months (Fig. 2.1A). For discussion, we arbitrarily define an earthquake ‘swarm’ as a continuous or near continuous period where the earthquake rate exceeds a ‘background rate’ threshold of 80 events within a one-week bin (Fig. 2.3A, Fig. 2.S2). Our choice of 80 events per week is to distinguish periods of relatively high or low event rates and does not reflect a criterion defined by previous studies. The long-term changes in event rates discussed here closely resemble those corrected for by the time-dependent magnitude of completeness (Fig. 2.S2), which suggests that they reflect true changes in seismic activity at Mount St. Helens.

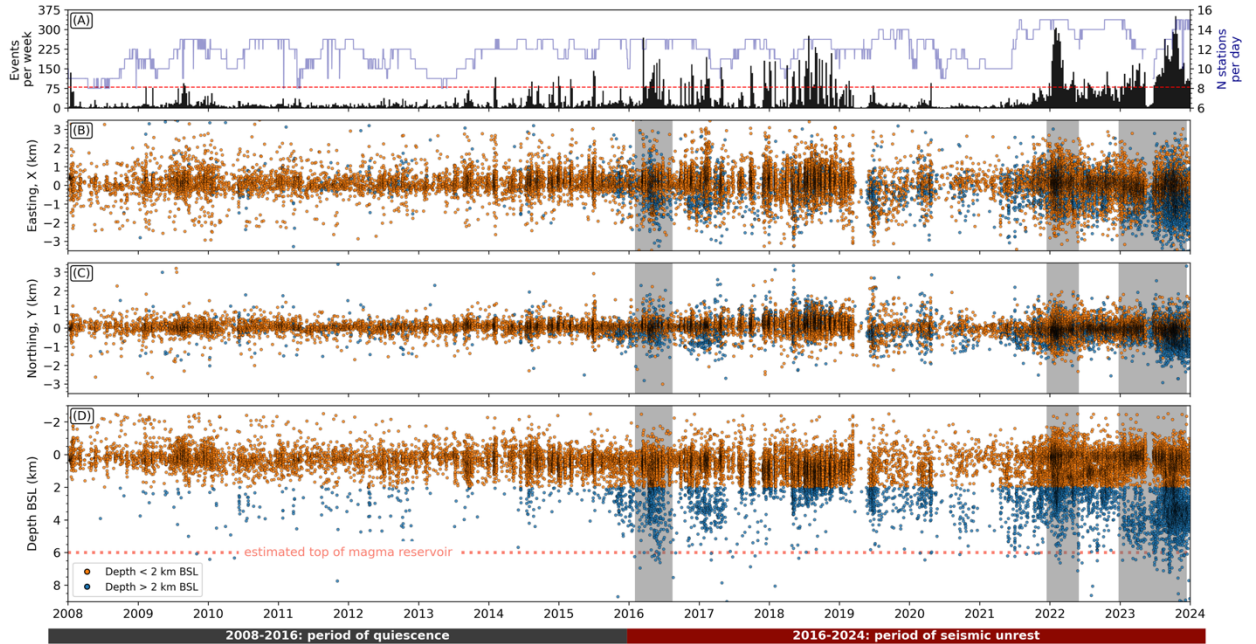


Figure 2.3: (A) Weekly earthquake rates (black histograms) and number of stations perday within 10 km of the edifice (blue line). Dashed red line in (A) depicts the 80 event per week rate threshold dividing earthquake swarm and background rate ranges. Individual days with fewer than 7 operating stations were removed to emphasize long term trends in station count. (B, C, and D) Earthquake locations along X, Y, and depth below sea level (BSL) for entire 2008 to 2024 study period. Earthquakes shallower than 2 km BSL and deeper than 2 km BSL are colored orange and blue, respectively. The estimated top of the magma reservoir (red dotted line in D) is constrained from numerous seismic, petrologic, geodetic, and tomographic studies listed in the Introduction. Light grey patches are swarms examined in Fig. 2.4A, 2.4B, and 2.4C. See Video S4 for animation of the entire time series. See Section 3.1 for discussion regarding depth datum and the boundary between ‘shallow’ and ‘deep’ seismicity depth ranges.

The period of quiescence begins in January 2008, during the final weeks of the 2004–2008 eruption (Dzurisin et al., 2015; Dzurisin, 2018). Small shallow swarms in early January mark the final stages of the eruption, after which seismicity drops to fewer than 10–30 events per week. Over the next eight years, activity is dominated by low-rate shallow seismicity near the seismogenic volume (Fig. 2.2A), occasional deep events, and weak inflation (Fig. 2.3A–D; Fig. 2.2A; Fig. 2.S4B; Dzurisin et al., 2015). A small number

of shallow swarms occur in January 2008 (Fig. 2.S5A) and August–September 2009 (Fig. 2.S5B), after which, seismicity rates remain low. Semi-periodic shallow swarms reappear in mid-2013, July–October 2014, and July 2015, with weekly rates peaking at 100–150 events (Fig. 2.3A; Fig. 2.S2C).

Beginning in 2016, the period of seismic unrest hosts several recurring earthquake swarms with relative increases in earthquake activity at shallow (less than 2 km BSL) and deep (2 to 8 km BSL) depth ranges. The years 2016, 2017, and 2018 each featured several swarms per year where seismic event rates reached 100 to 200 events per week, before decaying to background rates between 2019 and late 2021. The final two years of the study period, 2022 and 2023, feature the highest peak and persistent seismic activity levels in the entire study period. In 2022 and 2023, seismicity rates reached as high as 240 to 350 events per week.

2.3.3) Swarms: Shallow and Deep event timing

We explore the evolution of along-depth seismicity distributions over several-month-long timescales by examining the relative timing between shallow and deep earthquake daily event counts for three well-recorded swarms in March to July 2016, January to June 2022, and January to December 2023 (Fig. 2.4A, 2.4B, and 2.4C, Videos 2.S2-2.S4). All three examples show that shallow and deep event activity rates within earthquake swarms are temporally concurrent (Fig. 2.4A, 2.4B, 2.4C). In all three cases, the number of shallow and deep event rates per day both far exceed 10 times the average rates of a three-month period leading up to each respective swarm.

In general, event rate increases of shallow seismicity can either precede those of deep seismicity by several weeks or occur at roughly the same time. During swarms,

increased deep earthquake activity can persist continuously for several weeks to months. In addition, shallow seismicity rates remain elevated throughout periods of increased deep event activity. Shallow seismicity earthquake rates can decay concurrently with those of deep earthquakes or persist for several weeks after the rate of deep events has declined to pre-swarm levels. Continuous migration fronts are rarely observed. Instead, most seismicity occurs within burst-like clusters with complex short-term migrations that collectively move along depth over several-week-long timescales. We choose not to interpret individual small-scale migration patterns as individual geologic structures, since this would require rigorous location precision testing to ensure robustness (Battaglia et al., 2004; Thelen et al., 2008)

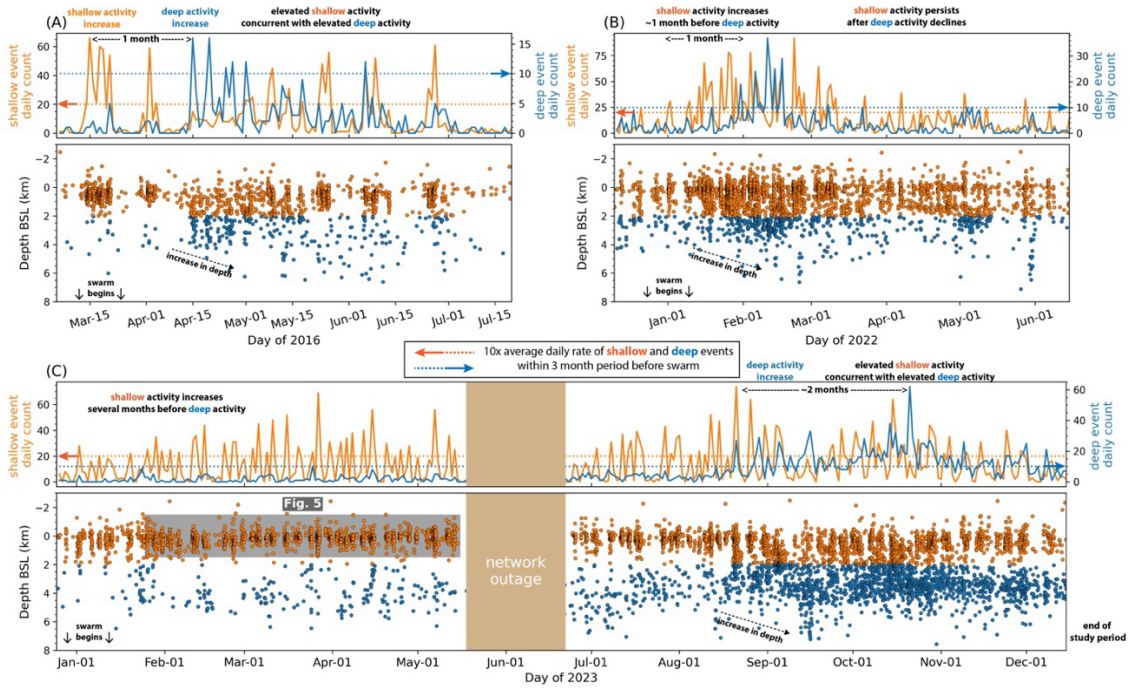


Figure 2.4: Earthquake swarms shown in grey patches from Fig. 2.3. (A) 2016, (B) early 2022, and (C) 2023 (C). Shallow and deep earthquake daily rates are shown in each top panel on left and right-side axes, respectively. The horizontal dotted lines show the equivalent of 10 times the average daily rate of shallow and deep events within a 3 month period before each swarm started. Earthquake depths plotted against

time on bottom axes. See Videos 2.2-2.4 for four-dimensional depictions of these swarms.

The 2016 swarm (Fig 2.4A, Video 2.2) features bursts of up to 60 shallow earthquakes per day, which precede the onset of near continuous deep activity by two to four weeks. As many as 60 shallow and 15 deep earthquakes per day were recorded between April 15 and July 1, 2016, before decaying to background activity rates. Similarly, the 2022 swarm (Fig 2.4B, Video 3) consists of shallow event rate increases that begin about four weeks before the onset of deep seismicity rate increase. Up to 80 shallow and 20 deep earthquakes per day were recorded between January 1 and April 1, 2022. A small, isolated cluster of around 60 earthquakes between 2 to 3 km depth occur between May 1 to May 15. An average of 50 earthquakes per week are recorded between June 2022 until the start of the third swarm in July 2023—significantly greater than rates observed during most of the 2008 to 2016 quiescent period.

The mid to late 2023 swarm (Fig 2.4C, Video 2.4) features the largest number of earthquakes and is the most long-lived swarm that we discuss in this section. Heightened seismicity rates during this period were originally reported by the Cascades Volcano Observatory (2023). A portion of this swarm is hidden by a network-wide service outage between May and June 2023. Like the 2016 and 2022 swarms, increases in deep seismicity rate occur after increases in shallow activity rates. However, higher-than-average levels of persistent shallow and deep seismicity started shortly after the end of the January to June 2022 swarm. Between 20 to 70 shallow and 10 to 50 deep earthquakes were consistently recorded per day between mid-August until the end of December 2023. This swarm may extend into 2024, beyond the end of the study period.

2.3.4) Shallow burst-like swarms

A close inspection of shallow seismicity during the 2016, 2022, and 2023 swarms indicates that many shallow earthquakes occur in quasi-periodic 'burst-like' clusters, which we refer to as 'shallow seismicity bursts' (also referred to as 'bursts'). We focus on a well-recorded sequence of shallow bursts from late-January to mid-May 2023 for in-depth analysis (Fig. 2.4C, Fig. 2.5, and Fig. 2.6). The events within shallow seismicity bursts during this period are concentrated in a compact source volume between -0.5 to 1 km BSL (Fig. 2.5A), located in the shallow seismogenic volume (Fig. 2.1, Fig. 2.2A, Fig. 2.7). The timing and recurrence intervals (Fig. 2.5B) show that each shallow burst lasts from one to two days in duration, recurs between one to four days of each other, and contains between 5 to 30 shallow earthquakes.

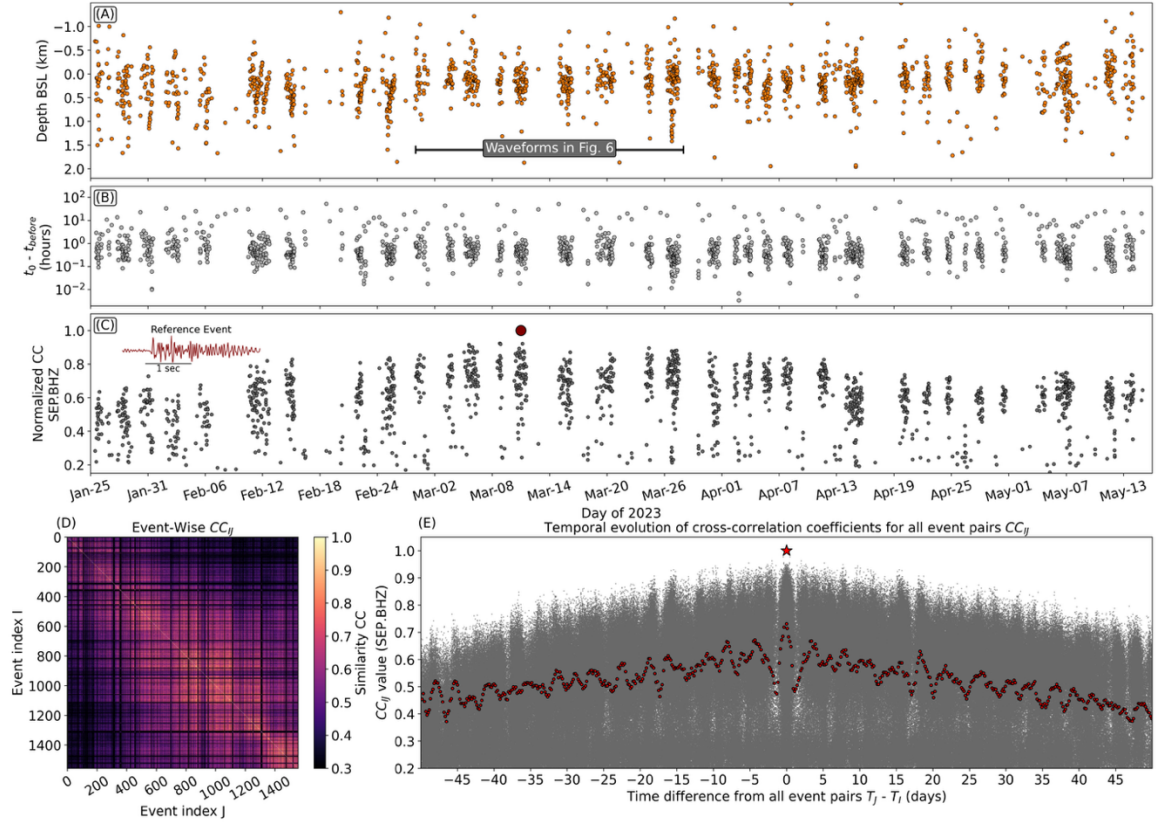


Figure 2.5: Shallow seismicity bursts between January and May, 2023, from Fig 2.4C. Time series of (A) depths, (B) recurrence intervals, and (C) cross-correlation coefficients of all events in this period with respect to a reference event (dark red dot in C) on March 11, 2023. (D) Event-wise cross-correlation matrix. (E) Change in event-wise cross-correlation values (grey dots) for all events with respect to any reference event (red star) event in (A) plotted versus the relative time difference in each event pair. This graph shows the same information as (D) but is projected with respect to the event pair time difference. Red dots are the mean values for each four-hour long bin. See Fig. S7 for a detailed inspection of inter-event cross-correlation values of P-S and coda phases.

We explore inter-event waveform similarity for all shallow earthquakes within this period by computing the event-wise cross-correlation coefficient for every event pair recorded on the vertical EHZ channel of station SEP, as it is the nearest station to the source region. Prior to event-wise cross-correlation, the waveform of each event is trimmed to 4 seconds in duration beginning 0.5 seconds before the expected P-wave arrival, interpolated to 100 Hz, and high-pass filtered above 1 Hz. We show the resulting inter-

event cross-correlation between a reference event on March 11, 2023 (referred to as T_0) in Fig. 2.5C and the full inter-event cross-correlation matrix (CC_{ij}) in Fig. 2.5D. The inter-event cross correlation indicates a steady increase in waveform similarity from no greater than 0.65 in late January 2023 to roughly 0.9, a few hours before T_0 , followed by a decline at a similar rate until mid-May. The progressive increase and decrease of mutual waveform similarity is a common feature among several events within shallow bursts between January and May 2023 (Fig. 2.5D, 2.5E).

To investigate whether the long-term change in inter-event cross-correlation is caused by an evolution in the location, source mechanism, or path, we separate the waveforms into two sections--the first containing body wave or P- and S-wave arrivals and second containing coda phases (Fig. 2.S7). We then compute the inter-event cross-correlation for both sections independently. Most of the body wave inter-event cross-correlation values are between 0.75 and 1, suggesting a high degree of similarity with a slight increase leading to the reference event and slight decrease afterwards (Fig. 2.S7A, 2.S7D). By contrast, the evolution of inter-event coda phase cross-correlation coefficients closely resembles the trends observed in those of whole waveform inter-event cross-correlation values (Fig 2.5C, 2.5D, 2.5E). Inspection of the aligned waveforms shows a consistent first-motion polarity.

Individual events within shallow bursts are of low signal to noise, however their near-repetitive waveform character makes it possible to stack and characterize several hundred similar events. We align and compute the NCC-weighted average of 300 events from this sequence to form a characteristic multi-station waveform stack (Fig. 2.6A). We then compute the continuous wavelet transform (CWT) spectrogram (Fig 2.6C) for the

stacked waveform on SEP.EHZ (Fig 2.6B), as it is the nearest station to the event source region. CWT analysis of the SEP.EHZ waveform stack shows that it is rich in coherent 2 to 20 Hz energy. In addition, the multi-station stack shows clear moveout of impulsive P-wave arrivals on vertical channels and S-wave arrivals on horizontal channels on seven stations nearest the source location (Fig. 2.1A, Fig 2.6A; all components shown in Fig. 2.S6). The amplitude of the coda phase decays to noise levels within roughly 10 seconds of the P arrivals on three of the nearest stations: SEP, VALT, and SUG.

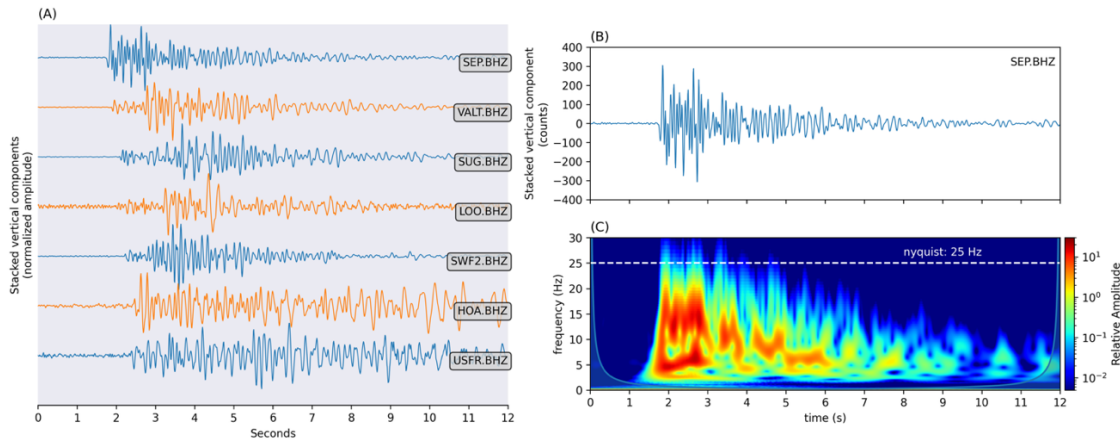


Figure 2.6: (A) Multistation waveform stack containing 300 shallow burst events from February to April 2023, shown in Fig. 2.5A. (B) Continuous wavelet transform (CWT) spectrogram (C) of stack from the SEP.BHZ channel.

Shallow bursts occur throughout most of 2023 (Fig. 2.4C), during the 2022 swarm from January to March 2022 (Fig. 2.4B), and to a lesser extent the 2016 swarm from March to June 2016. Do shallow burst swarms occur only during these select time periods or are they ubiquitous throughout the entire study period? The compact source region and consistent temporal clustering of events within burst-swarms (Fig. 2.5A, B) make it possible for a density- based clustering algorithm to identify individual shallow seismicity bursts. We apply ST-DBSCAN (Ester et al., 1996; Birant and Kut, 2007; Cakmak et al.,

2021) on all earthquakes between 2008 to 2024 to automatically detect groups of shallow earthquakes whose locations (X, Y, and depth) and origin times cluster together. ST-DBSCAN is a variant of DBSCAN (Ester et al., 1996) that is designed to accommodate both space and time dimensions. Clusters of data points are formed if the Euclidean distances of constituent points are within a user-defined space and time *epsilon* and minimum size thresholds relative to each other (Birant and Kut, 2007; Cakmak et al., 2021). We apply spatial and temporal *epsilon* values of 1 km and 5 hours respectively and require that each cluster include at least 15 events. Lastly, we consider all ST-DBSCAN-identified clusters with a mean location depth shallower than 2 km BSL as shallow burst swarms.

We identify a total of 336 individual shallow seismicity bursts between 2008 and 2024 (Fig. 2.7A). The labels assigned to individual bursts make it convenient to observe statistical patterns including the duration (Fig. 2.7B) and recurrence interval between bursts (Fig. 2.7C). Shallow seismicity bursts occur infrequently between 2008 and 2009, and 2013 and 2015 (Fig 2.7A). They occur frequently during seismic unrest between 2016 to early 2019 and late 2021 to 2024. All events within shallow seismicity bursts occur within a 1.5 km radius of the dacite domes (Fig. 2.1A, 2.7F, 2.7G), between depths of –1 to 2.5 km BSL (Fig. 2.7G, 2.7H). The depth distribution of all shallow burst events is densest at approximately 0.25 km BSL. Similarly, with all detected earthquakes in our catalog, events in shallow bursts do not preferentially occur during any season of the year (Fig. 2.7D), or time of day (Fig. 2.7E).

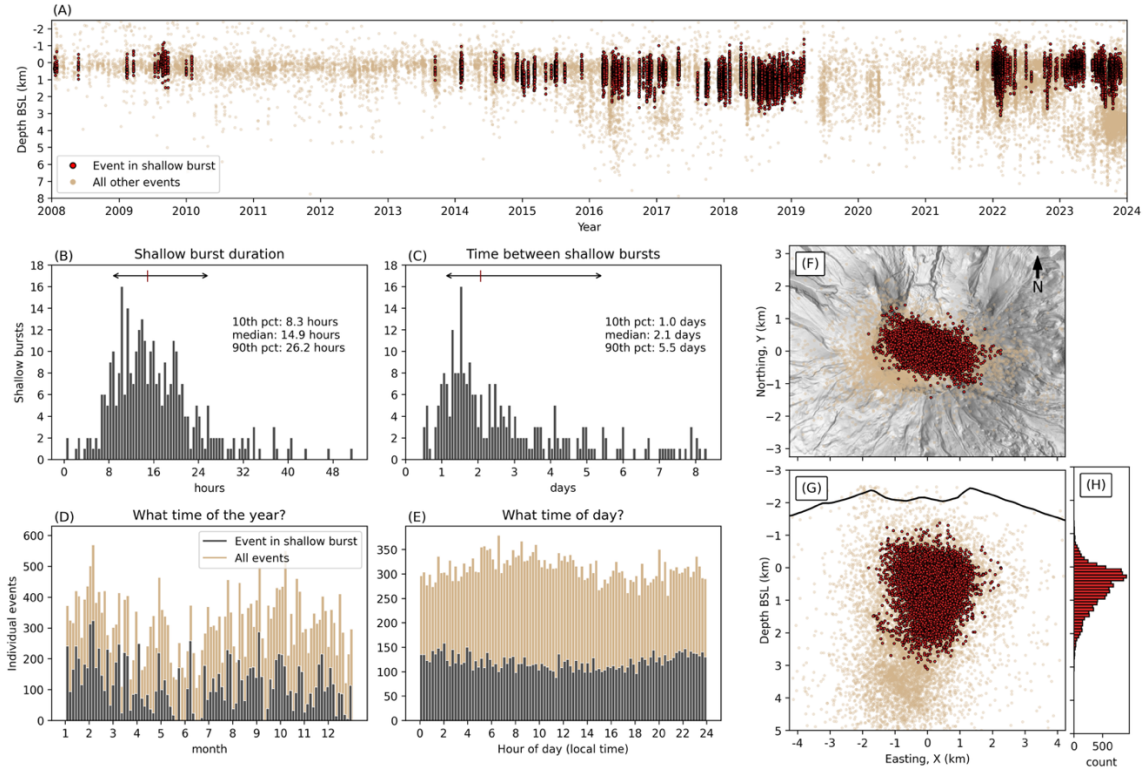


Figure 2.7: (A) Depth time series between 2008 and 2024 showing all detected shallow bursts using ST-DBSCAN. (B) and (C) are histograms showing distributions of durations of and time between individual shallow bursts within their respective sequences. Shallow bursts that are more than 8 days apart from each other are not a part of the same sequence. Horizontal arrow and red tick show the 10th to 90th percentiles and median values. (D) and (E) show distributions of month and hour of day for all individual events within shallow bursts and the entire dataset. Location of shallow burst events in map view (F), along depth (G), and depth distribution (H).

2.3.5) Source mechanisms

Our attempts to constrain moment tensor and double-couple solutions are unsuccessful (Fig. 2.S9 and 2.S10). We examine 31 of the most well-recorded events in the PNSN catalog, each of which has at least 15 analyst-picked first-motion polarities. Despite reasonable focal sphere coverage, the resulting moment tensor and double-couple solutions are poorly constrained and contain several misfit polarities (Fig. 2.S9 and 2.S10). This could be due, in part, to difficulties in accurately constraining take-off angles within a complex velocity structure. However, from these 31 events, we find seven events between

2019 and 2022 located within the edifice between 2 to 3.5 km BSL that feature 15 or more positive and no negative first-motion polarities (Fig. 2.8A, 2.S11). These events feature impulsive P-wave arrivals, show high-frequency energy on borehole stations, and do not correlate with changes in shallow seismicity rates.

2.4) Discussion

2.4.1) Volatile flux as a driver of VT seismicity

The driving mechanisms of VT earthquakes are debated and likely specific to the volcanic environment of interest. At Mount St. Helens, we propose that VT seismicity is driven by volatile flux from the shallow magmatic system based on its repetitive swarm-like nature and proximity to the conduit. Although regional-scale tectonic earthquakes occur along the Saint Helens Seismic Zone (SHZ) (Fig. 2.S1A) we do not find evidence that tectonic earthquakes on the SHZ are related to seismicity near the Mount St. Helens volcanic edifice (Qamar et al., 1981; Weaver et al., 1987; Musumeci et al., 2002; Lehto et al., 2013).

In this section, we propose that VT seismicity at Mount St. Helens is generated by brittle deformation resulting from increases in fluid pressures on preexisting crack networks within the conduit and the surrounding bedrock (Hill, 1977; Goto, 1999; Roman and Cashman, 2006; Benson et al., 2008; Tuffen et al., 2008; Scholz, 2019). This increase in fluid pressure likely occurs due to volatile flux from the deeper portions of the shallow magma system (Edmonds and Woods, 2018; Edmonds et al., 2019). These volatiles could originate from deeper, aseismic magma injection or from second boiling—the continued release of volatiles during crystallization of magma as it cools post-emplacement (e.g. Roman and Cashman, 2018; Wech et al., 2020). Both processes can increase volatile

saturation and have been shown to drive repeating, non-eruptive seismicity in other volcanic environments (Cannata et al., 2013; Wech et al. 2020). Systematic analysis of the long-period seismic records at Mount St. Helens from 2008 to 2024 could help decipher the origin of these volatiles and better constrain their timing and depth.

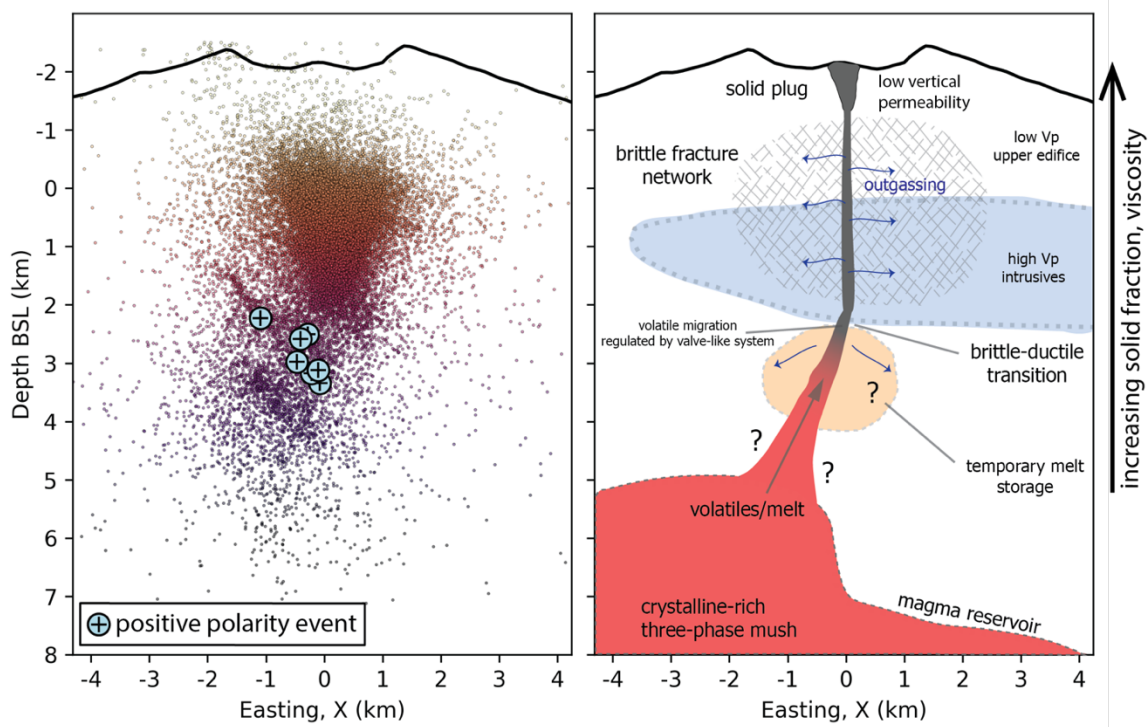


Figure 2.8: Interpretation of seismogenic features at Mount St. Helens. (A) All earthquake locations along depth, including the locations of 7 events with at least 15 positive and no negative first motion polarities (see section 2.3.5 and Fig. 2.S11, Fig. 2.S15). (B) Physical features affecting earthquakes. The magma reservoir and high Vp intrusives are traced from the Vp images of Kiser et al. (2018). The temporary melt storage zone is loosely based off the shallow magma storage zone from Lees (1992) and Waite and Moran (2009).

VT earthquakes at Mount St. Helens have been previously proposed to be associated with shear faulting induced by pressurization within the magmatic conduit and reservoir (Barker and Malone, 1991; Moran, 1994; Musumeci et al., 2000, 2002; Roman and Cashman, 2006; Moran, 2008; Lehto et al., 2010, 2013). However, VT earthquakes

can encompass a diverse range of source types with varying degrees of double-couple, isotropic, and dipole components (Julian et al., 1998; Templeton , 2006; Foulger and Julian, 2015; Mildon et al., 2016; Hrubcová et al., 2021). Moreover, we find seven earthquakes with all positive first-motion polarities across the focal sphere, suggest that some earthquakes at Mount St. Helens can contain isotropic components (Fig. 2.S11, Fig. 2.S15) (Aki, 1984; Foulger and Julian, 2015).

Spatial and temporal patterns of VT occurrence are also consistent with VTs being driven by volatile flux. Shallow seismicity is much more abundant than deep seismicity between 2008 and 2024 and it is persistent throughout the entire study period. Deep seismicity does not become common until 2016, when it occurs in recurring, semi-periodic swarms that last for several months. Event-rate changes within swarms could signal periods of elevated volatile flux over several weeks to months-long timescales. Shallow seismicity can occur with or without deep seismicity, whereas deep seismicity rarely occurs without concurrent shallow seismicity. If seismicity at all depths is driven by gas flux that originates from the upper-crustal reservoir, it appears to traverse several kilometers of conduit, often without triggering seismicity until it reaches shallow depths. Moreover, this spatial distribution of VTs indicate that the shallow edifice is more prone to brittle deformation than the deep edifice.

We suggest that these macroscopic observations can be explained by (1) a brittle-ductile transition (Fournier, 2007; White and McCausland, 2019) at about 2 km BSL, and (2) reduced vertical permeability towards shallow depth (Fig. 2.8). The proposed brittle-ductile transition at 2 km BSL is roughly consistent with the 80th percentile threshold of focal depths (1.8 km BSL) suggested by Smith and Bruhn (1984). To accommodate these

rheological and permeability changes, we consider along-depth changes in the bulk solid fraction of resident magma in the conduit (Fig. 2.8). Upon ascent, magma decompresses and crystallizes as it exsolves volatiles (Cashman and Blundy, 2000). In turn, increases in solid fraction within highly crystalline, silicic magma produce substantial changes in viscosity and permeability (Kozono and Koyaguchi, 2012; Schneider et al., 2012; Wong et al., 2017, Cassidy et al., 2018). Small increases in solid fraction can cause increases in magmatic viscosity by several orders of magnitude and affect brittleness of the shallow edifice (Cashman et al., 2017). Moreover, the resulting loss of volatiles and collapse of pore spaces triggers reduction of porosity, which can decrease permeability by several orders of magnitude in the upper conduit (Rust and Cashman, 2004; Heap et al., 2016; Wong and Segall, 2020). Thus, brittle deformation is favored in shallow depths owing to higher solid volume fraction and reduced vertical permeability. The resulting mechanical and permeability structures remain in place and serve as the initial conditions that produce VT seismicity thereafter.

Previous imaging of the Mount St. Helens magmatic systems provides some insight into the conduit structure and spatial patterns of VTs. Above 2 km BSL, the shallow seismogenic volume partially overlaps with a tomographically imaged high V_p anomaly, which is interpreted as a laterally extensive region of solidified granitic plutons (Fig. 2.8B) (Waite and Moran, 2009; Kiser, 2018). In addition, the base of the seismogenic volume and proposed brittle-ductile transition at 2 km BSL overlaps with a low V_p zone previously interpreted as a shallow magma storage region between 2 to 4 km BSL (Lees, 1992; Waite and Moran, 2009; De Siena et al., 2014). This feature, however, is not visible in more recent V_p models that use data recorded by the iMUSH experiment between 2014

to 2016 (Kiser et al., 2018; Ulberg et al., 2020). Both Kiser et al. (2018) and Ulberg et al. (2020) suggest that this may have been a temporary feature during and prior to the 2004 to 2008 eruption. At these depths, melt bodies may be unstable and transient (Cashman et al., 2017). Although the persistence of a shallow magma storage region after the 2004 –2008 eruption is unclear, this region may still accumulate transient batches of melt and gases from the upper-crustal reservoir below 5 to 6 km BSL as it repressurizes (Battaglia et al., 2018). Here, melt could be inhibited from migrating upwards due to the lower permeability of the overlying solidified magma. The movement of exsolved volatiles may then be regulated by a valve-like mechanism in the conduit at 2 km BSL, which may either permit it to migrate upwards through the conduit into the shallow fracture network or force it into the surrounding wallrock. Alternatively, upwelling volatiles may concentrate within zones of focused tensile stresses due to stiffness contrasts between the melt-rich zone and high-Vp intrusive body at 2 km BSL (e.g. Gudmundsson and Brenner, 2004; Kiser et al., 2018).

We infer that VT hypocenters occur within the conduit and surrounding wallrock. Additionally, they trend towards the WSW direction with increasing depth. Deep earthquake hypocenters are less spatially clustered than shallow earthquakes, possibly owing to their larger uncertainties (Fig. 2.3, Fig. 2.S8). Conduit geometry between 2 to 6 km BSL is only loosely constrained as its small aperture makes it difficult to image through tomographic methods. It is unclear if the WSW offset of deep seismicity reflects the geometry of the conduit between 2 to 6 km BSL. However, southwest offset of the lower extent of deep seismicity is consistent with the top of the upper-crustal magmatic reservoir in tomographic images from the iMUSH experiment (Kiser et al., 2018; Ulberg

et al., 2020). In addition, the WSW trend of deep hypocenters, given uncertainties, is roughly consistent with the location of a modelled pressurized cylinder proposed by Moran (1994) to explain focal mechanism P-T axis orientations of earthquakes at similar depths from 1987 to 1992.

2.4.2) Temporal trends

Shallow background-level seismic activity between 2008 to 2014 can be explained by steady or residual outgassing into the shallow brittle volume and near-magmatic conditions at the end of the eruption (Mastin, 2008, 2009; Moran et al., 2011; Anderson and Segall, 2013; Segall, 2013; Wong et al., 2017; Wong and Segall, 2019, 2020; Zhang et al., 2022). Battaglia et al., (2018) and Makus et al., (2024) interpret magmatic recharge of the upper-crustal reservoir between 2014 and 2016 from observations of mass addition and reduced Vp values, respectively. These two observations could indicate the onset timing of precursory recharge leading to seismic unrest, which we observe, beginning in 2016.

Episodic deep seismicity swarms between 2016 to 2024 bear some resemblance to deep seismicity swarms between 1987 to 2002 and could similarly be explained by repressurization of exsolved gases from the upper-crustal reservoir that are sealed within the plumbing system by the shallow impermeable dacite plug (Weaver et al., 1983; Barker and Malone, 1992; Moran, 1994; Musumeci et al., 2002; Moran et al., 2008; Zhang et al., 2022). High seismicity rates between 2022 to 2023 may also be explained by repressurization in addition to improved detection thresholds permitted by the installation of four permanent broadband seismometers within 5 km of the edifice between 2018 and 2022 (CC.LOO, CC.REM, CC.USFR and CC.HOA; Fig. 2.1A, 2.2A, 2.S1).

Deep and shallow seismicity rates within swarms are concurrent within several-

month-long timescales but are burst-like and complex within shorter timescales (Zhang et al., 2022). During the initiation of swarms, shallow seismicity rates increase either before or at the same time as deep event rates (Fig. 2.4). Additionally, maximum daily shallow seismicity rates are achieved several weeks to several months before maximum deep earthquake daily rates. These patterns could be explained by a combination of 1) temporary increases in vertical permeability by the formation of volatile flow channels during increased volatile output (Parmigiani et al., 2016, 2017; Degruyter et al., 2019) and 2) the brittle-ductile transition at 2 km BSL (Fournier et al., 2007; White and McClausland, 2019). Channelized flow pathways may be able to support volatile flux from the magma reservoir to shallow depths without triggering upward migrating seismicity. In addition, brittle deformation within the conduit beneath 2 km BSL may be inhibited by lower viscosities, however downward-migrating seismicity clusters during peak swarm activity rates indicate that this rheological transition can temporarily migrate to greater depths with high enough strain rates (Fournier, 2007, Cordonnier et al., 2012). Similar rheological transitions are additionally used to explain the lack of pre-eruption seismicity deeper than 0 km BSL (Dzurisin, 2018).

None of the swarms we observe between 2016 to 2024 are associated with detectable ground motion signals recorded by local GNSS stations (Fig. 2.S1, Fig. 2.S6), tiltmeters, or strainmeters. However, deformation consistent with inflation is recorded at local GNSS stations between 2008 to 2012 (Palano et al., 2012; Dzurisin et al., 2015; Poland et al., 2017; Makus et al., 2024). Post-eruptive inflation is ambiguous and could be explained by viscoelastic relaxation with or without contributions from magmatic recharge (Segall, 2016, 2019; Battaglia et al., 2018). Deep earthquake swarms in 1989 to 1992 and

1997 to 1999 coincided with explosive tephra emissions (Mastin, 1994; Moran 1994) and elevated CO₂ levels (Gerlach et al., 2008) respectively. However, there are no detectable outgassing events of CO₂, H₂S, or SO₂ between February 2008 and 2024 (USGS, 2023). The lack of observed H₂S and SO₂ emissions may be explained by scrubbing interactions with liquid water (Gerlach et al., 2008), but no anomalies that reflect increased outgassing of these chemical species have been detected in measurements from nearby hydrothermal springs (Symonds et al., 2001; Bedrosian et al., 2007; Bergfeld et al., 2017). Based on the absence of eruptive activity, recorded ground deformation consistent with inflation, and gas emissions at Mount St. Helens since January 2008 (Dzurisin et al., 2015b; Dzurisin, 2018), we suggest that the ‘failed eruptions’ model proposed by Moran et al. (2011) does not apply to the seismicity swarms we observe in the 2008 to 2024 study period. Thus, the ascent of large volumes of magma through the conduit does not explain the seismicity patterns we observe.

2.4.3) Shallow bursts and repeating earthquakes

Shallow burst-like VT swarms (Fig. 2.4C, 2.6, 2.7) are a curious new discovery revealed by the matched-filter detection and location workflow. Their compact source region, semi-regular periodicity, and waveform similarity may evoke comparison to drumbeat earthquakes recorded during the 2004 to 2008 eruption (Iverson et al., 2006; Harrington and Brodsky, 2007; Moran et al., 2008; Thelen et al., 2008; Waite, 2008; Matoza and Chouet, 2010; Thelen et al., 2011; Matoza et al., 2015). However, individual events within shallow seismicity bursts are distinct from drumbeat earthquakes mainly for their lack of dominant long-period energy, which indicates a different source type. Additionally, they are found during non-eruptive earthquake swarms and are roughly 2 km

deeper than drumbeat earthquakes. In addition, the lack of long-lived harmonic coda indicates that they do not trigger observable fluid resonance in the conduit (Chouet and Matoza, 2013; Glasgow et al., 2018). In comparison to repeating earthquakes on tectonic structures (e.g., central San Andreas Fault, USA (Nadeau and McEvilly, 2004)), shallow bursts at Mount St. Helens do not repeat over several year-long timescales but similarly occur in pulse-like swarms albeit over much shorter timescales.

Shallow burst-like activity rates are loosely concurrent with deepening swarms (Fig. 2.4, 2.7A), indicating a response of the shallow seismogenic volume to gas or fluid input from greater depths. Near-repeating waveforms with 2 to 20 Hz frequency energy content and consistent locations at -0.5 to 1 km BSL indicate repetitive fracturing within or adjacent to the shallow conduit beneath the dacite domes. In addition, semi-regular time intervals between bursts may reflect quiescent accumulation and valve-like release of fluid or gas overpressure from the conduit into a brittle fracture network (Farge et al., 2021). At Mount St. Helens, a valve at 2 km BSL within the edifice may represent a transient increase in permeability by hydrofracturing or capillary fracturing processes that are characteristic of high-crystallinity magmas (Oppenheimer et al., 2015; Degruyter et al., 2019; Parmigiani et al., 2016, 2017; Farge et al., 2021).

Shallow burst-like events persistently repeat within roughly one month of their respective origin times. Their P- and S-wave arrivals in early to mid 2023 are remarkably similar across this 5-month period (Fig. 2.5C, 2.S7). By contrast, the steady increase and decay of inter-event cross-correlations leading up to a reference event in Fig. 2.6B can be explained by changes in the coda waveform character (Fig. 2.S7). Hypocenters of individual events within a shallow burst can vary by 1 to 2 km (Fig. 2.5A). Further

refinement of location solutions may be required to verify whether their locations vary by this much or if they collapse into a smaller source volume. S-wave detections, of low signal to noise, that contain segments of shifted or altered coda wave trains could degrade the cross-correlation measurements, leading to increased hypocenter uncertainty. It is unclear if the progressive change in coda wave character can be explained by changes in location that are not resolved, or due to changes within the scattering medium. However, the consistency of inter-event body-wave cross-correlation values indicates that most events in this time share similar source mechanisms and locations. Well-recorded shallow burst events with mutually similar body waves and changing coda phases may provide an opportunity to image temporal velocity changes near the shallow edifice and flanks

(Snieder, 2006; Hotovec-Ellis et al., 2014, 2015; Kanu et al., 2013).

2.4.4) Environmental sensitivity

Changes in seismicity rates in response to external factors are well documented in volcanic environments (Hill et al., 1993; Neuberg, 2000; Brodsky and van der Elst., 2014). Stress field alterations due to steep thermal gradients between hot solidified magma and rock containing cool meteoric water have been proposed as a mechanism for driving fracturing potential swarm-like seismicity (e.g. Warren and Latham, 1970). However, the role of meteoric fluid in generating VT seismicity at Mount St. Helens is unclear. There are no known instances of seismicity rate changes during non-eruptive periods (i.e. when the vent is shut) in response to large rainfall events. Infiltration times longer than one seasonal cycle are possible; however, electric self-potential measurements made near the edifice indicate that fluid and circulation in the hydrothermal system above roughly 2 km of the

surface is fed by magmatic exsolution (Bedrosian et al., 2007). Moreover, at Mount St. Helens, there are few known geologic structures that serve to transmit meteoric fluid deep into the crust (e.g. Lassen Volcano, Long Valley Caldera; Janik and McLaren, 2010; Montgomery-Brown et al., 2019).

Correlations between annual peak seismicity rates and hydrologic well data have been identified at Mount St. Helens and other Cascade Range volcanoes between 1984 and 2004 (Saar and Manga, 2003; Christiansen et al., 2005). The time-of-year frequency histogram of Fig. 2.7D shows a slight increase in event count during the late summer months, when infiltration of snowpack melt is expected to reach its annual maximum flux (Christiansen et al., 2005). However, declustering is necessary to prevent bias due to overrepresentation of seismicity rates from anomalously large swarms. One may also need to explain why there is an additional peak in the January to March months, when snowpack is expected to accumulate. The hypothesis testing required to verify these apparent seasonal correlations are out of scope for this study.

Lastly, we do not find any evidence for seismicity rate changes immediately after phase arrivals from global teleseisms of M7 or greater, or regional earthquakes of M3 or greater. Tidal modulation, particularly for periodic shallow seismicity bursts, may be worth consideration. However, most shallow bursts recur within 1 to 2 days of each other (2.7E). If tidal stresses were a primary triggering mechanism, we might expect them to recur within diurnal or semidiurnal periods (Thomas et al., 2009, Scholz et al Incl., 2019; Wang et al., 2022).

2.4.5) Implications for future eruptive potential

Increased seismic activity eight years after the end of the 2004 to 2008 eruption of Mount St. Helens indicates increased levels of volatile flux in response to recharge or second boiling within the upper-crustal reservoir, which will eventually lead to the next eruption given sufficient time. Within the past 1,500 years, Mount St. Helens has experienced four 60 to 100 year-long eruptive periods, each consisting of several dome-building eruptions within decades of each other (Hoblitt et al., 1980; Evarts et al., 1987). Despite the apparent increases in subvolcanic activity inferred by seismicity rates, our observations should not be used to forecast the next eruption. Neither the 1980 to 1986 nor 2004 to 2008 dome-building eruptions featured precursory deep seismicity (Dzurisin, 2018; Roman and Cashman, 2018). Moreover, pre-eruptive seismicity is usually shallow and occurs at rates of hundreds of earthquakes per hour--much higher than we observe in our dataset (Moran, 1994; Moran et al., 2008; Roman and Cashman, 2018). Further investigations of non-eruptive seismicity, particularly their connection to observations from other disciplines, could be useful to better understand the physical processes that govern eruption cycles.

2.5) Conclusion

We derive a high-resolution earthquake catalog between 2008 and 2024 to better understand non-eruptive volcanic seismicity at Mount St. Helens. We observe widespread volcano-tectonic VT seismicity in proximity to the conduit between -2 and 7.5 km below sea level (BSL). Shallow seismicity beneath the volcano's recently emplaced dacite domes within -0.5 to 2 km BSL is persistent from 2008 to 2024, indicating that the shallow edifice is consistently brittle. Both shallow and deep seismicity rates increase dramatically within semi-periodic earthquake swarms beginning in 2016, indicating increased volatile

flux due to recharge or crystallization in the upper-crustal reservoir. Within swarms, peak shallow seismicity rates occur before peak deep seismicity rates. The long-term spatial and temporal seismicity patterns can be explained by a brittle-ductile transition at 2 km BSL and a decrease in vertical permeability within conduit magma towards the shallower edifice—both of which can be attributed to the upward solidification of magma within the conduit. The shallow edifice hosts semi-periodic shallow seismicity bursts that consist of low amplitude repeating earthquakes. Shallow seismicity bursts become frequent beginning in 2016 and tend to occur in temporal proximity to deep earthquake swarms, indicating a valve-like response to gas input from greater depths. Seismicity rates do not appear to change according to seasons of the year nor in response to phase arrivals from large earthquakes at teleseismic or regional distances. Considering these observations, we suggest that non-eruptive VT seismicity at Mount St. Helens is modulated by varying amounts of volatile flux from the upper crustal magma reservoir. Volatiles then interact with structural variations within the conduit that are the result of increasing solid fraction and decreased permeability toward shallow depths.

2.6) Acknowledgements

The earthquake catalog used in this study is provided in Supporting Data 2.S1. This material is based upon work supported by the U.S. Geological Survey under Grant No. G22AP00299 and G22AP00300 and by the National Science Foundation under Award 1848302 and Cooperative Agreement # 2225286. Part of this study was completed during the Pathways Internship at the USGS Geological Hazards Science Center. We offer thanks to Alicia Hotovec-Ellis, Robin Matoza, and an anonymous reviewer for helpful reviews that improved quality of this manuscript. In addition, we extend appreciation to

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2.7) Open Research

The earthquake catalog from this study is available in Hirao et al., (2025) and its headers are described in Supporting Text S6. Template event data, including their waveforms and phase arrivals, were obtained from the Pacific Northwest Seismic Network (PNSN; University of Washington, 1963).

2.8) Supporting Text

2.8.1) Supporting Methods – additional details for detection, quality control, and location

Each template cross-correlation function is evaluated at every 1/100th of a second, allowing for both positive and negative correlations. Multi-station normalized cross-correlation (NCC) functions are generated by stacking individual day-long NCC functions from each channel and phase. Day-long waveform traces are shifted by the respective phase and station travel time, allowing detection peaks to stack at the event origin time. We restrict the search window for differential times and NCC peaks for each channel and phase within ± 0.5 seconds of the event origin time. Sub-sample timing precision is achieved by a three-point parabolic interpolation of each cross-correlation peak. Station templates with an S minus P time difference of less than 0.8 seconds are discarded to prevent cross-correlating the S-wave arrival with the P-wave arrival.

Event detections within plus or minus 1 second of each other are considered duplicates and their corresponding measurements are merged into a single event. To prevent cycle skipping, we use individual cross-correlation measurements with a minimum NCC value difference of 0.05 between the absolute value of the highest peak and second highest peak. Candidate newly detected events must have at least a total of 20 P and 20 S-wave differential time and cross-correlation measurements to be eligible for location.

Newly detected events are assigned a starting location determined by the median of its corresponding template starting locations. We use the normalized cross-correlation value as the weight for each differential time measurement. For each station and phase,

we select the differential time observation with the highest ‘quality weight’ defined by Shelly et al., (2016). Thus, we include only one differential time and cross-correlation measurement for each station and phase.

$$quality = (0.1 + 3cc_{diff}) + cc_{max}^2$$

(Equation 2.S1)

Here, cc_{max} is the absolute value of the largest interpolate NCC peak, and cc_{diff} is the difference in height between cc_{max} and the absolute value of the second highest NCC peak whether positive or negative. We then set a minimum weight (rmin) of 0.65 to be used for relocation within GrowClust3D (Trugman et al., 2023). The final catalog contains all previously detected and located events from the PNSN catalog, whether successfully relocated by GrowClust3D or not, and new detections that have been relocated by GrowClust3D.

2.8.2) Additional details for preparation of 3D velocity model, topography, travel time grids, and depth datums

We locate earthquakes using 3D P- and S-wave travel-time grids solved from the 3D velocity models of Kiser et al., 2019. The velocity models are originally defined in map view as kilometers in the UTM coordinate system, within a 100 km by 90 km rectangular area centered around Mount St. Helens, and between -3.05 to 14 kilometers below sea level along depth (i.e. 0 sea level datum). The original grid spacing is 1 km along longitude (x), latitude (y), and depth (z). To maintain a consistent reference frame, we project the UTC source coordinates to degrees latitude and longitude according to the WGS84 geodesic.

Here, we describe the steps used to interpolate and apply topographic correction to the Vp and Vs grids for input into Stingray, our travel-time solver. To increase the precision of the travel-time solutions, we define a grid of target coordinates (X_i) in degrees latitude and longitude with a grid spacing of 0.25 km along x, y, and z. The X_i grid target coordinates in map view are defined between the same latitude and longitude boundaries as the original Vp and Vs models and between 0 and 40 km along depth below the topographic surface. This change of depth datum, where 0 is the Earth's surface, is intended to correct for topography. We then apply the `Scipy.interpolate.griddata` function to interpolate the Vp and Vs grids to the target coordinates (X_i) between the topographic surface, to roughly 14 km BSL (i.e. the maximum depth of the Kiser velocity models). For depths below 14 km BSL, we use a linear along-depth gradient of Vp values between 14 to 40 km BSL from the S4 1dimensional velocity model used by the PNSN. Vs values are obtained using a Vp/Vs ratio of 1.73. Both Vp and Vs values from the 1D velocity model are extended horizontally to the longitude and latitude dimensions into the interpolated 3D X_i grids. The resulting grids are then input into Stingray, which returns P- and S-wave travel time grids in the X_i coordinate system as described above. We then convert the travel time grids to NonLinLoc-compatible header (.hdr) and buffer (.buf) file formats, which are also used for relative location solutions in GrowClust3D. In both applications, the depth is with respect to the topographic surface.

To restore the original depth datum, where depth of 0 is sea level, we estimate the corresponding elevation of the epicenter of each earthquake location by interpolating the DEM. The elevation (negative depth) at each epicenter is then added to the depth

solution returned by either NonLinLoc or GrowClust3D. The depth solutions presented through this manuscript and supporting information are the corrected depths following the above procedure. We do not apply correction terms to any step in the location procedure.

2.8.3) Relative location uncertainty estimation

The distributions of single-sided, 1 standard deviation uncertainties are reported in Fig. 2.S5. Individual uncertainties for each event are provided in Supporting Data 2.S1.

GrowClust3D supports uncertainty estimation via bootstrapping (Trugman et al., 2023). To capture uncertainties in all three spatial dimensions, we perform 100 bootstrap iterations for each location. We then estimate the 3D Gaussian uncertainty estimators by computing covariance matrix of the bootstrap iterations for each event.

$$cov[B] = B * B^T = \begin{pmatrix} \sigma_{xx}^2 & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy}^2 & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz}^2 \end{pmatrix}$$

(Equation 2.S2)

B is a 3 by 100 array containing the bootstrap locations with mean subtracted, and standard deviations along X, Y, and depth (Z) are the square root of the diagonal elements. The uncertainties should be viewed with respect to their cluster centroids (Trugman et al., 2017, 2023; Lomax and Henry, 2023). Additionally, precision can vary temporally in accordance to changes in the seismic recording network and intermittent station outages.

2.8.4) Additional details for magnitude estimation

To estimate the size of newly detected events, we apply the relative amplitude procedure defined by Shelly et al., (2016) (Eq. 2.S3).

$$\mathbf{M}_{new} = \mathbf{M}_{template} + c * \log_{10}(\alpha) \quad (\text{Equation 2.S3})$$

The magnitude of a new event ($\mathbf{M}_{new}$) is computed relative to the known template magnitude ($\mathbf{M}_{template}$), the waveform amplitude ratio between the template and new event (α), and a scaling coefficient (c) which relates the differential catalog magnitudes to measurable amplitude ratios (α) (Fig. S9). For each template-event detection, each successful detection is associated with an amplitude ratio (α) from a principal component analysis of the ratio of amplitudes from the new event and template waveforms (see Shelly et al., 2016).

We derive the c coefficient from pairs of catalog events by relating the event-wise differential magnitude and amplitude ratio (α) (Fig. 2.S9) using a robust linear regression. We try three different linear regression solvers including the Random Sample Consensus (RANSAC) algorithm of Fischler and Bolles (1981), the Huber loss algorithm of Huber (1964), and ordinary least-squares. For all three methods, we apply the `linear_model` module of Python Scikit distribution (Pedregosa et al., 2011). In the end, we choose a c value of 0.34, derived from the linear scaling between inter-event difference of newly solved Richter magnitudes ($\mathbf{M}_L$) and $\log_{10}$ of α . We prefer the solution from the Huber robust linear regression algorithm (Fig. 2.S12) because it is deterministic and robust against outliers.

We choose to recompute template starting magnitudes ($\mathbf{M}_{template}$) in order to: derive magnitudes with consistent scaling, reconcile new template location solutions, and improve scaling between amplitude ratio and relative magnitudes (Fig. 2.S12). Most template events in the PNSN catalog are associated with coda duration magnitudes ($\mathbf{M}_d$).

However a fraction of template events are only associated with Richter magnitudes (M_L). For events that have both, we do not find a clear scaling between the M_d and M_L magnitude scales. Furthermore, the scaling between differential M_d and α is poorly constrained—possibly owing to large uncertainties in M_d values.

2.8.5) Frequency Index estimation

To observe trends in spectral characteristics of earthquakes between 2008 to 2024 at Mount St. Helens, we compute the Frequency Index (FI) P-wave arrivals from all 4,044 templates using the definition from Matoza et al., (2014).

$$FI = \log_{10} \frac{\overline{A_{high}}}{\overline{A_{low}}}$$

(Equation 2.S4)

Here $\overline{A_{high}}$ and $\overline{A_{low}}$ are the mean amplitudes of the high and low frequency bands, between 7 to 15 Hz and 1 to 7 Hz, respectively. We compute the amplitude spectra for individual P-wave arrivals recorded on vertical components using the multitaper power spectrum estimation method of Prieto (2022).

We show the distributions of FI values for shallow and deep events in the Figure 2.S13. We show the distributions (2.S13a) of station-averaged FI values as well as a selection individual stations with varying ray paths and site effects (SEP, VALT, SUG, and B202; Figure 2.S13b). In addition, we show the FI values for individual events along depth. There is little difference between the FI distributions of deep events (depth > 2 km BSL) and shallow events (depth < 2 km BSL). Stations SEP and VALT show minor increases in FI with depth. However, borehole station B201 shows a slight decrease in FI

along depth. These trends are probably due to individual path and site effects and cancel each other in the station averaged FI.

2.8.6) Along-depth b-value estimation

We compute b-values using the magnitudes of all events in our catalog. We also compute frequency index (FI) values for templates. Because b-values are dependent on the choice of cutoff magnitude and binning, we compute the b-value using cutoff magnitudes determined by two different methods: (1) the maximum curvature (MAXC) definition of Wiemer and Wyss (2000) and (2) the MBS method of Cao and Gao (2002). For both methods, we use modified code from the Jupyter notebook of Lapins (2022). We additionally perform compute along-depth b-values using two bin widths of 1 and 2 km (Fig. 2.S14).

The along-depth trends in b-value and M_c are inconsistent across bin width and computation method. For the 2-km bin width, the b-values computed by the MAXC method steadily increase along depth between the -2 to 6 km depth bins. However, the b-value curve computed by the MBS method appears to oscillate between those same depths. B-value trends computed with 1-km bin widths are comparatively noisier but show an increase in b-value from the 2 to 5 km depth bins. This demonstrates that the b-value computation, in this application, is quite sensitive both method and binning. For this reason, we choose not to interpret along-depth changes in b-values as an indicator of structural variations in the volcano.

2.8.7) Captions for Videos 2.S1-2.S4

Event locations for 2016, 2022, and 2023 swarms are shown in Videos 2.S2, 2.S3, and 2.S4. Events from entire 2008 to 2023 study period shown in Video 2.S1.

Panels (A, B, and C) are the same from Figure 1. Panel (D) shows the depth time series from Figure 2.4 or Figure 2.2D. Videos 2.S2, 2.S3, and 2.S4 contain frames generated every 3 hours for high resolution. Video 2.1S shows one frame per day to reduce file size. All videos are rendered at 60 frames per second in the MPEG4 format.

2.9.8) Captions for Supporting Data 2.S1

The final earthquake catalog may be accessed through

<https://doi.org/10.5281/zenodo.17089016>.

The earthquake catalog analyzed in this study is uploaded here as a CSV dataframe file. The headers are as follows:

- 'event_id' : unique event identifier.
- 'utc' : event origin time in UTC
- 'lat', 'lon' : event location latitude and longitude in decimal degrees
- 'depth_bsl' : event location depth in km BSL
- 'x_km', 'y_km' : event location in X and Y in km from the geographic origin (Fig. 1A)
- 'rms_P', 'rms_S' : root mean squared misfit for P and S phases
- 'et' : event origin time misfit
- 'type' : whether an event is 'pnsn' or 'new' detection
- 'mag' : event magnitude
- 'sig_x', 'sig_y', 'sig_z' : event location uncertainty in X, Y, and depth (Z) in km

2.9) Supporting Figures

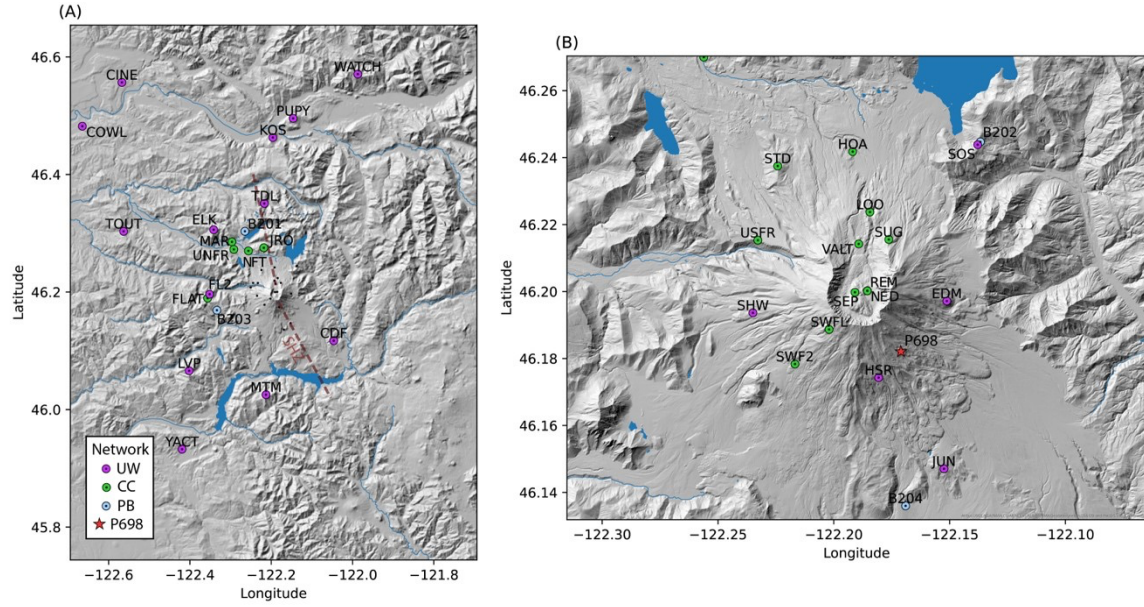


Figure 2.S1. Map of all seismic recording stations and GNSS station P698 used in this study. (A) Extent of the velocity model. The St. Helens Seismic Zone (SHZ) is drawn from Weaver et al, (1987). (B) Stations near Mount St. Helens. Note that some UW stations were moved to the CC network between 2008 and 2024.

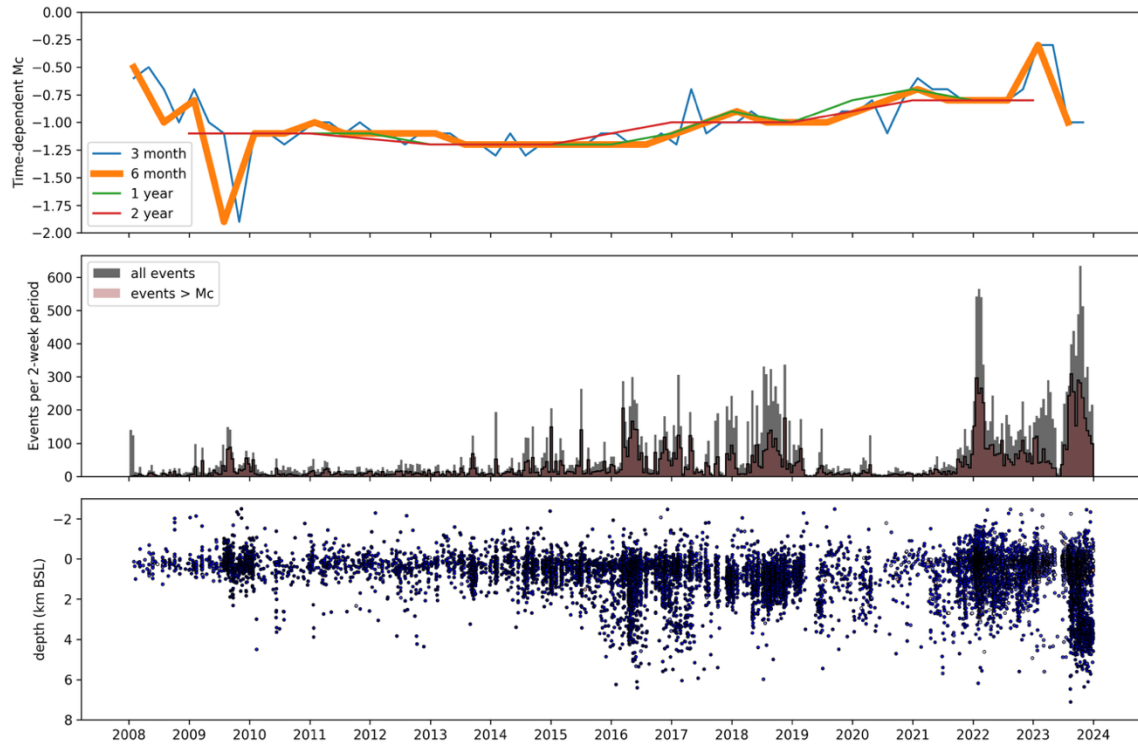


Figure 2.S2: Details for time-dependent magnitude of completeness correction. Time-dependent completeness magnitude (M_c) on top panel, corrected event count

(number of events with magnitude at or above M_c at that time) in middle panel, and all event depths with M greater than or equal M_c on bottom panel. We explore the effect of binning size on M_c by computed within 3 month, 6 month, 1 year, and 2 year-long bin widths. For this analysis, we settle on 6 month long bins, although the effect of bin width is minimal. The uncorrected event rate in the middle panel is the same as shown in Fig. 2.2A.

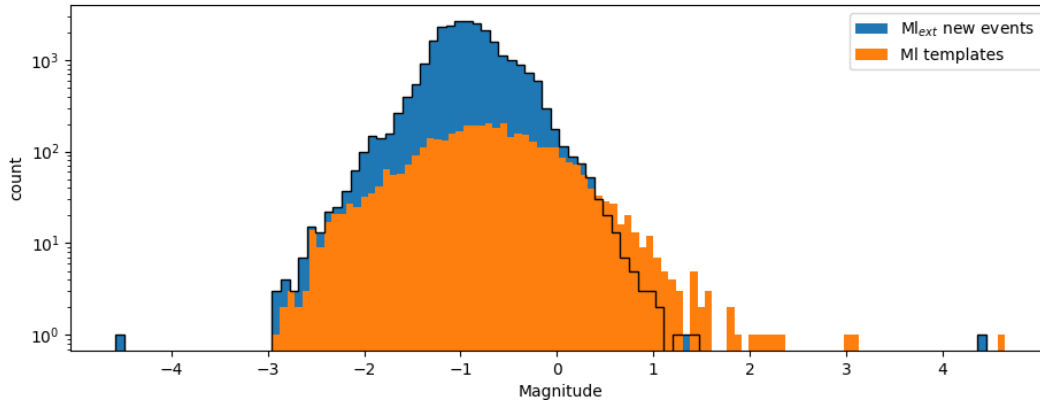


Figure 2.S3: Magnitude distributions for all events and templates. See Section 2.2.3 and Supporting Text 2.S4 for details.

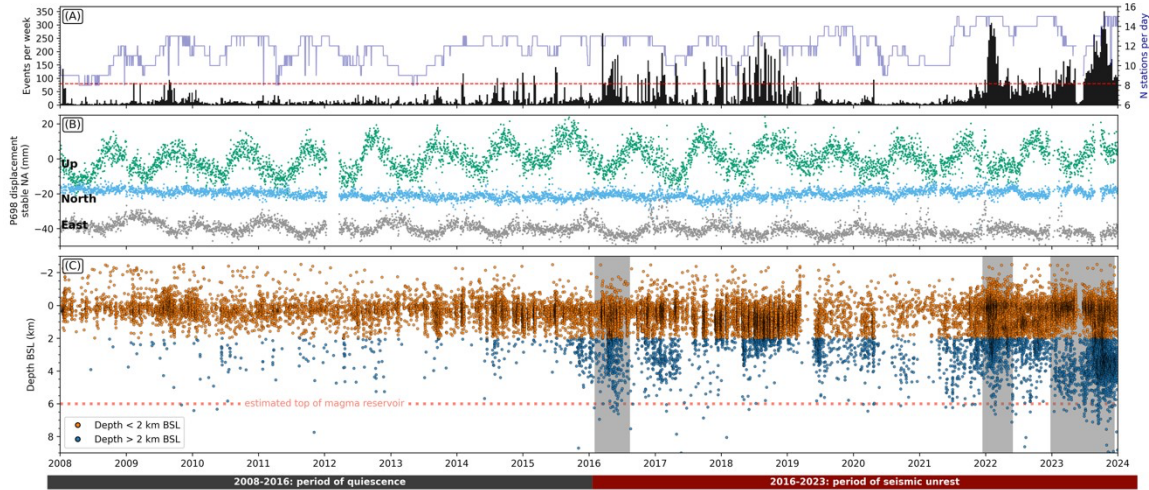


Figure 2.S4: Event rates (A), displacement time series recorded by station P698 in (B) (Fig. 2.S1a), and depths below sea level (BSL) of all locations in (C). Panels A and C are the same as in Figure 2.2 of the manuscript. P698 data access and downloaded through the USGS (<https://earthquake.usgs.gov/monitoring/gps>) and processed using the methods of Murray and Svarc (2017). We choose P698 for its location on the southern volcanic flank of Mount St. Helens and continuous data with few prolonged outages.

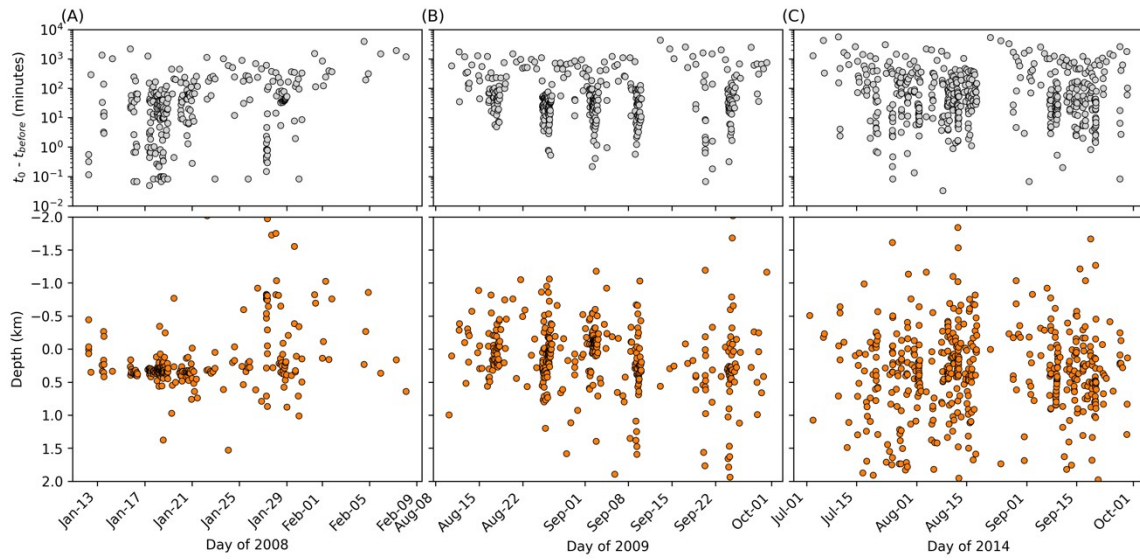


Figure 2.S5. Recurrence intervals and depths (km below sea level) of shallow earthquake swarms from 2008 (A), 2009 (B), and 2014 (C).

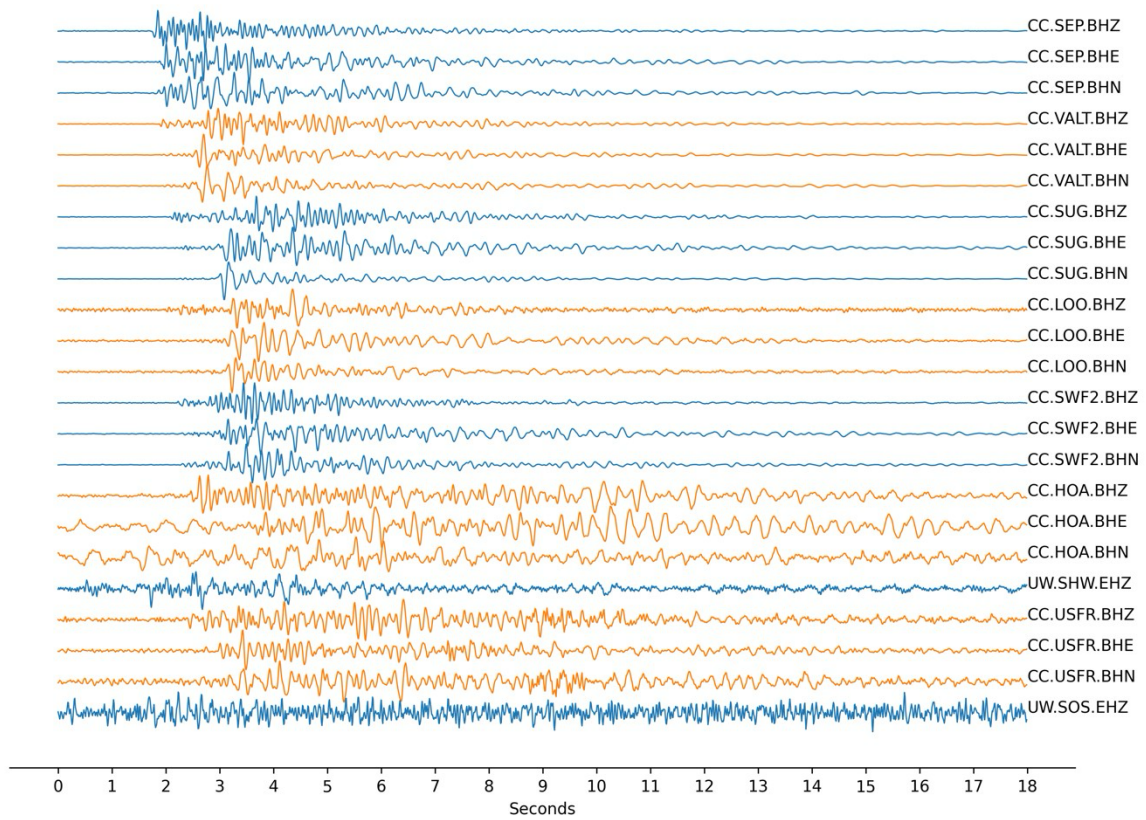


Figure 2.S6 Stacked waveforms for shallow burst-like swarms between February and April 2023, shown in Fig. 2.6. Unlike Fig. 2.6A, this includes vertical and horizontal components.

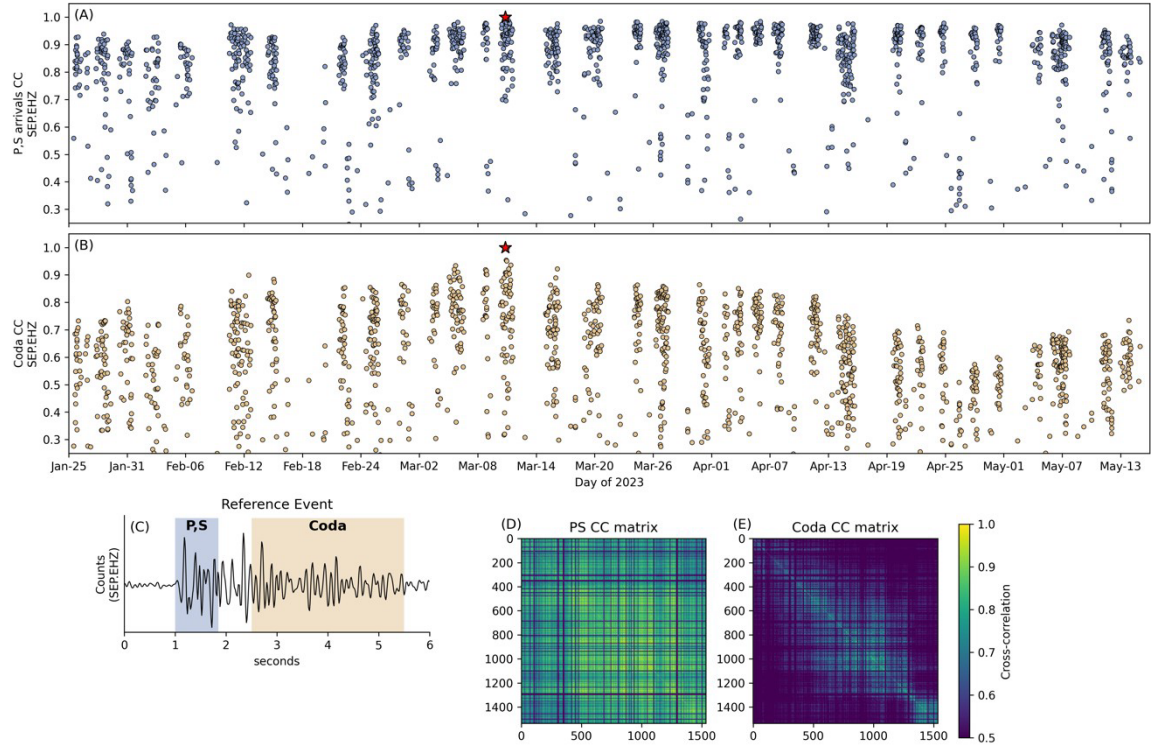


Figure 2.S7. Inter-event P–S (A), and coda (B) waveform cross-correlations with respect to the waveform shown in (C). P–S and coda time windows shown in (C). (D) and (E) show cross-correlation matrices for P–S and coda inter-event cross-correlations.

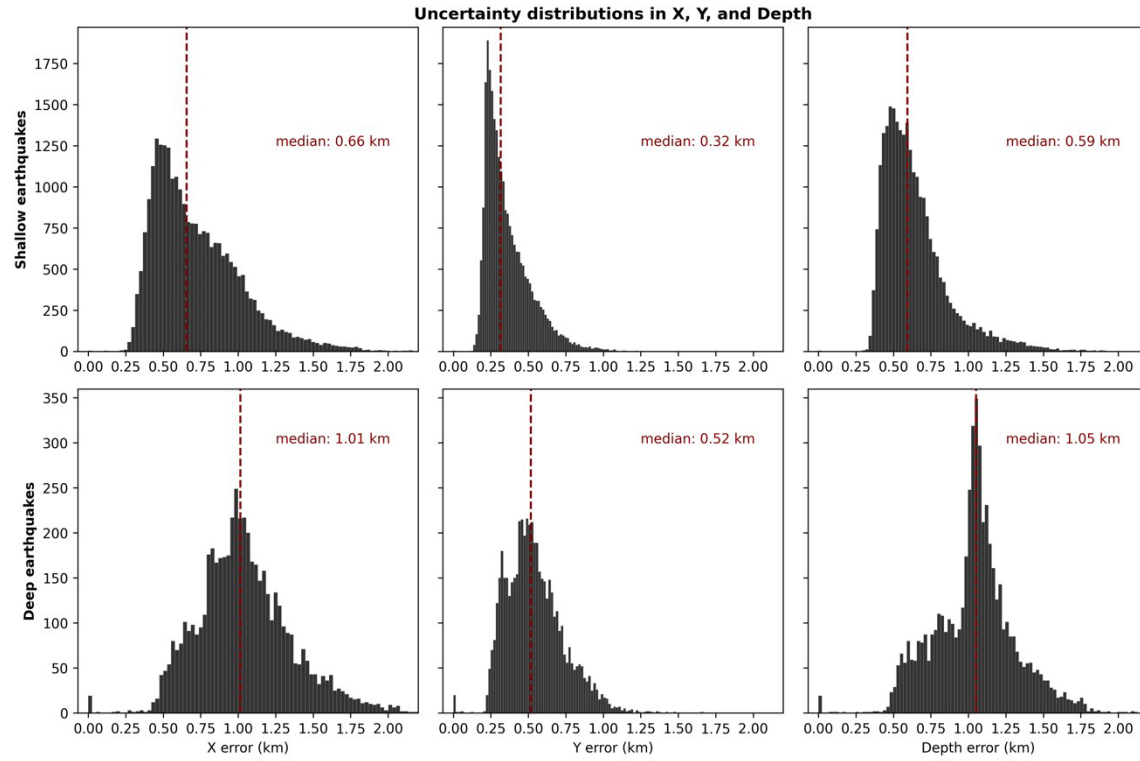


Figure 2.S8. Distributions of single-sided 1 standard deviation earthquake location uncertainties along, X (east-west), Y (north-south), and depth in km. Shallow earthquakes in top row and deep earthquakes in bottom row. See Supporting Methods for description of computation.

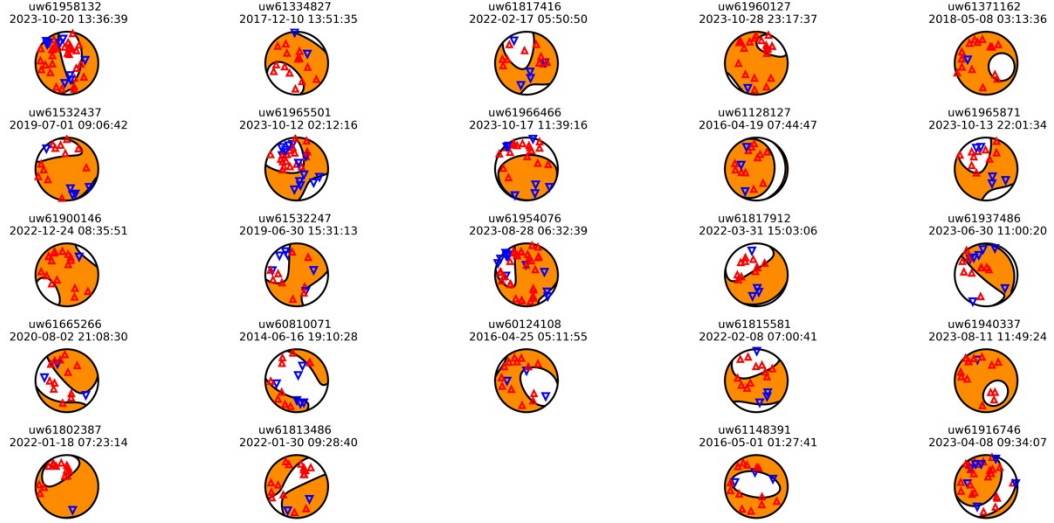


Figure 2.S9. Moment tensor solutions for events with at least 15 analyst-picked polarities. P quadrant is in orange. Positive polarities are red triangles. Negative polarities in blue upside-down triangles. None of these are well constrained.

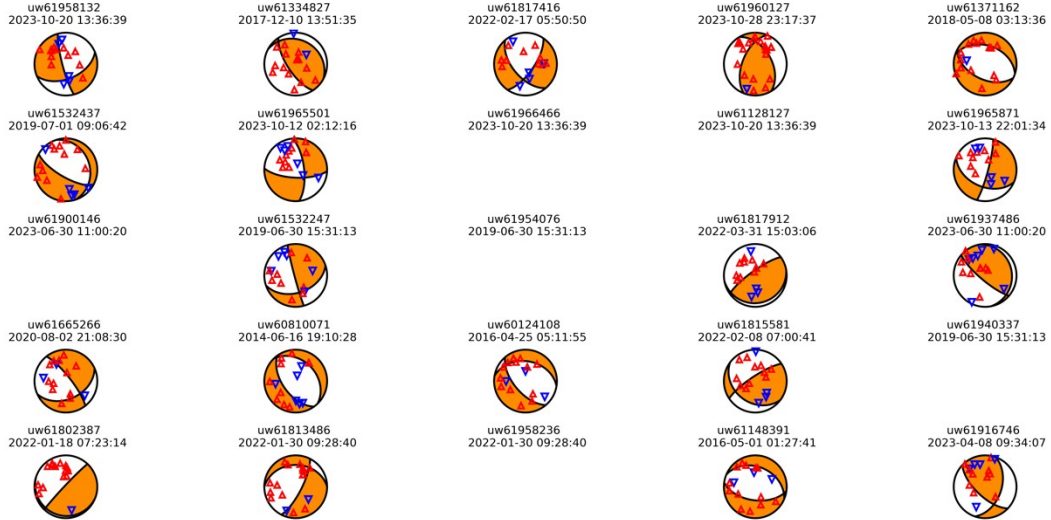


Figure 2.S10. Double-couple fault plane solutions for events with at least 15 analyst-picked polarities. P quadrant is orange. Positive polarities are red triangles. Negative polarities in blue upside-down triangles. If the beach ball is missing, it does not have a nonzero probability solution. None of these are well constrained.

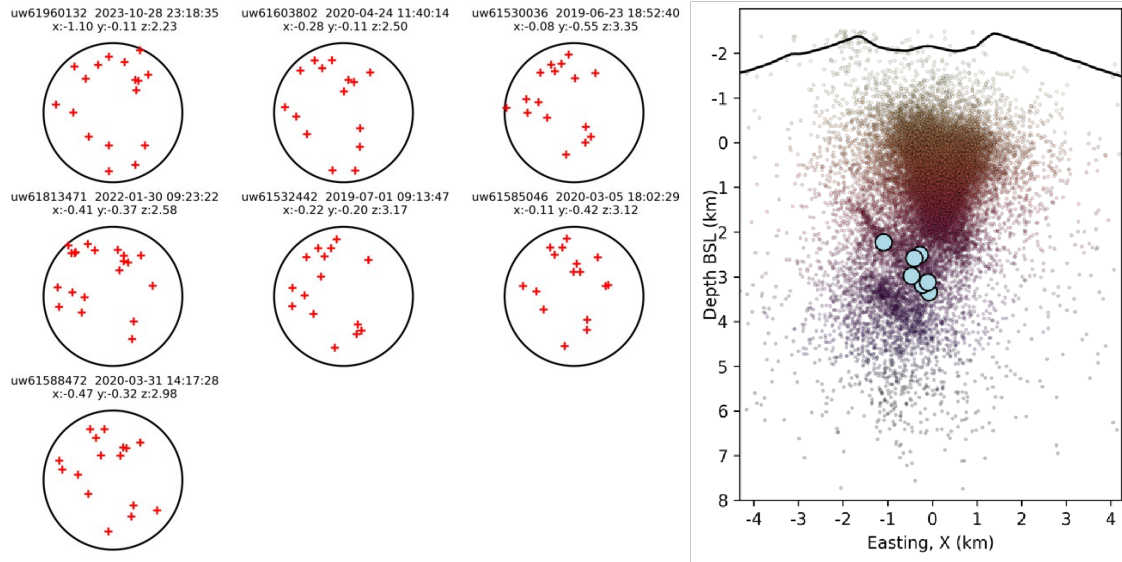


Figure 2.S11. Earthquakes with at least 15 positive analyst-picked P-wave polarities and no negative polarities (left). Location depths in km below sea level (BSL) on right side, plotted in blue circles over all locations in the catalog. Waveforms for the event uw61960132 are plotted in Figure 2.S15.

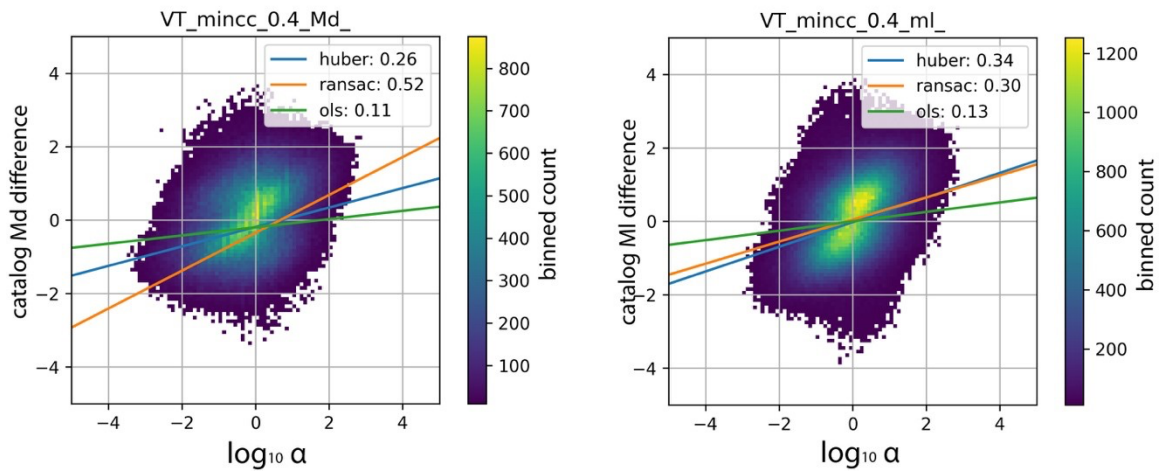


Figure 2.S12. Relative magnitude c coefficient calibration for differential coda duration magnitude (M_d) on left, and differential local seismic magnitude (M_L) on right. For both magnitude scales, the relationship between differential magnitudes and amplitude ratios (α) are noisy. The differential M_L vs α relationship is slightly better constrained. We use a c value of 0.34 determined by the Huber algorithm (right). See Supporting Text 2.4 for details.

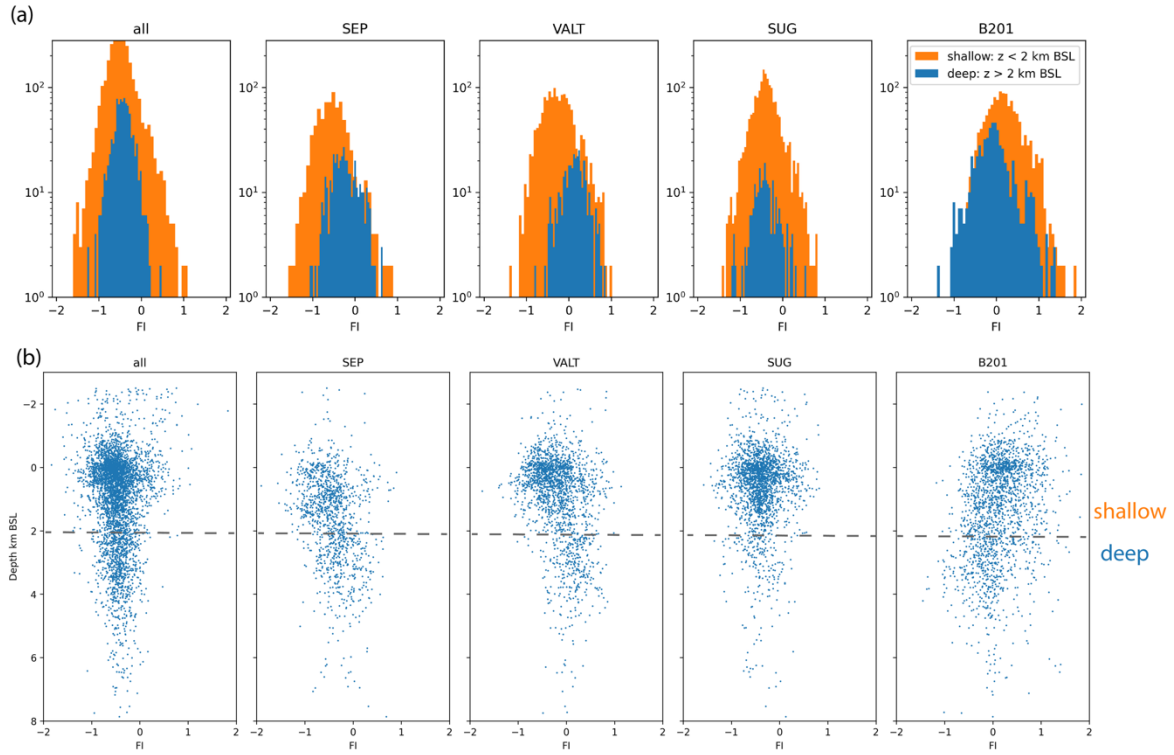


Figure 2.S13. Frequency index (FI) distributions in (a) of station-averaged FI values as well as a selection individual stations with varying ray paths and site effects (SEP, VALT, SUG, and B202). In addition, we show the FI values for individual events along depth in (b). See Supporting Text 2.5 for details.

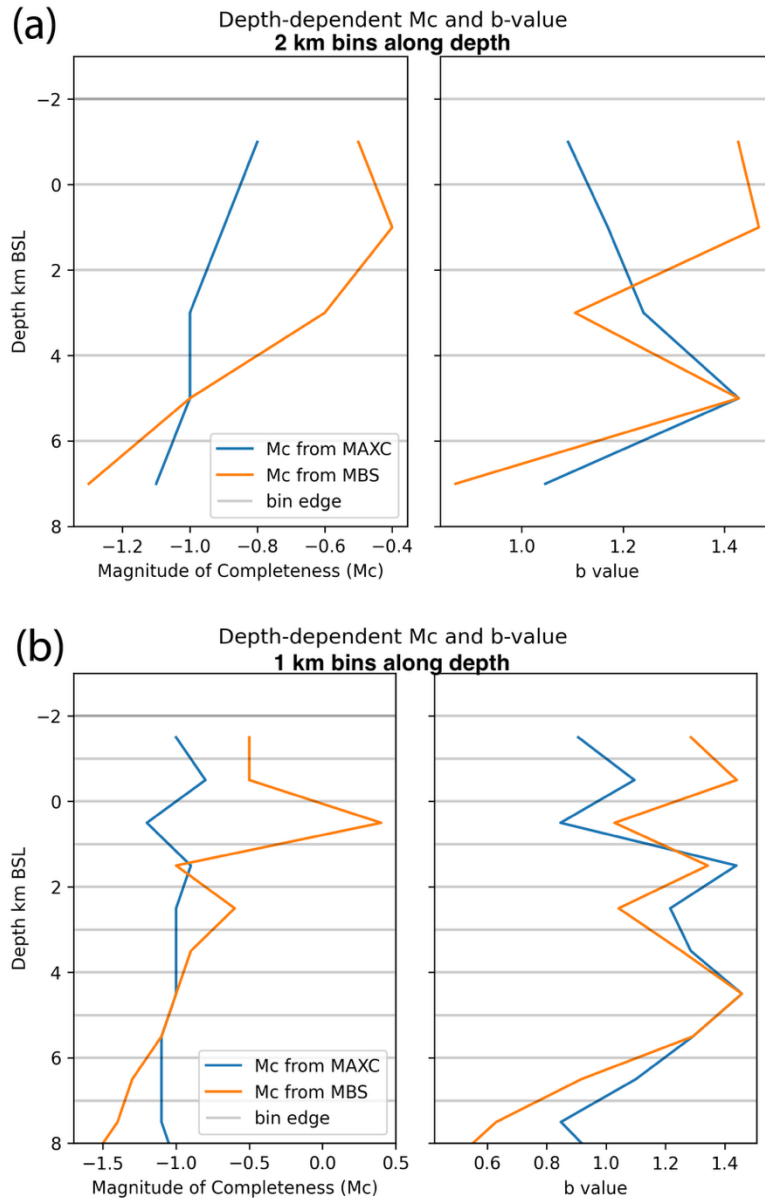


Figure S14. Along-depth profiles of Magnitude of Completeness (Mc) and resulting b-values using (a) 2 km bins and (b) 1 km bins. To compute b-values, we apply two different methods to complete the Magnitude of Completeness cutoff magnitude—the Maximum Curvature of Wiemer (MAXC) and Wyss (2000) and MBS method of Cao and Gao (2002). See Supporting Text 2.6 for details.

2023-10-28T23:18:35.000000Z; uw61960132

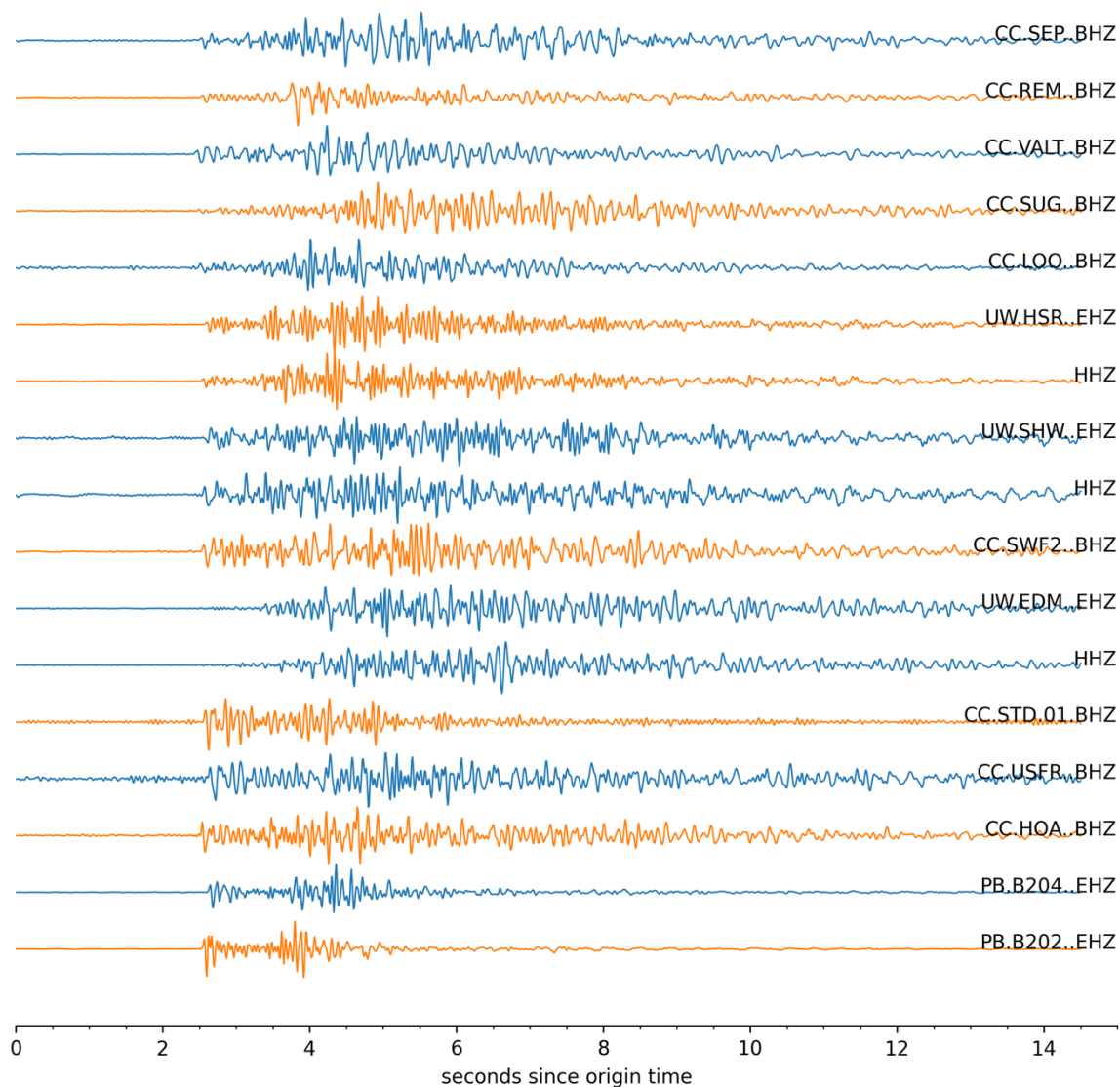


Figure 2.S15. Waveforms of an event (id: uw61960132) in October 2023 with all positive first motion polarities. Plotted are the vertical channels of stations that recorded the event. Note that station UW.SHW.EHZ is inverted. Waveforms are bandpass filtered between 1 to 20 hz and detrended. The traces are arranged from nearest to furthest from the event epicenter. Note the high frequency character and long-duration coda phases on surface stations with the UW and CC network code. Borehole stations (network PB) do not have such long coda phases. See Figure 2.S11 and Figure 2.8 for polarities and locations.

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2.11) Bridge

In this chapter, we apply a precise template matching and location procedure to observe non-eruptive seismic activity between 2008-2024 at the Mount St. Helens volcano in Washington at an unprecedented resolution. Here, we find evidence for repressurization of the upper-crustal magmatic reservoir beginning roughly 8 years after the end of the last eruption. In addition, we find surprising modes of volatile and gas release from the shallow conduit. We infer that volcano-tectonic seismic activity during non-eruptive periods is primarily modulated by varying amount so volatile release from the upper-crustal magmatic reservoir. In the next chapter, we apply similar

detection and location methods to a larger region surrounding Mount St. Helens to analyze seismic activity along the adjacent faulting structures. We then use this information to infer why Mount St. Helens is positioned in its current location. Mount St. Helens is briefly revisited in Chapter 4, where we inspect seismic activity during its most recent eruption.

CHAPTER III

SEISMOTECTONICS OF THE MOUNT ST. HELENS REGION

3.1) Introduction

Seismicity adjacent to volcanic systems can be attributed to the superposition of stresses imposed by the magmatic transport system, local geologic heterogeneities, and the regional-scale tectonic stress field. The Saint Helens Seismic zone (SHZ) is a right-lateral NNW-SSE oriented strike-slip crustal fault that underlies the modern-day Mount St. Helens volcanic complex (e.g. Weaver et al., 1987). The SHZ was initially characterized using fault-plane solutions of events in the February 1981 M5.5 sequence near Elk Lake, WA (Fig. 3.1). Displacement rates of the SHZ at the surface are too low to be detected through Global Navigation Satellite System monitoring. Moreover, there are no known subaerial exposures or geomorphic expressions of the SHZ. Thus, the SHZ remains characterized primarily through seismic studies—including earthquake locations, fault-plane solutions (Grant et al., 1984; Weaver et al., 1987; Qamar et al., 1987), tomographic imaging, and geotechnical surveys.

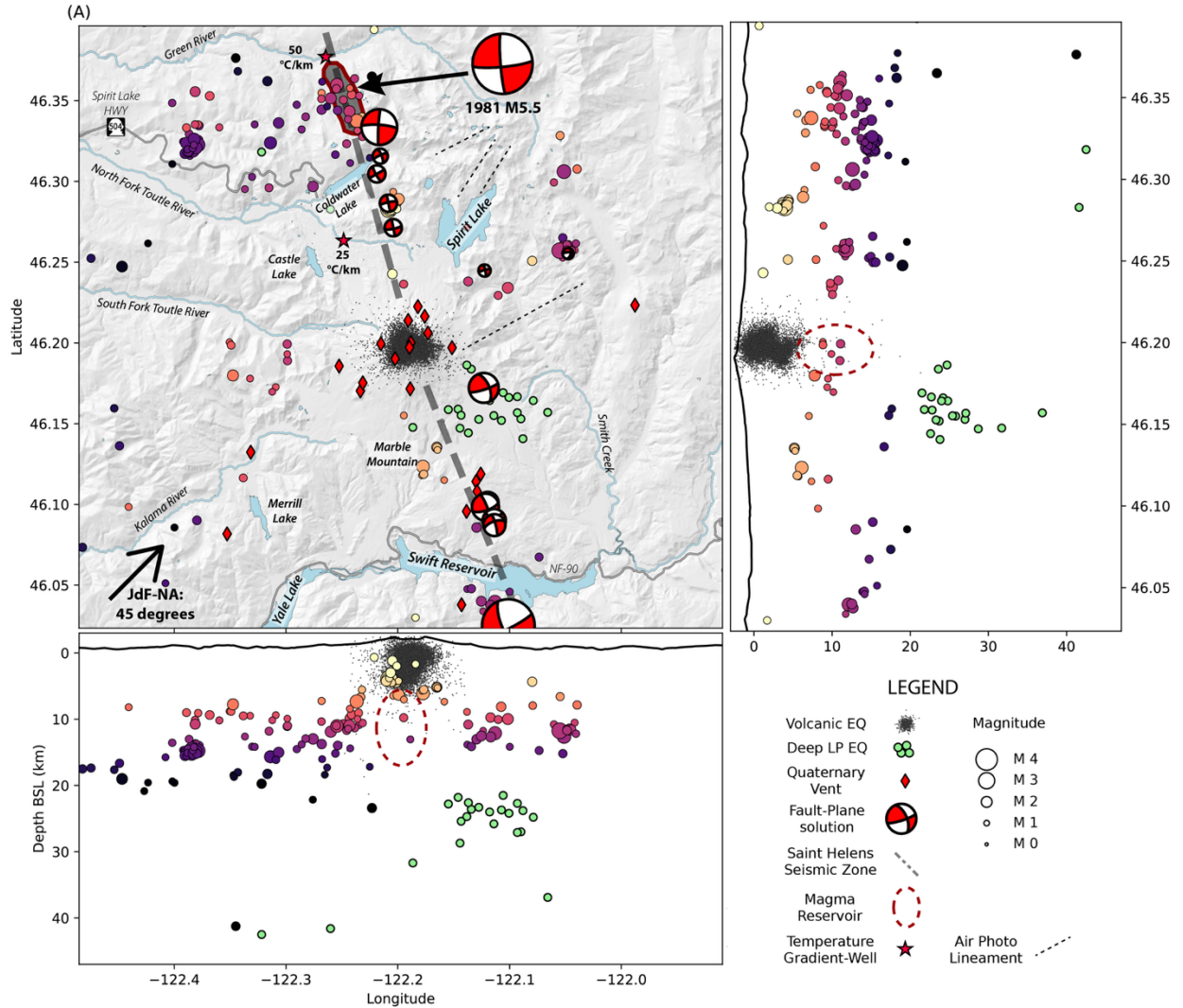


Figure 3.1: Study area with seismicity from the USGS catalog with magnitude greater than M1.5, volcanic earthquakes (EQ) from Hirao et al., 2025, Deep Long Period (LP) events from the PNSN catalog, Quaternary volcanic vents from O'Hara et al. (2020), historical fault-plane solutions from 1981-1985 and the Saint Helens Seismic Zone trace from Weaver et al., 1986, and approximate upper-crustal magma reservoir location from Kiser et al., 2018. Temperature gradient well data from Norman et al. (2015). Juan de Fuca (JdF) and North American (NA) convergence vector is sourced from the ITRF2020 model (Altamimi et al., 2023). Air Photo Lineaments reflect pyrite-bearing dikes and are simplified from Evarts et al., 1987.

The Saint Helens Seismic Zone may partially explain the location of the modern-day Mount St. Helens volcano. Mount St. Helens is unique from other Cascade volcanoes in southern Washington and northern Oregon for its high background earthquake activity rates (e.g. Hirao et

al., 2025), and for its anomalously large west-ward offset from the main volcanic axis (e.g. Hildreth, 2007; Hansen et al., 2016; Bedrosian et al., 2018). Magnetotelluric and seismic imaging of central and southern Washington indicate that the SHZ at midcrustal depths forms a lithologic boundary that separates the Siletzia oceanic basalts and Spirit Lake Batholith granodiorites to the west and east, respectively (Hill et al., 2010; Bedrosian et al., 2018; Kiser et al., 2019; Ulberg et al., 2020). In addition, observations of low Vp anomalies, high conductivity, persistent deep long-period event activity (e.g. Han et al., 2018), and elevated fluid geothermal observations taken at Green River (Norman et al., 2015) suggest that the SHZ may serve to channel melt and volatiles from midcrustal to shallow depths (Hill et al., 2010; Bedrosian et al., 2018). A slight offset of strike of the SHZ fault is thought to form a small pull-apart extension beneath Mount St. Helens volcano, allowing for decompression melting and the upward movement of melt towards shallow depths (Weaver et al., 1987).

Earthquake locations, timing, and their associated fault-plane solutions can be helpful to characterize deformation along active structures. To understand the potential for interaction between faulting and magmatic systems, we generate two enhanced seismicity catalogs—both consisting of earthquakes surrounding the volcano. The first catalog spans between January 1, 2008 and December 31, 2024 and consists of earthquakes recorded on a total of 59 permanent broadband seismometers within roughly 50 km of the modern-day Mount St. Helens volcanic edifice. The second catalog is shorter in duration, spans between July 2014 and July 2016, and consists of earthquakes that are recorded on the permanent network in addition to 65 broadband nodal seismometers included in the Imaging Under Mount St. Helens (iMUSH) dense array experiment (e.g. Kiser et al., 2016). Compared with the first catalog, the second catalog contains fewer events in total; however, the dense network configuration of iMUSH stations should allow

for the detection and location of larger quantities of small magnitude earthquakes during the 2014 to 2016 period. In addition, the configuration of the iMUSH network is well suited to detecting earthquakes with good azimuthal coverage—which we leverage to solve for focal mechanisms to improve the characterization of the Saint Helens seismic zone and adjacent faults.

3.2) Methods

3.2.1) Detection, location, and magnitude

To generate both earthquake catalogs, we apply the shift-and-stack template-matching detection and location procedure described in Hirao et al. (2025), which is briefly summarized here. We generate individual P- and S-wave waveform templates from analyst-verified earthquakes from the event catalog distributed by the Pacific Northwest Seismograph Network (PNSN; University of Washington (1963)). We cross correlate each waveform template through continuous day-long data streams at 1/100 samples per second using the Super-Efficient Cross Correlation package of Senobari et al., (2018)) and consider event detections where peak values of the summed normalized cross-correlation (NCC) value exceed 12 times its median absolute deviation (MAD), for that day. Cross-correlation peak values and timing for individual traces and phases within 0.5 seconds of the shifted differential times are stored if they exceed 8 times their respective MAD for that day. The resulting differential times and NCC values are then input into the GrowClust3D algorithm (Trugman et al., 2023) for relative location. We localize both newly detected and previously catalogued template events using three-dimensional P and S-wave travel time grids from the 3D velocity models of Kiser et al. (2019).

To quantify the size of newly detected earthquakes, we extend the coda duration magnitudes (M_d) included in the PNSN catalog, to new detections using the relative amplitude scaling procedure of Shelly et al., 2016 (Equation 3.1).

$$M_{new} = M_{template} + c \log_{10} \alpha$$

(Equation 3.1)

Here, α represents the amplitude ratio between a given template and new event waveform and c is a scaling constant that maps the measured amplitude ratio (α) to a magnitude (M_d) differential. To determine the c value, we compute the amplitude ratio for each catalog template pairwise against their respective catalog M_d difference. We then empirically determine a value of $c = 0.6$ by using the Huber robust linear least squares algorithm included in the SciKit Learn package (Fig. 3.4; Pedregosa et al., 2011). Values of α are computed at maximum waveform alignment, through an eigenvector decomposition of the covariance matrix of the template and new event waveforms.

We favor the M_d values as starting magnitudes because most of templates (3,001 of 3,171) are assigned M_d values whereas only 1,752 events that are assigned local magnitudes (M_l). For the remaining 170 event templates with M_l values without M_d , we estimate the equivalent M_d using a linear scaling relationship derived from a total of 1,249 events that have both M_d and M_l values (Equation 3.3).

$$M_d = 0.876 * M_l + 0.230$$

(Equation 3.2)

3.2.2) Long-term catalog construction

To construct the long-term catalog between 2008 and 2024, we apply the template matching procedure to stations of the regional network—including a total of 59 broadband seismometers in the PB, UW, CC, PB, and UO networks within the 100 by 90 km region

around MSH. We construct P and S-waveform templates from 3,171 analyst-picked and verified events contained in the PNSN catalog, recorded between January 2008 to December 2023. All templates and new detections analyzed in this catalog are at least 4 km away from the center of the present-day MSH volcanic edifice.

3.2.3) Methods for short-term (iMUSH/XD) catalog

To construct the short-term catalog (also referred to as iMUSH or XD catalog), we apply the template-matching and location procedure described above to a total of 65 stations from the iMUSH dense array experiment, in addition to all stations from the permanent network that recorded from July 2014 to July 2016 (XD network). The Imaging Under Mount St. Helens (iMUSH) dense array experiment is a temporary geophone and seismometer network (network code XD) that continuously recorded ground motion in the region surrounding Mount St. Helens region between July 2014 through August 2016 (e.g. Kiser et al., 2016). Although originally designed to produce high-resolution tomographic models of the MSH volcanic complex after its latest eruption (ending in 2008; e.g. Dzurisin, 2015; 2018), its dense and widespread configuration is also favorable for capturing microseismicity distal to the MSH volcano where permanent network coverage is poor (especially east and south of the MSH edifice; Fig. 3.2)). Here, the detection and location procedures are performed separately from the long-term catalog.

Similar to the long-term catalog, we construct separate waveform templates from P- and S-waveform arrivals of 540 PNSN catalogued events recorded between July, 2014 to August, 2016. Each event template includes analyst-verified phase picks on permanent network stations in addition to arrivals recorded on the XD network. P- and S-wave phase arrivals on XD network stations are identified by a combination of manual and automated phase picking with the PhaseNet convolutional neural network (Ronneberger et al., 2015; Zhu and Beroza, 2019). We

apply the PhaseNet algorithm with the STEAD-trained model weights included in the Seisbench package (Woollam et al., 2022) to a 60 second window beginning at the known event origin time to all XD stations within a 50 km radius of its epicenter. We save all P- and S-wave phase arrival picks with a minimum confidence of 0.3. We then manually inspect the arrivals of each event, discard mis-timed and misclassified arrivals, and manually impute additional missing arrivals using the Snuffler graphical user included in the Pyrocko distribution (Heimann et al., 2017). We use these phase picks to solve for starting locations to templates with the NonLinLoc package.

3.2.4) Fault-plane solutions

Fault plane solutions (FPS, a.k.a. focal mechanisms) are a convenient way to represent the orientation of slip along a faulting surface during an earthquake with an assumed double-couple point source (e.g. Aki and Richards, incl. 1980; Vavrycuk, 2015). We make further use of the station coverage offered by the XD network to resolve fault-plane solutions for PNSN catalogued events recorded between 2014 to 2016. We manually assign polarity values between -1 (negative) and 1 (positive) through visual inspection of all first-arrival P values on vertical channels. Each polarity is a value between -1 (dilatation, down) and 1 (compression, up) and is weighted to reflect the subjective measurement confidence. We also include the first motion polarities assigned by the PSNS analysts on permanent network stations (IRIS DMC, University of Washington, 1963), which are either 1 or -1 for compression and dilatation, respectively.

To further aid in constraining FPS's, we include S/P amplitude ratios for all vertical channels that include a P-wave arrival time (Hardebeck and Shearer, 2003). If an S-wave arrival pick is not present, we crudely predict the S-wave arrival with an S to P travel-time ratio of 1.73. The peak-to-peak amplitude for each phase is then computed as the maximum value minus the

minimum values of the waveform for a 2.5 second window, beginning at 1 second before its respective arrival time pick. We take the S/P amplitude ratio as simply the ratio of these two values for S and P phases and assume that the noise levels prior to both arrival times are the same. Next, we estimate ray path take-off angles and azimuths for polarity and amplitude ratio observations using the NonLinLoc Time2EQ program and 3D P-wave travel time grids and event location from GrowClust3D. We require that each event be associated with a minimum of 8 polarity observations and a maximum azimuthal gap of 90 degrees and assume that 10% of both polarity and S/P amplitude ratio observations are incorrect. Lastly, we input into SKHASH a list of polarity reversals reported by the PNSN to account for stations with inverted vertical channel orientations. The first motion polarities and S/P amplitude ratios are then jointly inverted for the optimal FPS using the method of Skoumal, Hardebeck, and Shearer (SKHASH, 2024; Hardebeck and Shearer, 2002; 2003). Here, we report FPS for a total of 181 events recorded between 2014 to 2016. To be selected for the result, we require a maximum nodal plane uncertainty of 25 degrees and a quality score of ‘A’ or ‘B’ (see manual of SKHASH, 2024).

3.2.5) USGS-sourced moment tensors

For discussion, we include from the USGS catalog, the moment tensor solutions for two of the largest earthquakes recorded within the study period: a Md 4.3 (Mw 3.84) event of February 14, 2011 and Ml 3.9 (Mw 3.76) event of January 3, 2018 (Fig. 3.3). The Md 4.3 event of February 14, 2011 is located roughly 7 km NNW of the Mount St. Helens edifice, at a depth of 4 km BSL. This event is fit by only a 33 percent double-couple component and a steeply dipping NNE-SSW oriented primary fault plane with a strike of 256 degrees azimuth. The Ml 3.9 event of January 3, 2018 is located roughly 10 km ENE of the Mount St. Helens edifice and 10 km depth BSL. It is fit completely by a double-couple mechanism, NNE-SSW oriented normal-

faulting auxiliary planes (strike of 55 degrees azimuth and dip of 40 degrees). Because moment tensor solutions are sensitive to location parameters, we assume the location provided by the USGS catalog when discussing the two events.

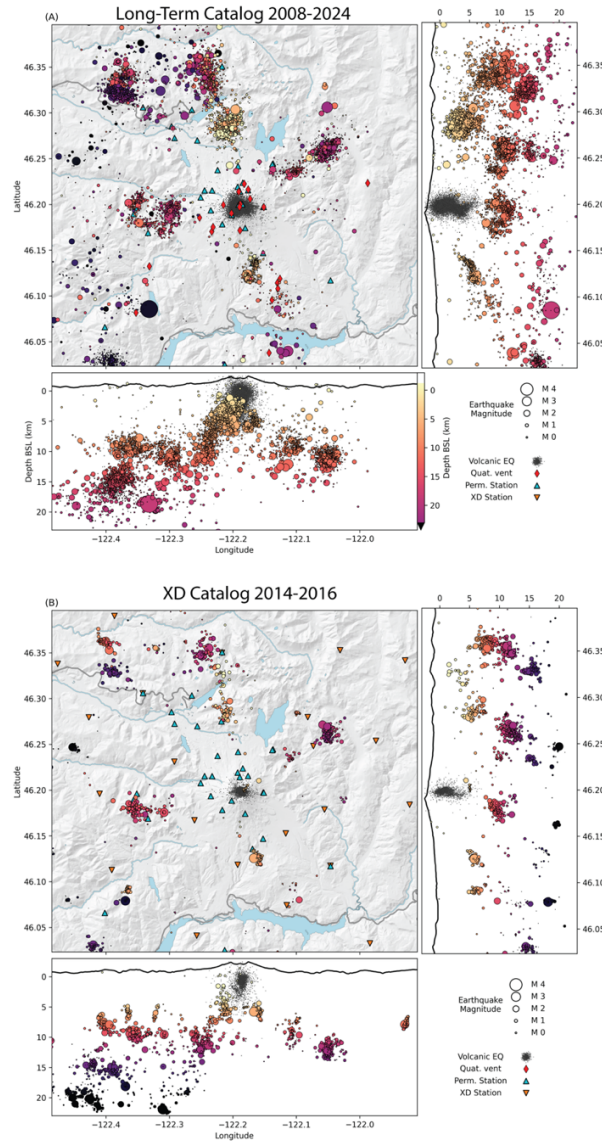


Figure 3.2: Locations of events in the Long-Term catalog from 2008 to 2024 in (A), and XD (iMUSH) catalog from 2014-2016 in (B).

3.3) Results

Between 2008 and 2024 we record a total of 10,331 earthquakes (approximately 3 times greater than in the PNSN catalog) within a region approximately 5 to 50 km of the MSH

volcanic edifice frequent seismicity in the region surrounding MSH between 5 to 30 km of the modern-day edifice (Fig 2, 3). Earthquakes directly beneath the volcanic edifice between 2008 to 2024 are analyzed in Hirao et al., (2025) and are only considered for reference in this study (Fig. 3.1, 3.2, 3.3). Approximately 65% (6,696 of the 10,331) of all observed events adjacent to the MSH volcano are organized into four geographically distinct groups that are positioned to the NNW, SSE, WSW, and ENE of the edifice (Fig. 3.2, 3.3).

For spatial and temporal analysis, we isolate and refer to these four groups of earthquakes according to their position relative to the middle of the MSH edifice at a latitude and longitude of 46.19 and -122.19 degrees and consider the rest as a part of the geographic ‘background’ (Figure 3.2, 3.3, 3.4). In addition, we treat the NNW—SSE and WSW—ENE groups as antipodal pairs with respect to the middle of the MSH edifice for additional spatial analysis. Here we seek to compare the relative positions of each cluster in both pairs not only with respect to each other, but in addition to the MSH edifice. To do so, we project the distance of the two pairs along two curvilinear profiles: A-A’ and B-B’, respectively (Figure 3.3). Both profiles are defined by three points and intersect that the middle of the MSH edifice, which we treat as the origin. We use three points for each profile instead of two because antipodal pairs cannot be adequately fit by a straight line defined by two points that also intersect at the MSH edifice.

To characterize spatial patterns of seismic energy release, we combine the earthquakes of the long-term catalog from 2008 to July 2014, the XD catalog from July 2014 August 2016, and the long-term catalog thereafter. We then separate them geographically into 0.1 by 0.1 km² bins and then sum the cumulative seismic moment. For each event, we solve for the seismic moment M_0 in Equation 3. Here we assume that M_0 is in units of newton-meters and that moment magnitude M_w is represented by our extended coda-wave magnitude defined in the Methods section.

$$M_w = \frac{2}{3} \log_{10} M_o - 6.07$$

(Equation 3.3)

In addition, we define a magnitude of ‘completeness’ (M_c) to describe the lowest event magnitude that is completely detected by our methods. For each catalog and each seismicity group, we bin all event magnitudes into 0.1 magnitude bins, then assume that M_c is the magnitude bin with the highest count (i.e. the ‘peak’ of the magnitude distribution). We consider this simple M_c adequate for comparing detection thresholds between network geometries used to construct each catalog.

The geographic footprint of the NNW and SSE seismicity groups is consistent with the orientation and extent of the SHZ (Fig 3.1), previously defined by Weaver et al. (1987). The NNW seismicity group is the most populous of all four and contains a total of 2,952 events within a continuous, NNW-oriented quasi-linear feature. The southernmost extent is near the head of the North Fork Toutle River at the Pumice Planes, extends NNW beneath Coldwater Lake, and reaches its northernmost extent at the Green River, between roughly 5.5 to 28 km NNW of the MSH edifice. The SSE groups a total of contains 794 events (about 3.7 times fewer than in the NNW) and spans from its northernmost end at about 5 km south of the MSH edifice and forms a trend that deviates from the NNW group towards the NW-SE. Here, it extends beneath a series of Quaternary volcanic vents (O’Hara et al., 2019) southwards towards the eastern end of Swift Reservoir. The SSE cluster is less spatially continuous than the NNW cluster, owing to its smaller event count. While earthquakes within the NNW cluster occupy a greater range of depths than in the SSE cluster, they share similar median depths of 4.2 and 5.3 km BSL, respectively.

The WSW and ENE seismicity groups are positioned roughly opposite each other on either side of the MSH edifice, and each is separated from seismicity directly beneath the edifice by approximately 5 km (Fig. 3.2, Fig. 3.3). The WSW group is bounded to the north and south between the South Fork Toutle River and Kalama River, respectively, between 5 to 15 km west of the MSH edifice. The ENE cluster is separate from earthquakes within the edifice by a similar distance and forms ENE-striking lineation southeast of Swift Lake that extends from 5 to 18 km away from the edifice. The WSW and ENE clusters both occupy similar depth ranges between 7 to 13 km BSL with median depths of 10.3 and 11.3 km, respectively. We record a total of 1,966 earthquakes within the ENE cluster, about twice as many as in the WSW cluster.

All four seismicity groups feature persistent activity within the 2008 to 2024 study period (Fig. 3.3). Normalized binned seismicity rates for all four groups are shown in Figure 3.3. Here, we show rates of all four groups relative to each by bin the seismicity times into their own non-overlapping two month-long time bins. We then normalize each time-series with respect to their own maximum rates. For reference, we include the seismicity rates of volcano tectonic events from the MSH edifice. Peak activity rates for all groups occur between the years 2009 to 2018. Unsurprisingly, local peak seismicity rates within each of the four seismicity groups are temporally correlated to their largest magnitude events.

The largest event recorded in this study period is a M_l 4.3 in the NNW group between Spirit Lake and Coldwater Lake on February 14, 2011. The second highest magnitude event, also the largest event in the ENE group, is a M_l 3.9 east-southeast of Spirit Lake on January 3, 2018. Both earthquakes have been assigned moment tensor origin products by the USGS and will be briefly discussed in a later paragraph. The largest magnitude earthquake of the SSE and WSW

groups are a M_L 2.9 near Marble Mountain on February 8, 2016, and a M_L 2.2 near the South Fork Toutle River on July 9, 2013.

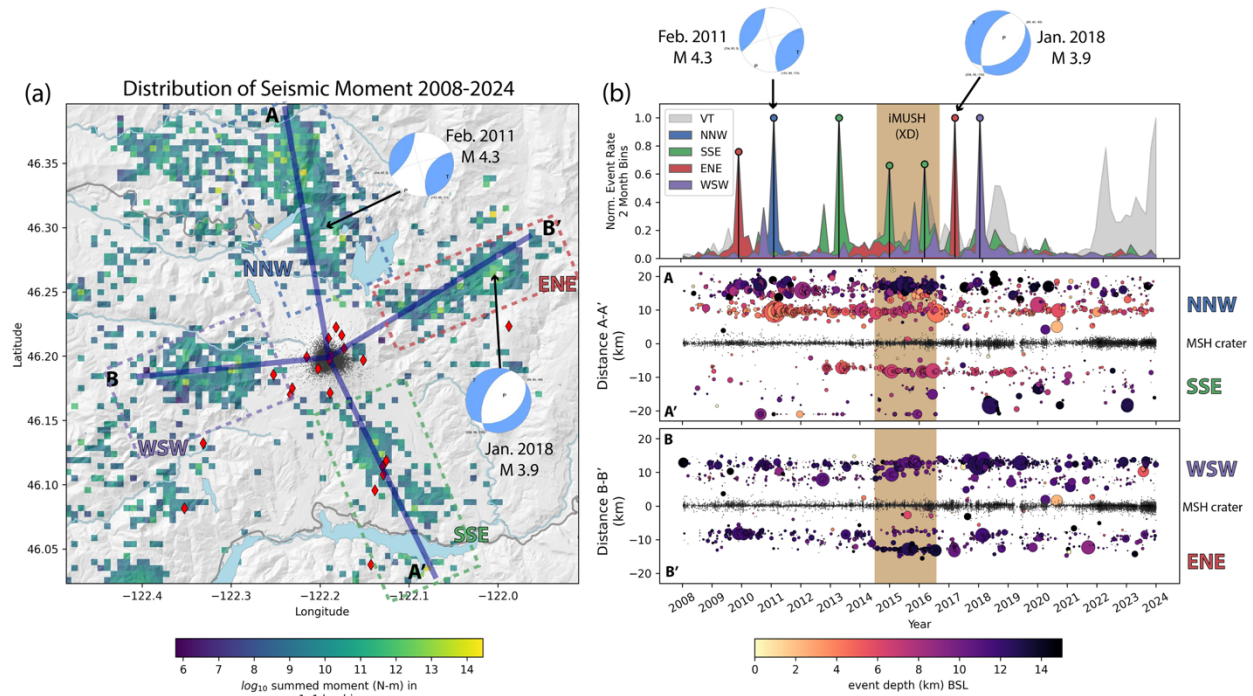


Figure 3.3: The four seismicity groups from 2008 to 2024. (a) Distribution of seismic moment from 2008 to 2024 in 1x1 km bins, A-A' and B-B' profiles used to analyze spatial distributions of all four clusters. Beachballs of moment tensor solutions of two of the largest events from the USGS catalog from from the 2008 to 2024 study period. (b) Normalized event rates in 2 month-long bins for all four seismicity clusters in addition to VT seismicity from Hirao et al. (2015), along-profile distances for the NNW and SSE clusters along the A-A' profile and WSW and ENE clusters along B-B' profile.

We do not find any clear correlation for elevated seismicity rates between the four groups, nor from within the volcanic edifice at any time. In addition, seismicity within the respective groups does not appear to migrate to any other group, or in/out of the MSH edifice. Thus, there is no clear evidence for triggering or interaction among seismicity groups or the volcanic system.

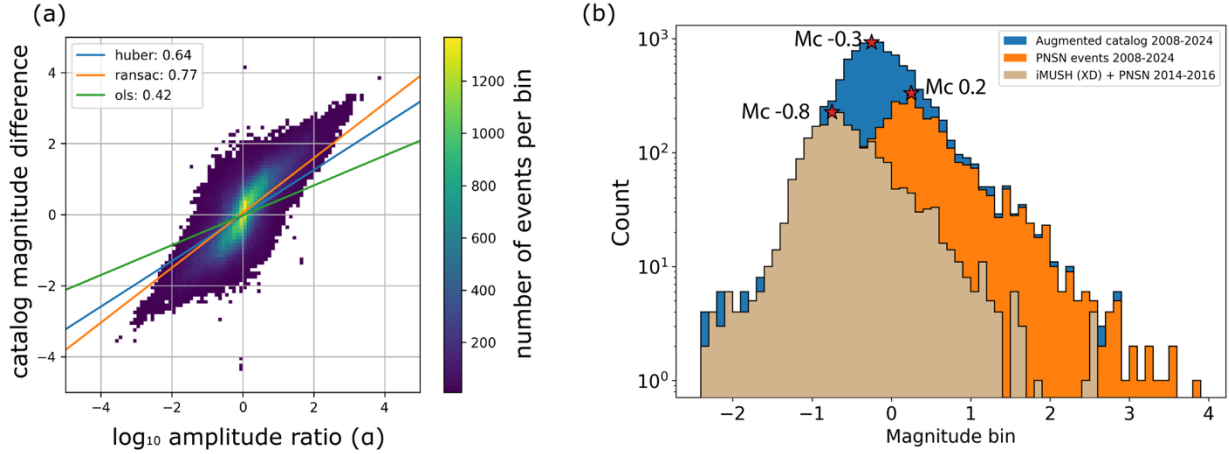


Figure 3.4: Catalog magnitude difference and $\log_{10}$ amplitude ratio scaling in (a). We adopt the Huber solution of $c = 0.64$ to solve for relative magnitudes. Magnitude frequency distributions in (b) for PNSN catalogued events 2008-2024, the long-term augmented catalog 2008 and 2024, and the iMUSH XD augmented catalog 2014-2016. The M_c for each distribution is assumed as the peak magnitude bin according to the Maximum Curvature definition of Wiemer and Wyss (2000).

The amplified detection capacity of the permanent network by including 65 stations from the temporary XD network permits us to detect and locate a total of 2,236 events between July 1, 2014 and July 31, 2016—roughly 1.6 times more numerous than the long-term catalog in the same period. The improved station coverage enables detection and location of a greater number of small magnitude earthquakes, which is made apparent by a M_c of -0.8 in the XD catalog, as opposed to -0.3 in the long-term catalog (Fig. 3.4). In addition, the improved network allows for earthquakes to be more precisely located across the entire study region. The distributions of median absolute deviation uncertainties for all four seismicity groups are shown in Figure 3.5. Median horizontal and depth errors for the long-term catalog range from 0.4 km to 0.7 km and 0.4 to 0.8 km, respectively. By comparison, the median horizontal and depth uncertainties for events in the XD catalog for those same groups range from 0.1 to 0.3 km and 0.1 to 0.5 km—two to five times smaller than in the long-term catalog.

We report the fault-plane solutions for a total of 181 events (Fig. 6), which have a solution quality score of A or B, recorded by the XD and permanent networks between July 2014 to August 2016. Without augmentation from the XD network, few events have adequate focal sphere coverage to reliably constrain a fault-plane solution.

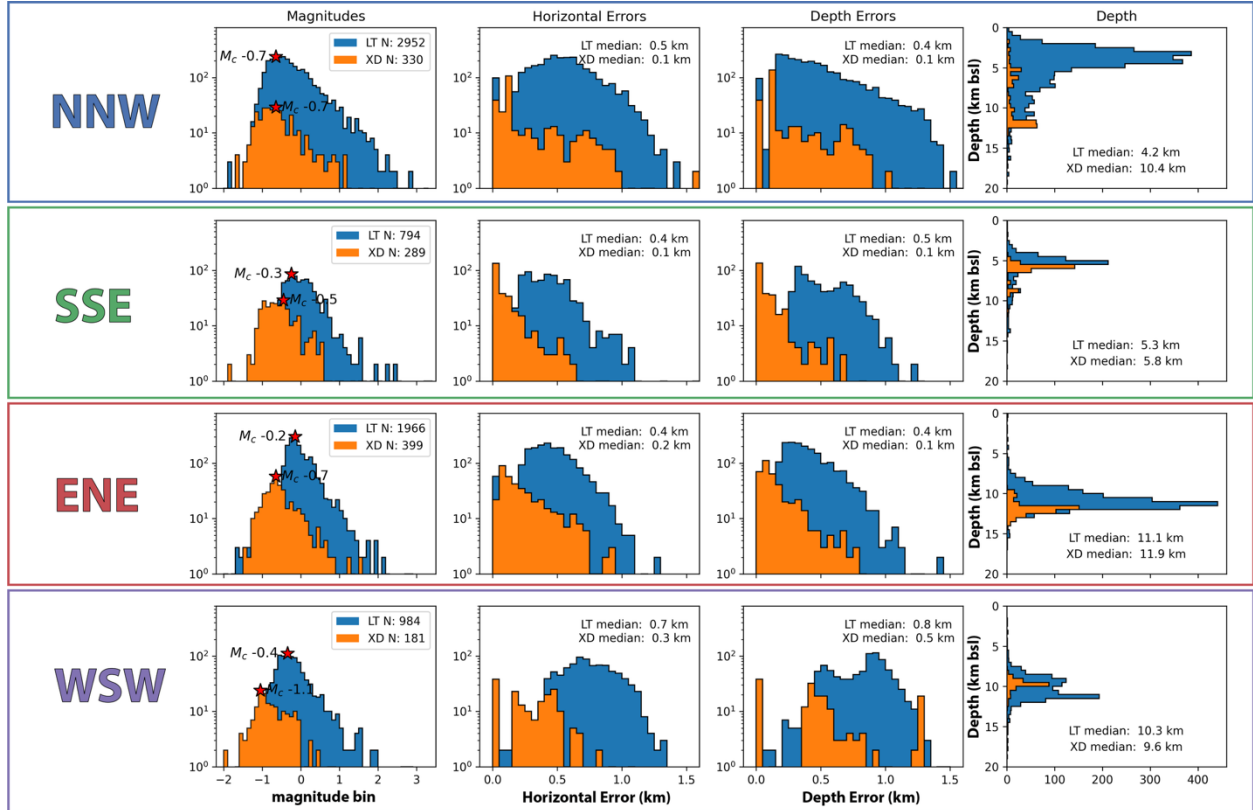


Figure 3.5: Location statistics for all four seismicity clusters (see Fig. 3.3), including magnitude distributions, horizontal and depth uncertainties, and depth distributions for both long-term 2008-2024 catalog and iMUSH (XD) 2014-2016 catalog.

In the NNW group we observe a total of 59 predominantly NNW-SSE striking, steeply dipping strike-slip FPS along the NNW-SSE trending lineation of seismicity hypocenters (Fig. 3.6). The distributions of primary fault-plane strikes, and corresponding pressure and tensile axes indicate relatively low variability among FPS in the NNW cluster. In addition, the NNW group contains the M_d 4.3 event of February 2011 which is associated with a large non-double couple

component and steeply dipping NW-striking primary plane, described in section 2.4. The SSE group contains comparatively fewer well-resolved solutions—only 18 in total, which lie along a similar NW-SE striking trend observed in epicenters of both the long-term and XD catalogs. In addition, fault-plane solutions in the SSE group are more heterogeneous than in the NNW group. Here, we observe only a few FPS which are consistent with strike-slip motion along the southern Saint Helens Seismic Zone, as observed in the NNW group. The remaining solutions generally consist of mechanisms consistent with dip-slip thrust-faulting motion along either NW-SE or SW-NE oriented fault planes.

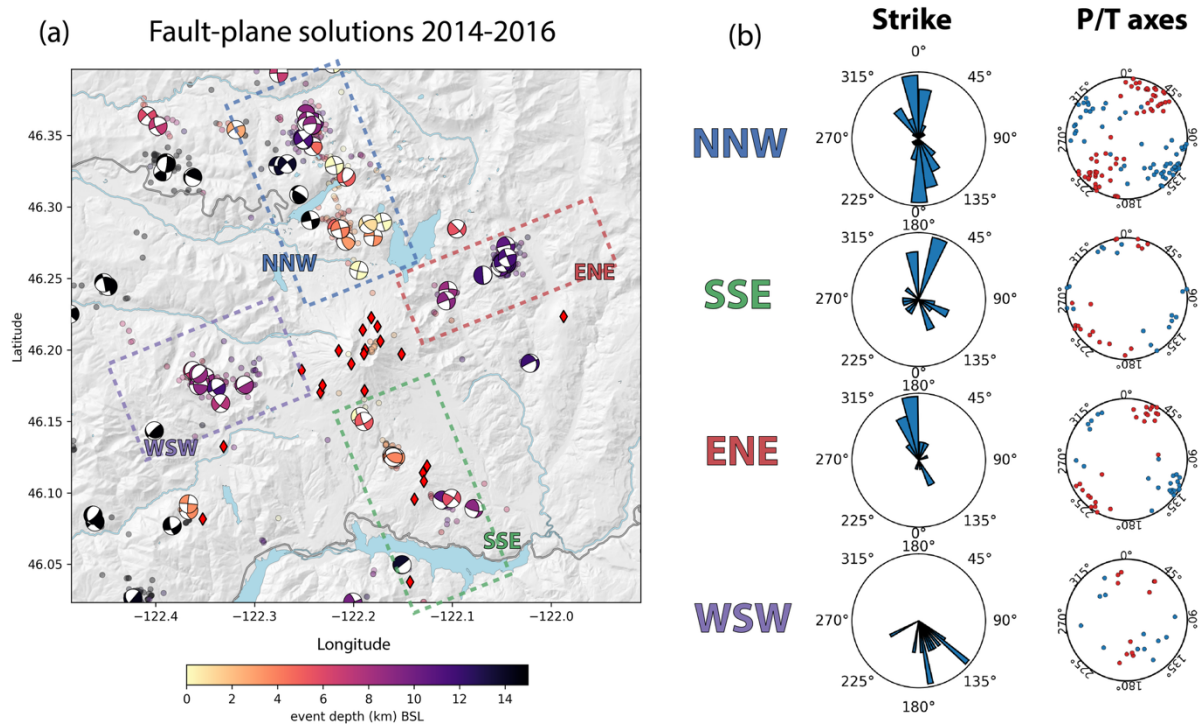


Figure 6: Map (a) of fault-plane solutions using the XD and permanent networks from 2014-2016. Transparent colored dots are locations of events in the XD catalog that do not have a fault-plane solution. Strike and P/T axes (b) of fault-plane solutions in each of the four seismicity clusters.

The WSW group contains 12 well-resolved fault-plane solutions—the fewest of all four seismicity groups (Fig. 3.6). Fault-plane solutions in the WSW group fall into two categories: (1) FPS consistent with normal dip-slip along steep, ENE-WSW oriented primary planes that orient directly towards and away the Mount St. Helens edifice, and (2) strike-slip along near-vertical NW-SE or NE-SW oriented faults. The ENE group contains a total of 24 well-constrained fault-plane solutions—most of which are very similar to each other and those observed in the NNW group. These mechanisms are consistent with steeply dipping strike-slip faults along a NNW-SSE or ENE-WSW fault plane. Additionally, the ENE group contains the M_l 3.9 event of January 2018, which reflects dip-slip normal faulting along a NE-SW oriented fault plane—a mechanism which is not like any other fault-plane solution in the ENE group from the 2014-2016 period.

3.4) Discussion

Our enhanced earthquake catalogs show that most non-volcanic seismicity nearest to the Mount St. Helens volcano, organizes into four distinct groups that are located near its NNW, SSE, WSW, and ENE flanks (Fig. 3.2, Fig. 3.3). Seismic activity within all four non-volcanic seismicity groups, aside from the SSE group, is persistent throughout most of entire study period. In addition, each of the four groups are separated from volcano-tectonic earthquakes by approximately equal distances of roughly 5-7 km. Here, we examine whether the broad-scale spatial and temporal observations we list here are the result of interaction between the underlying tectonic structure and forces from the volcanic system and regional stress field.

3.4.1) Do the earthquake groups interact?

Seismicity within each of the four non-volcanic groups are spatially separated from both each other and volcano-tectonic events. At no point in time do earthquakes from any of the four non-volcanic seismicity groups appear to migrate from group to group, nor interact with the

Mount St. Helens volcanic plumbing system. In addition, the event rates of each of the four non-volcanic seismicity groups and those of volcano-tectonic earthquakes are uncorrelated. We likewise do not observe focal mechanism patterns near the volcanic complex that would indicate magmatic inflation or deflation, such as ring-like dip-slip faulting surrounding the edifice. Thus, the spatial separation of non-volcanic and volcanic seismicity, lack of temporal correlation of event rates, and lack of ring-like faulting around the volcanic complex indicate that direct interaction between volcanic and adjacent tectonic structures during the 2008 to 2024 study period is unlikely or not observable through seismicity alone.

3.4.2) Fault architecture near Mount St. Helens: possible orthogonal fracture zones?

The orthogonal configuration of the four non-volcanic seismicity groups around the Mount St. Helens volcano (Fig. 3.2) is striking. The locations of the four non-volcanic seismicity groups may be explained by near-orthogonal fracture zones that intersect beneath the modern-day Mount St. Helens edifice. If true, the apex of orthogonal fracture zones could form a high-permeability pathway, which may assist the upward migration of magmatic fluids towards the shallow depths, contributing to the formation of the upper-crustal Mount St. Helens volcanic complex. Alternatively, it is possible that the four non-volcanic seismicity groups simply occur along faults that are not related to the Mount St. Helens magmatic system and are coincidental. Here, we examine our catalog of focal mechanisms and compare them to mapped geologic features to test the possibility that the modern-day Mount St. Helens volcanic complex is underlain by intersecting fracture zones.

Double-couple fault-plane solutions yield two orthogonal nodal planes which can equally explain the observed polarity patterns. Due to this inherent non-uniqueness, prior knowledge of the geologic setting is necessary to constrain the most plausible nodal plane. The hypocenter

distributions and NNW-SSE to NW-SE oriented nodal planes of focal mechanisms within the NNW and SSE non-volcanic groups are consistent with the geometry of the Saint Helens Seismic Zone revealed by magnetotelluric (Hill et al., 2010; Bedrosian et al., 2018) and velocity imaging studies (Kiser et al., 2018; Ulberg et al., 2020). Moreover, the NNW-SSE strike-slip mechanisms in both sides of the volcano closely resemble those of Weaver et al. (1987), which were used to originally define the SHZ. Thus, the NNW and SSE seismicity groups can be explained by slip along the SHZ, and that the SHZ is continuous between the north of, beneath, and south of the Mount St. Helens volcanic complex.

The WSW and ENE seismicity groups both contain a significant number of fault-plane solutions with candidate nodal planes that reflect a WSW-ENE fracture zone that point towards the Mount St. Helens volcano. The WSW group contains focal mechanisms that suggest a combination of slip along NW-SE or NE-SW oriented strike-slip faults and steeply dipping normal faults along WSW-ENE oriented plane. Here, normal faulting is suggestive of localized crustal extension along an axis that is subparallel to the SHZ. This is unexpected considering its orientation to the plate convergence vector (Fig. 3.1), which we assume is the maximum horizontal compressive stress. In the WSW group, both the epicentral trend of earthquake locations and the NE-SW trend of nodal planes are approximately subparallel to the NE-SW trend of Quaternary vents between the northern and southwestern flanks of the Mount St. Helens volcano (Fig. 3.2, Fig. 3.6).

The focal mechanisms in the ENE group between 2014 and 2016 reflect strike-slip along either NNW-SSE or WSW-ENE oriented nodal planes. These mechanisms closely resemble those found in the NNW group and have little variation. Without measurable surface deformation or direct observation of the fault, it is difficult to determine whether slip occurs along either the

NNW-SSE plane or the WSW-ENE plane. If the NNW-SSE plane is favorable, it suggests that the ENE group could reflect an offset splay fault of the SHZ. On the other hand, the WSW-ENE candidate nodal plane is subparallel to a set of NE-SW air photo lineaments, or pyrite bearing dikes found in the Spirit Lake Batholith, east of the SHZ (Fig. 3.1). In addition, the NE-SW oriented normal faulting event of January 2018 (Fig. 3.3) provides some further evidence a roughly NE-SW oriented fracture zone that points towards the Mount St. Helens volcano, may exist in the ENE group.

Our analysis of hypocenters and focal mechanisms within both the WSW and ENE non-volcanic groups, and agreement with trends of known geologic structures, provide some evidence for fracture zones that are sub-orthogonal to the Saint Helens Seismic zone. Given the lack of rate correlation and spatial separation of the four non-volcanic seismicity groups with volcano-tectonic seismicity, it is unclear if these two fractures zones form a continuous structure that intersect the SHZ beneath the Mount St. Helens volcano. If we suppose that the NE-SW fracture zone and SHZ do intersect beneath volcanic complex, we speculate that this could form a high-permeability pathway that encourages the upward migration of melt and volatiles from midcrustal depths. We further speculate that the separation of volcanic and nonvolcanic seismicity could be attributed to a lateral brittle-ductile transition surrounding the magmatic reservoir between approximate 5-15 km depth BSL. In this framework, VT seismicity in the Mount St. Helens volcano is different from non-volcanic seismicity in that it is driven primarily by volatile flux (e.g. Moran, 1994, Hirao et al., 2025), rather than tectonic stresses.

3.4.3) Fault heterogeneity and fluid circulation along the SHZ

Non-volcanic seismicity within the northern section of the SHZ (i.e. the NNW group) is in strong contrast to seismicity in the southern section (i.e. the SSE group). Compared with the

southern section, seismicity in the northern section is more frequent, can reach higher magnitudes (M 5.5 of 1981; Weaver and Smith, 1984; Weaver et al., 1987), is distributed more continuously (Fig. 3.2, Fig. 3.3), and features less variation of fault-plane solutions (Fig. 3.6). By contrast, seismicity in the SSE group is less frequent, more disjointed, and has more heterogeneity among fault-plane solutions. More frequent slip cycles along the NNW group could lead to the formation of a more matured, homogeneous, and continuous faulting surface (e.g. Guo et al., 2023). Moreover, Quaternary volcanic vents could interfere with a throughgoing fault zone along the SSE group (Fig. 3.2, Fig. 3.6). Here, volcanic intrusions of different elastic properties could introduce local material heterogeneities and influence the local stress state, leading to variances in faulting style (Fig. 3.6). Despite being subject to a similar regional background stress state, it is still unclear why the SSE group has such lower seismicity rates.

Several observations indicate that the SHZ hosts hydrothermal fluid circulation. Magnetotelluric and seismic imaging at the regional scale (Hill et al., 2010; Bedrosian et al., 2018), in addition to the geothermal energy surveys at local field scales indicate that hydrothermal fluid circulation is present along the SHZ near both the NNW and SSE nonvolcanic groups (Norman et al., 2015; Steely et al., 2021). In addition, an elevated geothermal fluid gradient of 50 C/km is observed along the northern SHZ at a temperature gradient well along the Green River (Fig. 3.1). We speculate that higher levels of fluid circulation along a more focused pathway could support higher seismicity rates along the NNW group, compared with the SSE. It is unclear if fluids in the northern and southern SHZ are directly connected to the magmatic reservoir beneath the Mount St. Helens volcano, but it is plausible that they originate from a common source at midcrustal depths.

3.4.4) Revisiting the Weaver et al., (1987) hypothesis: crustal extension beneath Mount St. Helens

Weaver et al. (1987) suggest that the slight offset of strike along the SHZ lends itself to crustal extension and the formation of a pull-apart basin-like feature beneath the Mount St. Helens volcano. In turn, crustal extension is thought to aid in the transfer of magma and volatiles towards shallower depths through opening of the brittle crust and decompression melting. Pull-apart basins and crustal extension, which are frequently observed along transform faults, are also associated with en-echelon normal faulting oblique to the transform fault. While we do not find any evidence for normal faulting within the Mount St. Helens volcanic complex, we do find evidence for NE-SW oriented normal faulting in both the WSW and ENE non-volcanic groups (Fig. 3.6). However, we do not know of any geomorphic or sedimentary expressions for crustal extension, including sag ponds or shingling of basin deposits (e.g. Wakabayashi et al., 2004). Given that the SHZ is relatively young and its cumulative displacement, compared to major transform faults (e.g. San Andreas Fault) is low, these records may not have been well defined to begin with or could have been overprinted by recent volcanism. Through our analysis of high-precision earthquake locations, focal mechanisms, and geologic features, we suggest that crustal extension proposed by Weaver et al. (1987) is plausible but not required to explain the position of the Mount St. Helens volcano.

3.5) Conclusion

The modern-day Mount St. Helens volcanic plumbing system is underlain by the Saint Helens Seismic zone, a right-lateral strike-slip fault zone channels melt and magmatic volatiles between the Siletzia and Spirit Lake Batholith terranes. To better characterize the interaction between active volcanic plumbing systems and adjacent tectonic structures, we create two high-

resolution enhanced catalogs of non-volcanic seismicity. Here, we construct a long-term catalog spanning between 2008 and 2024 and a short-term catalog that leverages the high-density configuration of the iMUSH network, spanning between 2014-2016. We then make further use of the iMUSH catalog to solve for over 180 fault-plane solutions to resolve fault architecture near Mount St. Helens. Our enhanced catalogs show that nonvolcanic seismicity clusters into four distinct groups that lie within 5 to 20 km of the edifice. These groups appear to be temporally uncorrelated with volcano-tectonic activity—we do not find evidence for migration or correlated event rates between groups. Detailed analysis of hypocenters, fault-plane solutions, and mapped geologic features indicate a possible sub-orthogonal fracture zone that intersects at the Mount St. Helens magmatic plumbing system. This feature could assist in the upward migration of magmatic volatiles and melt which form the modern-day Mount St. Helens volcanic plumbing complex.

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3.7) Bridge

In this chapter, we apply a similar procedure as in Chapter 2 to analyze seismic activity at the regional scale along faulting structures that surround the Mount St. Helens volcano in southern Washington, USA. In addition, we augment the template matching and location procedures with machine-learning phase detection along over 50 temporary seismometers included in the iMUSH dense array experiment. In addition, we analyze newly resolved fault-plane solutions to characterize the orientation of faulting surrounding the volcano at an unprecedented resolution. Here, we find evidence for intersecting fault zones directly beneath the Mount St. Helens volcano.

The computational resources required to conduct the numerical analysis for Chapters 2 and 3 are vast. The cross-correlation for both chapters took over 60 years of CPU time, distributed over thousands of cores hosted by the Talapas high performance computing system. A single iteration of the relative locations, with bootstrap uncertainties, required 2 terabytes of memory to complete within 24 hours of human time.

In Chapter 4, we expand the study region to the sub-continental scale to cover the entire Pacific Northwest, USA. There, we construct an automated detection and location workflow to catalog over 154,000 earthquakes across the Pacific Northwest between 2002 and 2024.

CHAPTER IV

DEVELOPMENT OF AN ENHANCED SEISMICITY CATALOG FOR THE PACIFIC NORTHWEST, USA: THE CRESCENT SEISMICITY CATALOG VERSION 0

4.1) Introduction

The Pacific Northwest (PNW) of the United States (incl. Washington, Oregon, and Northern California) overlies the Cascadia Subduction Zone a roughly 1000 km-long megathrust fault that extends between offshore of Vancouver Island, Canada and the Mendocino Triple Junction in Northern California. The Cascadia megathrust is exposed at the surface from 50 to 130 km offshore west of the North American continental coastline, where the Juan de Fuca and Blanco plates converge with the North American plate at a rate of 3-4 cm/year (e.g. Nishimura et al., 1984). The Cascadia Subduction Zone is host to a wide range of seismic phenomena of large and small magnitudes (M)—many of which remain poorly understood. Such seismicity of interest includes the catastrophic and tsunamigenic $M9+$ earthquakes along the megathrust interface (e.g., Heaton and Kanamori, 1984; Atwater 1987; Satake et al., 1996), deep $M6+$ earthquakes within the subducting slab beneath the Seattle-Tacoma region in Washington (e.g. Ichinose et al., 2004), $M7+$ earthquakes along upper crustal faults (e.g. Brocher et al., 2017; Wirth et al., 2025),

widespread small-magnitude seismicity of both tectonic and volcanic origins, and nonvolcanic tremor (Bostock et al. 2019).

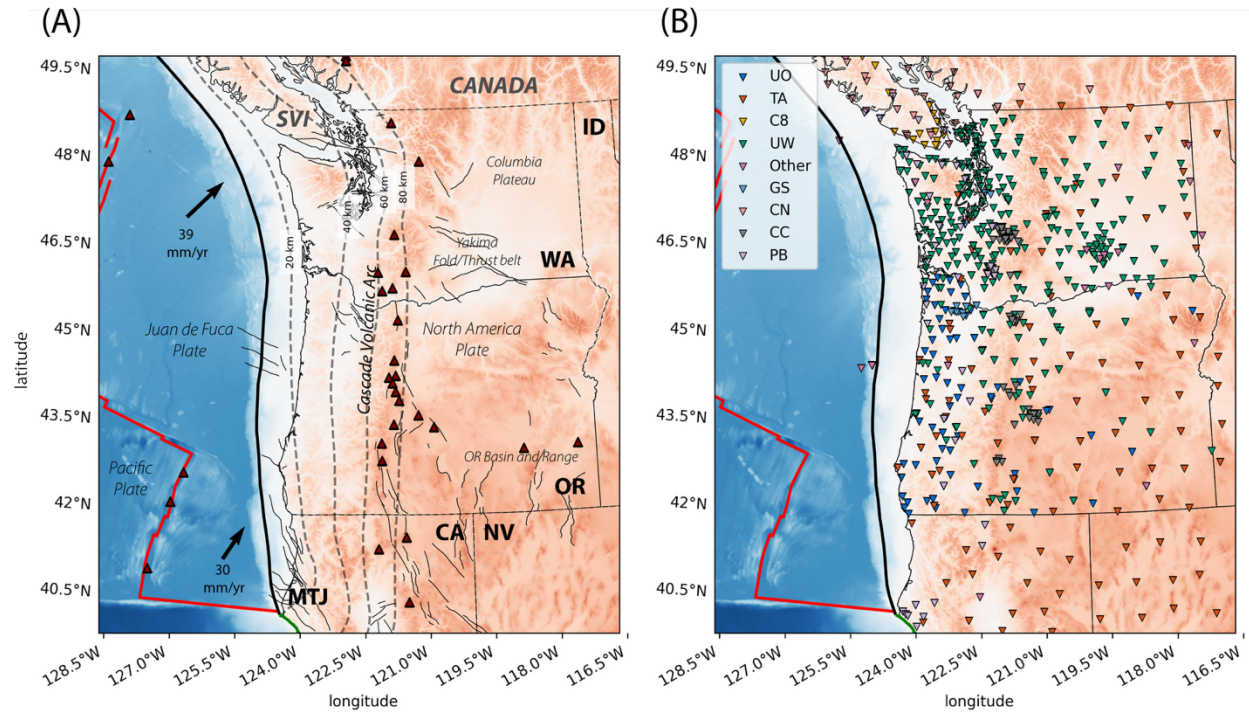


Figure 4.1: (A) Study region with plate boundaries, Quaternary volcanoes from the Smithsonian Volcanoes of the World database (Global Volcanism Program, 2025), plate convergence rate of Juan de Fuca plate with respect to stable North America from the NUVEL-1 model (Demets et al., 1990), slab contour depths from Demets et al., (2012), and crustal fault traces from the CRESCENT Community Fault Model viewer (Melgar et al., 2025). (B) Seismic recording stations used in this study. See section 4.2.2 for details. Topographic and bathymetric data are sourced from the SRTM+ 15 arcsecond digital elevation and bathymetric combined model. Mapping and projections are handled with Cartopy (Elson et al., 2022). Abbreviations: SVI = Southern Vancouver Island, MTJ = Mendocino Triple Junction, WA = Washington, OR = Oregon, CA = California, NV = Nevada, ID = Idaho.

The complexity of Cascadia’s seismic behavior raises several outstanding questions that remain central to understanding earthquake hazards in the Pacific Northwest. Despite clear geological evidence for repeated M9 megathrust earthquakes, the plate interface is anomalously quiet in the instrumental era, and the causes of this long-standing quiescence remain unknown (Trehu et al., 2015; Bostock et al., 2019). Significant along-strike heterogeneity in coupling,

tremor density, fluid distribution, mantle buoyancy, slab geometry, and other crustal properties is well documented, yet the mechanisms linking these variations to each other, to rupture segmentation, and the likelihood of full- versus partial-margin events are unresolved (Zandt and Humphreys, 2008; Bodmer et al., 2018, 2020; Delph et al. 2018, 2021). The physical processes governing episodic tremor and slip, its role in the megathrust earthquake cycle, and whether seismogenic slip penetrates into the ETS region remain first-order open questions with direct implications for hazard (Rogers and Dragert, 2003; Melgar et al., 2022). Uncertainties also remain in the distribution and activity of crustal faults throughout the Cascadia forearc: many Quaternary active faults are mapped, yet seismicity on them is sparse, paleoseismic coverage is uneven, and some geodetically inferred block boundaries do not coincide with known faults. Offshore paleoseismic data are largely absent, and several crustal structures remain poorly characterized, representing critical empirical gaps (Clarke and Carver, 1992; Oppenheimer et al., 1993; Frankel et al., 2015). Additional uncertainties include the origins of deep intraslab M6+ earthquakes beneath Puget Sound (Kao et al., 2008; Gombert and Bodin, 2021), the temporal evolution of strain accumulation between great earthquakes (Michel-Wolfe et al., 2022), the structure and strain partitioning on faults surrounding the Mendocino Triple Junction (Furlong et al. 2024, Shelly et al. 2025), and the poorly constrained interplay between megathrust, crustal, and intraslab seismicity. Together, these gaps underscore the need for more complete and higher-resolution seismic observations to illuminate how and where the Cascadia plate boundary accommodates strain and to refine seismic hazard assessments for the region.

Seismicity catalogs provide critical observations for the study of active tectonics and hazard assessment. Seismic activity in the Pacific Northwest is continuously monitored, recorded, expertly curated, and distributed via the ANSS Comprehensive Earthquake Catalog by

the well-established Pacific Northwest Seismograph Network (PNSN; University of Washington, 1963). The PNSN conducts reliable and operationally proven real-time seismicity monitoring with the Earthworm software distribution, which applies an amplitude-based (i.e. short-term average, long-term average ratio; STA/LTA) detection mechanism to identify phase arrivals from continuous data and trigger detections. Skilled PNSN analysts then manually reviews the associated phase arrivals and locates the event using a linearized form of the earthquake location solution via the Hypoinverse solver (Klein, 2002) and a set of 1-dimensional P-wave velocity models selected according to the region of interest (Hartog et al., 2020; Hellweg et al., 2020). Key strengths of the PNSN-curated catalog include rapid continuous updates with new events, high quality of detections reinforced by manual review, and a set of easily accessible origin products (including locations, phase arrivals, polarities, magnitudes, etc.). Because continuous seismic data is noisy, amplitude-based detection routines necessarily use higher detection thresholds to mitigate false detections, which reduces the likelihood of detecting small magnitude earthquakes. Moreover, the quality of event locations and uncertainties can be constrained by the practical need to use by the 1-dimensional velocity models, which rarely capture lateral heterogeneity and topography. In addition, the earthquake location procedure is inherently nonlinear (e.g. Lomax et al., 2002) and can have multimodal solution spaces. Because the PNSN mission rightly prioritizes rapid response and robust real-time operations, there are inevitable tradeoffs in event quantity and location precision relative to what may be possible in non-operational, offline analyses. Thus, while the catalog is excellent for operational monitoring, there may be opportunities to further refine small-event detection and event locations in complementary, offline scientific catalogs.

Recent years have seen substantial progress in using machine-learning methods to build enhanced catalogs of microseismicity (Zhu et al., 2022; Tan et al., 2021). These approaches typically begin with the detection of seismic body-wave arrivals using deep-learning phase pickers (e.g., Zhu et al., 2019; Ni et al., 2025). After picking, the resulting P- and S-wave arrival times from stations across a region are associated—that is, grouped into sets that correspond to the same earthquake source. A variety of techniques exist for both picking and association, including gridsearch-based methods (e.g. REAL of Zhang et al., 2019; PyOcto of Münchmeyer et al., 2024) and machine-learning-based methods (e.g., PhaseLink of Ross et al., 2019; GaMMA of Zhu et al., 2022; GENIE of McBrearty and Beroza, 2023). Once phases are successfully associated, earthquakes can be located using travel-time information, and additional source characteristics such as magnitude can be estimated. This workflow can be used to identify conventional small-magnitude onshore seismicity, but—when carefully tuned—can also reveal more atypical signals (e.g., low-frequency or hybrid events), volcanic seismicity (e.g. Wilding et al., 2023) offshore seismicity that is otherwise difficult to detect (Bornstein et al., 2024; Journeau et al., 202X).

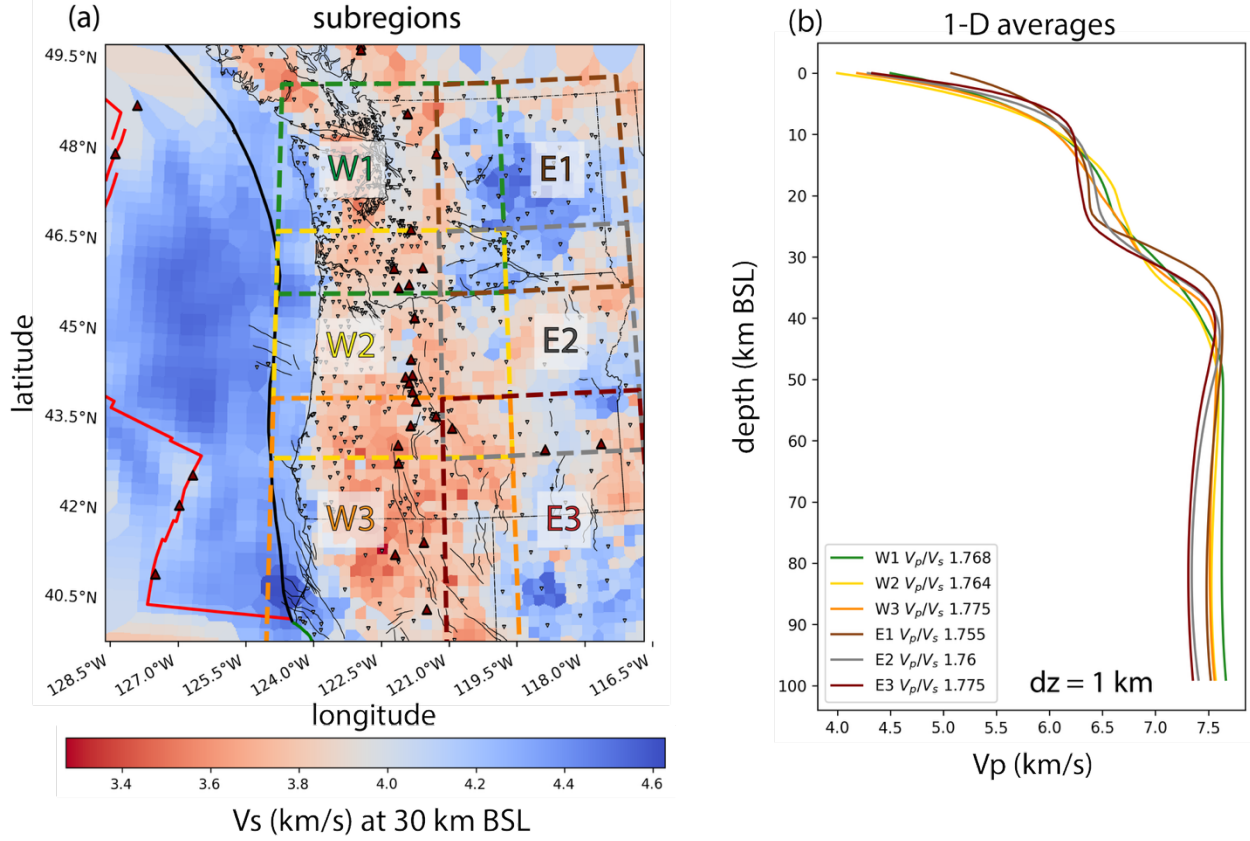


Figure 4.2: Experimental setup. W1, W2, W3, E1, E2, and E3 are the six subregions to which we apply the association and location steps towards building the CRESCENT Enhanced catalog. Additional details shown in Table 1 and Section 4.2.3. (A): Subregion boundaries, shear-wave velocity (V_s) at 30 km BSL from the CRESCENT Community Velocity Model (He and Delph, 2025), locations of volcanoes and seismic stations. The velocity model is interpolated to a grid spacing of 1 km in the x, y, and z directions for all six subregions. The V_s model is converted to V_p before resolving travel-time grids and locations. (B): 1-D averaged velocity models for each region for use with the Pyocto associator. We use a spacing of 1 km to avoid sharp interfaces.

4.2) Methods

Our goal is to compile an enhanced catalog of seismicity in the Cascadia subduction complex (Fig. 4.1) between the years 2002 to 2024 by applying an automated detection and location workflow that is uniform across the entire study region. Our workflow presented here is similar to the detection and location workflow presented in the Github notebooks of Bachelot et al. (2023). Here, we process phase arrival picks generated by a specially-trained model of the EQTransformer (Mousavi et al., 2020; Ni et al., 2022), using the PyOcto associator algorithm

(Münchmeyer et al. 2024) , and the NonLinLoc location solver (Lomax, 2002; Lomax et al., 2014) to search for optimal location parameters in 3-dimensional travel time grids derived from the 3-D CRESCENT Community Velocity Model (He et al., 2026).

Grid	Origin (lat, lon)	SW corner (lat, lon)	NE corner (lat, lon)	Dimension: x/y/z (km)	Vp/Vs	N Stations	File size per grid (32-bit float)
W1	47.50, -122.25	45.75, -125.00	49.25, -119.50	414/390/100	1.768	832	64.6 MB
W2	44.90, -122.25	43.00, -125.00	46.80, -119.50	434/422/100	1.764	487	73.3 MB
W3	41.00, -122.25	38.00, -125.00	44.00, -119.50	464/666/100	1.775	929	123.1 MB
E1	47.50, -118.79	45.75, -121.08	49.25, -116.50	346/390/100	1.755	173	54.0 MB
E2	44.90, -118.79	43.00, -121.08	46.80, -116.50	362/422/100	1.760	128	61.1 MB
E3	41.00, -118.79	38.00, -121.08	44.00, -116.50	386/666/100	1.775	318	102.9 MB

Table 4.1: Geographic and computational details of the six subregion grids. See Figure 4.2 for map and 1-dimensional averages.

4.2.1) Phase detection

Seismicity in the Pacific Northwest is highly varied in nature. Wide variations of source environments, mechanisms, ray paths, recording environments, and sensor types may all influence the observed waveform character. To cover over 20 years of study, an ideal phase picker should be fast and flexible, while retaining timing and classification accuracy. Recent adaptations of artificial intelligence-based image processing algorithms, trained on a vast accumulation of analyst-labelled seismic data, excel at identifying previously undetected earthquake signals of low signal-to-noise from large volumes of continuous waveform data (e.g. Ronneberger et al., 2015; Zhu et al., 2019). We utilize the phase arrival picks of Ni et al. (2023), who apply a regionally trained variant of the Earthquake Transformer (EQTransformer) to identify and classify P- and S-wave phase arrivals from stations in the Pacific Northwest (referred to as PNW-EQTransformer). Ni et al. (2023) enhance the original EQTransformer model for the PNW region by augmenting the training dataset with analyst-picked arrivals from

roughly 47,000 events recorded by the PNSN. They then apply the PNW-EQTransformer model to continuous waveform data recorded on 2,177 broadband seismic stations from 44 networks distributed across the Pacific Northwest (Fig. 4.1B), resulting in a database containing 188 million P- and 88 million S-wave picks between January 1, 2002 and December 31, 2023, where each phase pick is associated with its respective station coordinates, arrival timing, P- or S-wave phase arrival classification, and classification probability between 0 and 1.

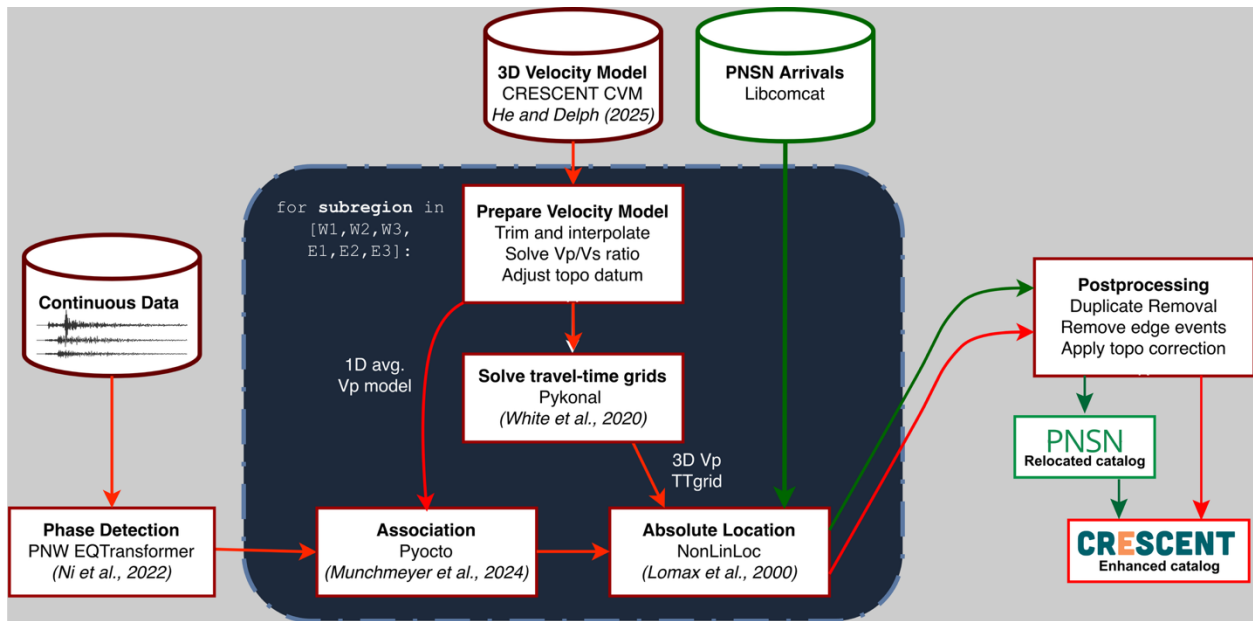


Figure 4.3: Schematic of workflow to develop the CRESCENT Enhanced catalog. Subregions, 1d averaged VP velocity model, and 3D Vp model of He et al. (202X), are shown in Figure 4.2. Modified from the workflow presented by Bachelot et al. (2023). The steps that involve preparation of the velocity model, solution of travel-time grids, association, and absolute location are performed independently for each of the six regions. The outputs are then merged in Postprocessing. The resulting PNSN and CRESCENT enhanced catalogs are shown in Figure 4.4. Waveform is from the July 12, 2019 M 4.4 earthquake in Monroe, WA. See Section 4.2 for details regarding individual steps. Figure created with draw.io (2025).

4.2.2) 3D Velocity model

The Pacific Northwest is vast and geologically complex. Unresolved variations in velocity structure and topography will introduce epistemic errors in final location solutions. To minimize such errors, we apply the recently developed CRESCENT Community Velocity Model

(CVM)—a 3-dimensional shear-wave velocity model that covers southern Vancouver Island, Washington, Oregon, and Northern to Central California (He et al., 2026) roughly 100 km along depth. Due to the size of the study region, we divide it into six subregions—three along West and East (W1-3, E1-3; Table 4.1), which overlap their adjacent grid by between 110-130 km. The positions and sizes of each grid were defined in consideration of relative station density and memory constraints required to solve for location solutions. In addition, the amount of overlap was chosen to reduce edge-effects that are often a consequence of azimuthal gap increases due to stations being excluded from the respective model space. Ideally, each event should have at least 55 km of space between itself and the edge of its respective grid. We perform the association and location steps for each of the six grids separately and remove duplicate events from the catalog in an additional post-processing step.

To prevent short-lived spikes in detection rates, we only include permanent stations that continuously record for a minimum of 365 days, not considering unplanned service outages. In addition, we omit picks recorded on temporary networks since their short-lived, experiment-specific configurations and higher noise levels make them sub-optimal for continuous monitoring. We make an exception to include the USArray Transportable Array (network code TA), which includes 165 stations in the study area--most of which operated continuously between 2006 to 2009 (IRIS Transportable Array, 2003). With the selection criteria in place, we use arrival data from a total of 939 stations from 14 permanent networks including the Transportable Array (Fig 4.1B).

4.2.3) Six geographic subregions

For each of the six subregions, we trim and interpolate the velocity model using locally assigned coordinate system with respect to the origin at its center latitude and longitude (Table

4.1; Figure 4.2). We define interpolant coordinates using an azimuthal equidistant projection with respect to the origin with 1 km spacing in x, y, and z directions. To represent onshore topography and offshore bathymetry, we interpolate the velocity models up to a depth of 100 km below the earth's surface defined by the SRTM15+ model, which contains topographic and bathymetric elevation values at a 15 arcsecond resolution (Tozer et al., 2019). This step is intended to account for source-station delays that result from topography, where stations be positioned at varying elevations. In model space, a depth of 0 km is the surface of the Earth regardless of its elevation. This approach is advantageous since it constrains the model space to the solid earth and eliminates the need to use placeholder velocity values (typically the V_p in air, ~ 0.343 km/s) in the space above the earth's surface. Topography is restored to each of the location solutions in a later step, where we interpolate the digital elevation model for the elevation at an event's epicenter, which is added to the event solution depth in model spaces—thus restoring the sea level datum.

For event association and location, we convert each of the six subregional V_s models to V_p by estimating subregion-specific V_p/V_s ratios using the definition of Wadati (1933) (Equation 4.1; Fig. 4.2b; Table 14.).

$$(t_s - t_p) = (V_p/V_s - 1) \times (t_p)$$

(Equation 4.1)

Here, we include the observed S- and P-wave travel time for analyst-confirmed events in the PNSN catalog. Since seismicity is unevenly distributed and that V_p/V_s ratios are specific to unique subsurface conditions, we apply a stratified sampling approach by binning each subregion into $10 \times 10 \times 2$ km³ volumes along the longitude, latitude, and depth directions. From each bin, we select at most 10 events whose P- and S-wave travel times contribute towards solving the

subregion-representative V_p/V_s ratio. We then estimate the regional V_p model by multiplying the V_s model by the resulting V_p/V_s ratio, which all range from 1.755 to 1.775. The resulting V_p models are then used to solve for theoretical P-wave travel time grids for each station in each subregion using the Fast-Marching Method implemented in Pykonal (White et al., 2020).

4.2.4) Phase association and location

For detections to be located, their phase picks must be grouped and associated to an event. The Pyocto associator (Munchmeyer, 2024) is designed for high-throughput applications with an ML-based phase detector, whose challenges may include unknown true event counts, large quantities of phase picks, false phase detections, and uneven and changing station geometry. Pyocto uses an efficient octotree nested grid-search algorithm to associate phase picks to a coarse grid of theoretical sources, based on their timing and station coordinates. Theoretical travel times are computed using an Eikonal solver informed by a 1-dimensional velocity model. For each of the six subregions, we solve for region-specific 1-dimensional velocity model by computing the along-depth average of its respective 3D V_p model in 1 km increments. We then apply the Pyocto associator to phase picks from all stations within each subgrid and save events with a minimum of 3 P- and 3 S-wave picks recorded by at least 6 stations. We then locate the resulting associated events with the non-linear probabilistic framework presented in NonLinLoc (Lomax et al., 2000; Lomax et al., 2014). We apply the OCT-TREE nested gridsearch of the 3-dimensional travel time grids with Equal-Differential Time (EDT) weighting to solve for optimal location parameters and 3-dimensional uncertainty ellipses of all events within their respective subregion grids.

4.2.5) Postprocessing

We detect and locate a total of over 1.7 million events—many of which are likely of poor quality and may include non-earthquake sources, duplicate detections, and events outside of the study region. We apply a series of postprocessing steps to remove events of poor quality. For each of the six subregions, we remove events that are within 10 km of its respective subregion boundary and 0.2 km of the top of the model space (i.e. the Earth’s surface). To remove duplicate events in overlap zones, we first identify events located within the overlapping regions between adjacent subregions. We apply a Ball-Tree search (Omohundro, 1989) implemented in the Scikit-learn package (Pedregosa et al., 2011) to efficiently retrieve, for each event, a list of neighboring events within a 30 km radius. We then iterate through all neighboring pairs and classify as potential duplicates those that occurred within 2 seconds of each other and originated from different subregional grids. Within each cluster, we identify duplicates as pairs of events that share one or more common phase arrivals and retain the event with the smallest azimuthal gap.

4.2.6) Benchmarking

To evaluate the performance of our enhanced detection and location workflow, we generate for comparison a separate catalog consisting of a total of 72,005 events recorded by the PNSN between 2002 and 2024. We obtain phase arrivals for these events with the USGS libcomcat package (Hearne and Schovanec, 2020) and locate them by applying the same location procedure as described. For the sake of comparison, we present events of an acceptable quality from both the Enhanced and PNSN relocated catalogs, which have an uncertainty (1 standard deviation) of less than 5 km in the X, Y, and Z dimensions and an azimuthal gap of less than 250 degrees. Since the PNSN authoritative region does not include Northern California and Northern Nevada, we remove events in the Enhanced catalog south of the 42nd parallel from discussion in this

paper. After applying these criteria to both catalogs, we retain a total of 153,836 events in the Enhanced catalog (Fig. 4.4B)—roughly 3.6 times greater than the 41,806 events which remain in PNSN catalog using the same filtering criteria (Fig. 4.4A)

4.3) Results

We observe increased event counts systematically throughout the entire study region (Fig. 4.4C), most of which is broadly distributed between Southern Vancouver Island, Puget Sound and the Western Cascades in western Washington, and northwest Oregon. In addition, we observe significant increases in seismicity near regions of mapped faults, including the Spencer Canyon Fault and Yakima Thrust belt in central and southern Washington (Fig. 4.4). In both relocated PNSN and Enhanced catalogs, seismicity is uncommon in the Columbia Plateau of Eastern Washington and almost nonexistent within the Basin and Range of Eastern Oregon (Fig. 4.4). We observe a steady increase in the number of recorded events from the beginning of the study period in 2002 until the end in 2024 (Fig. 4.5). An exception to this trend is the spike in event count between 2004 and 2008, which may be explained by high event counts during the Mount St. Helens eruption (Fig. 4.9, Fig. 4.10) and deployment of the USArray (Fig 4.1B, Fig. 4.5).

To explore the contents of the Enhanced catalog, we inspect seismicity related to three environments of interest within the Pacific Northwest including: 1) seismicity near the downgoing Juan de Fuca Plate (Fig. 4.1A), 2) crustal seismicity and earthquake swarms in the overriding North American plate, and. We compare the results of the Enhanced catalog to those of the relocated PNSN catalog for 1) seismicity near the downgoing Juan de Fuca plate (Fig. 4.6, Fig. 4.7), 2) crustal seismicity within the overriding North American plate (Fig. 4.8), and 3) seismicity near Mount St. Helens (Fig. 4.9, Fig. 4.10).

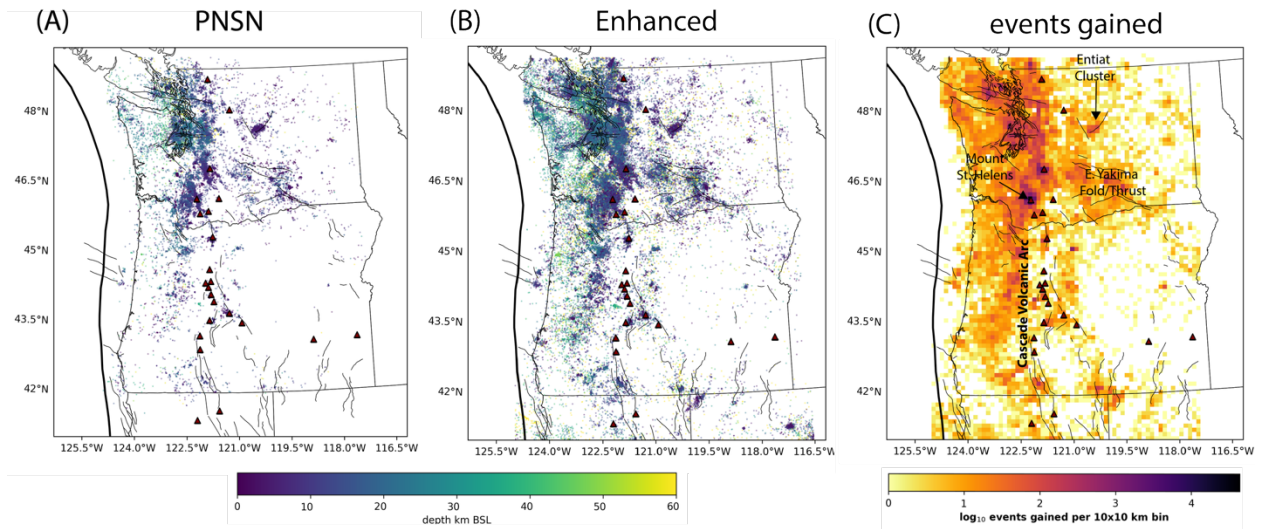


Figure 4.4: Comparison of (A) PNSN and (B) Enhanced earthquake catalogs. Both contain events that have a maximum error of 5 km in x, y, or z and a maximum azimuthal gap of 250 degrees. (C) Number of events in Enhanced catalog minus the number of events in the PNSN catalog per 10 x 10 km bin.

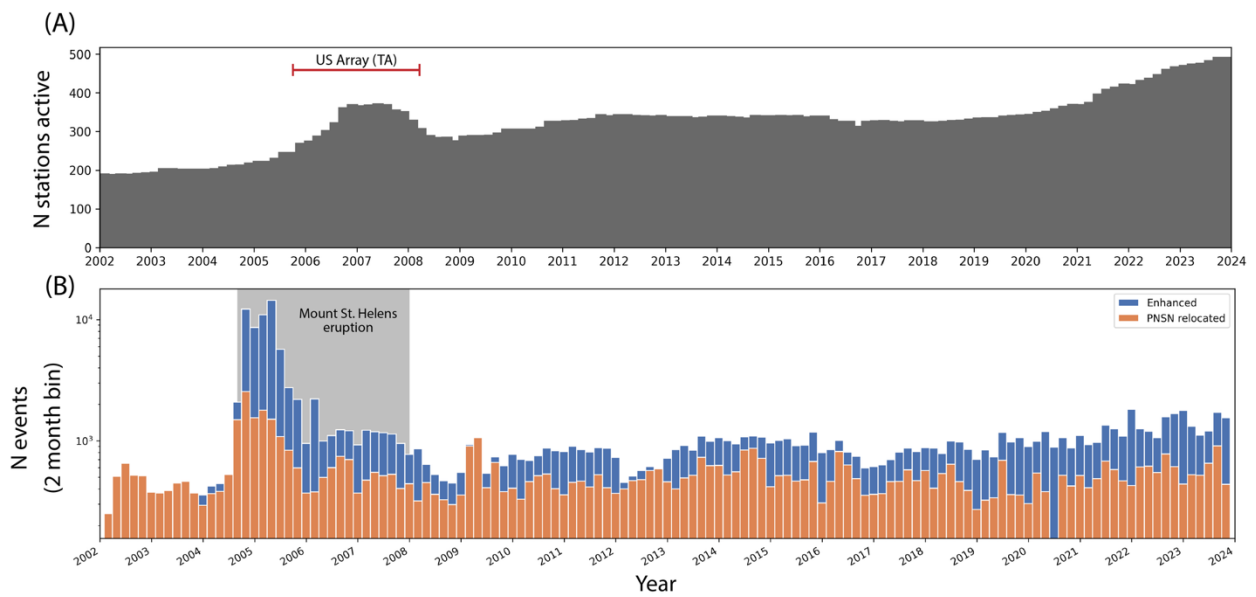


Figure 4.5: Data and event statistics. Number of stations active in 2-month bins in (A), number of events in 2-month bins for both Enhanced and PNSN relocated catalogs in (B).

4.3.1) Seismicity near the Cascadia Megathrust

The Cascadia megathrust is unusual for its low recorded seismicity rates along the plate interface, despite an abundance of evidence that it can host a catastrophic M9+ earthquake (e.g. Heaton and Kanamori, 1984; Atwater 1987; Satake et al., 1996; Small et al., 2025). Numerous lines of evidence show that deformation along the plate interface is accommodated aseismically via episodic slow slip, or through unconventional seismicity types including non-volcanic tremor and low frequency earthquakes. It is unclear if the apparent lack of recorded seismicity near the subducting plate interface reflects the seismogenic nature of the plate interface or is due to a lack of catalog completeness owing to the high detection thresholds of conventional methods. Whether this quiescence reflects true dearth of seismicity or observational limitations, the scarcity of recorded seismicity along the plate interface poses a major challenge for accurately assessing seismic hazard in the Pacific Northwest's major population centers.

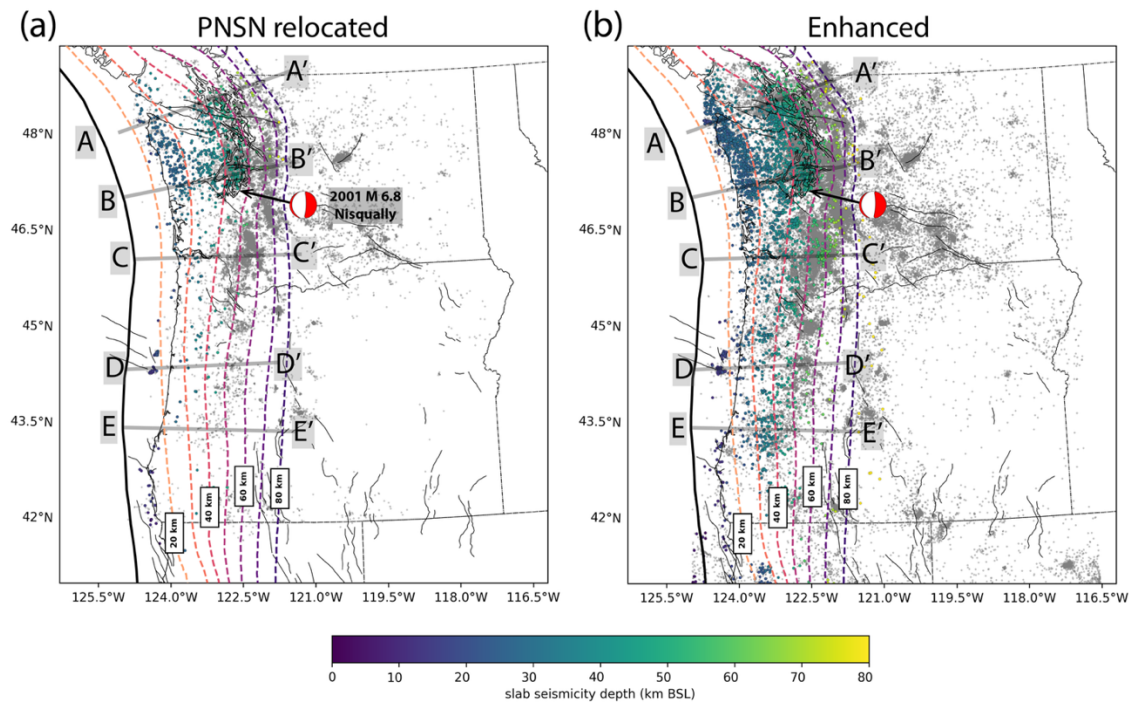


Figure 4.6: Comparison of seismicity near the subducting Juan de Fuca oceanic slab in the (a) PNSN relocated catalog and (b) Enhanced catalog. Seismicity within 6 km of the McCrory et al. (2012) model slab is colored according to the colorbar. Downdip depths of the McCrory et al. (2012) slab shown in colored contours in 10 km intervals between 20 to 80 km. Seismicity away

from the slab shown in grey dots. Red moment tensor is from the Global Centroid Moment Tensor project solution of the February 2001 M 6.8 Nisqually intraplate earthquake (Dziewonski et al., 1981; Ichinose et al., 2004; Ekström et al., 2012). Depth cross sections are shown in Figure 4.7.

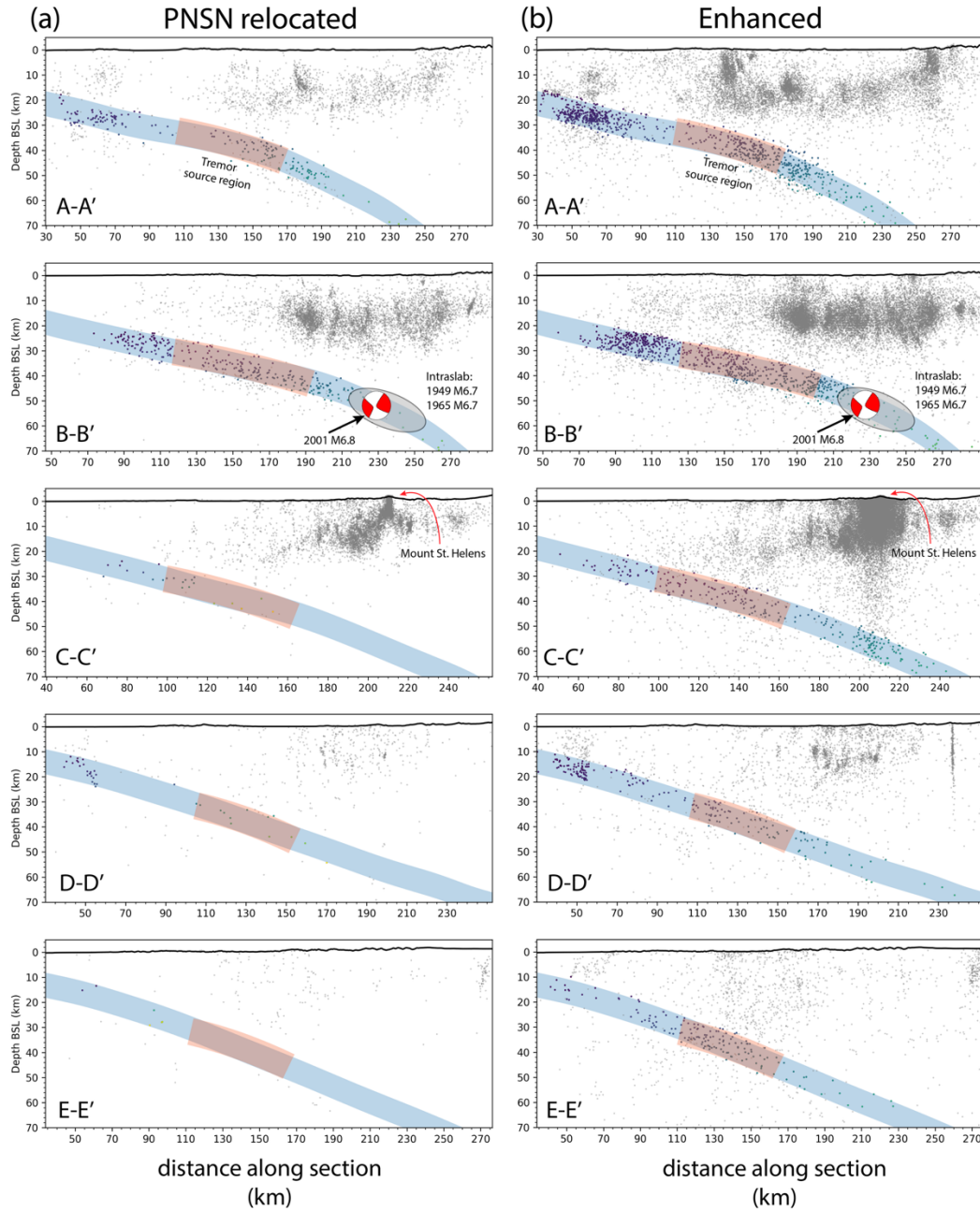


Figure 4.7: Along-depth cross-sections of seismicity near the subducting Juan de Fuca oceanic slab in the (a) PNSN relocated catalog and (b) Enhanced catalog. Map views of cross-section profiles in Figure 4.6. For each profile, we include all seismicity within 50 km of its respective profile. McCrory et al. (2012) slab model is depicted with a blue stripe, assuming a 10 km thickness. Approximate tremor source region between 30-45 km BSL from Bostock (2019) is

depicted along the slab with a light red stripe. Seismicity within 6 km of the model slab is colored according to the colorbar in Figure 4.7. Seismicity away from the slab is shown in grey dots.

To evaluate whether an automated AI-based detection and location workflow can improve detection of seismicity near the downgoing plate, we examine seismicity within 6 km to its nearest point to the McCrory et al. (2012) slab model in both the PNSN relocated and Enhanced catalogs (Fig. 4.6, Fig. 4.7). After removing events with uncertainties greater than 5 km and azimuthal gaps of 250 degrees, we find a total of 7,243 events in the Enhanced catalog within 6 km of the McCrory slab model between depths of roughly 25 to 60 km—a roughly 11-fold increase over 656 events contained in the PNSN relocated catalog. In most cases, newly detected seismicity near the plate interface locates in similar source regions to previously detected events in the PNSN relocated catalog.

Seismic activity near the plate interface is most active beneath Southern Vancouver Island and the Olympic Peninsula and gradually declines southwards of the Puget Sound (Fig. 4.6, Fig. 4.7). Beneath Southern Vancouver Island and the Olympic Peninsula, events near the interface cluster into two along-depth groups between roughly 20 to 30 km and 35 to 60 km and are separated by roughly 20-30 km along dip (Fig. 4.7, A-A' profile). These two groups become less distinct and begin to merge into a single group beneath the Puget sound region, between depths of 20 to 50 km—slightly updip of the source regions of several M 6.7+ intraslab events within the past 80 years—including the M 6.7 earthquakes of 1949 and 1965 and the M 6.8 Nisqually earthquake (Fig. 4.7, B-B'; Ichinose et al., 2004). By comparison, seismicity near the plate interface beneath southern Washington and Oregon is rare and diffuse (Fig. 4.7, C-C', D-D', E-E').

4.3.2) Upper-crustal seismicity

Earthquakes from upper-crustal sources are the most frequently recorded type of seismicity in both the PNSN relocated and Enhanced catalogs (Fig. 4.4). Upper-crustal seismicity is most frequently recorded in western Washington and Oregon between the Olympic Peninsula, Oregon Coastal Ranges and the Cascade volcanic arc. Upper-crustal seismicity is less frequently recorded in southern Oregon and east of the Cascades but is still present within focused clusters near the 1872 Entiat earthquake, and eastern Yakima fold and thrust belt in southern Washington (Fig. 4.1, Fig. 4.4).

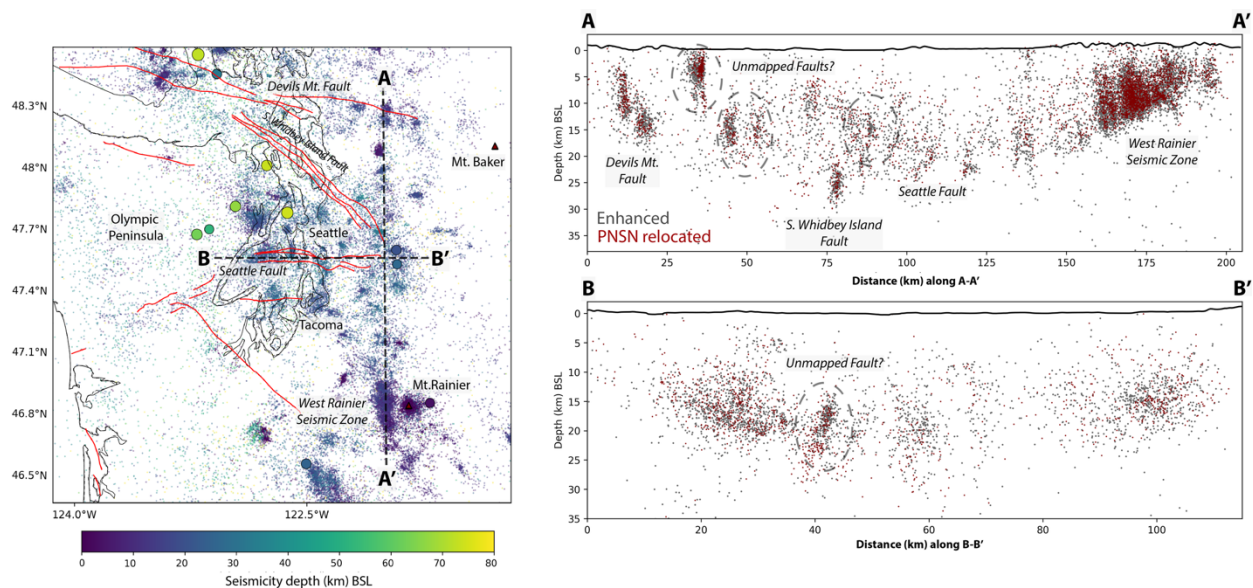


Figure 4.8: Crustal seismicity near the Seattle/Tacoma, WA area. Crustal faults from the CRESCENT CVM viewer in red lines. Both A-A' and B-B' cross sections show seismicity within 5 km of their respective profiles. Both profiles are measured in km. The Enhanced catalog features 7,519 and 2,811 events within 5 km of the A-A' and B-B' profiles respectively—over twice as many events in the PNSN Relocated catalog along both profiles (3,468 along A-A' and 1,213 along B-B'). Colored circles in map view are events with a magnitude of greater than 2.5.

We inspect the contents of both Enhanced and PNSN relocated catalogs along two transects near the Puget Sound region in western Washington (Fig. 4.8). The A-A' transect spans over 200 km between Puget Sound and the Western Cascade arc, where it intersects the Devils Mountain Fault, Southern Whidby Island Fault, Seattle Fault, and West Rainier Seismic Zone.

The B-B' transect spans roughly 100 km along strike of the Seattle Fault between the eastern Olympic peninsula to roughly 50 km east of the Seattle area. Within 5 km of each transect, we find that there over double the number of events within the Enhanced catalog compared with the PNSN relocated catalog.

The increased quantity of well-located events along both A-A' and B-B' transects reveals a series of lineations which resemble dipping fault planes between depths of 0 to roughly 30 km BSL (Fig. 4.8). Along the A-A' transect, we observe clear expressions of the Devils Mountain Fault, South Whidbey Island Fault, and West Rainier Seismic Zone, in addition to several crustal faults which are not contained in the CRESCENT Community Fault Model database. Interestingly, seismicity distributions are shallowest nearest to the edges of both transects and deepen towards the middle.

4.3.3) Volcanic seismicity: Mount St. Helens

Mount St. Helens is the most active and densely instrumented volcano in the Cascade Range. During the 2002 to 2024 study period, Mount St. Helens hosts a long-lived dome-building eruption from 2004 to 2008 (e.g. Sherrod et al., 2008) followed by a period of quiescence and persistent seismicity which continues to the present day. Seismicity during both the eruptive and quiescent periods are of highly varied character. Additionally, Mount St. Helens features a complex velocity structure (e.g. Kiser et al., 2020), varied site effects, and high-relief topography, making it a challenging environment to accurately detect and locate earthquakes in. Lastly, seismic signals from non-earthquake sources which may originate from rockfalls, debris flows, or other unknown types are frequently recorded at Mount St. Helens. These factors present an opportunity to stress-test our workflow in a challenging volcanic environment and evaluate its applicability towards a better understanding of volcanic seismicity.

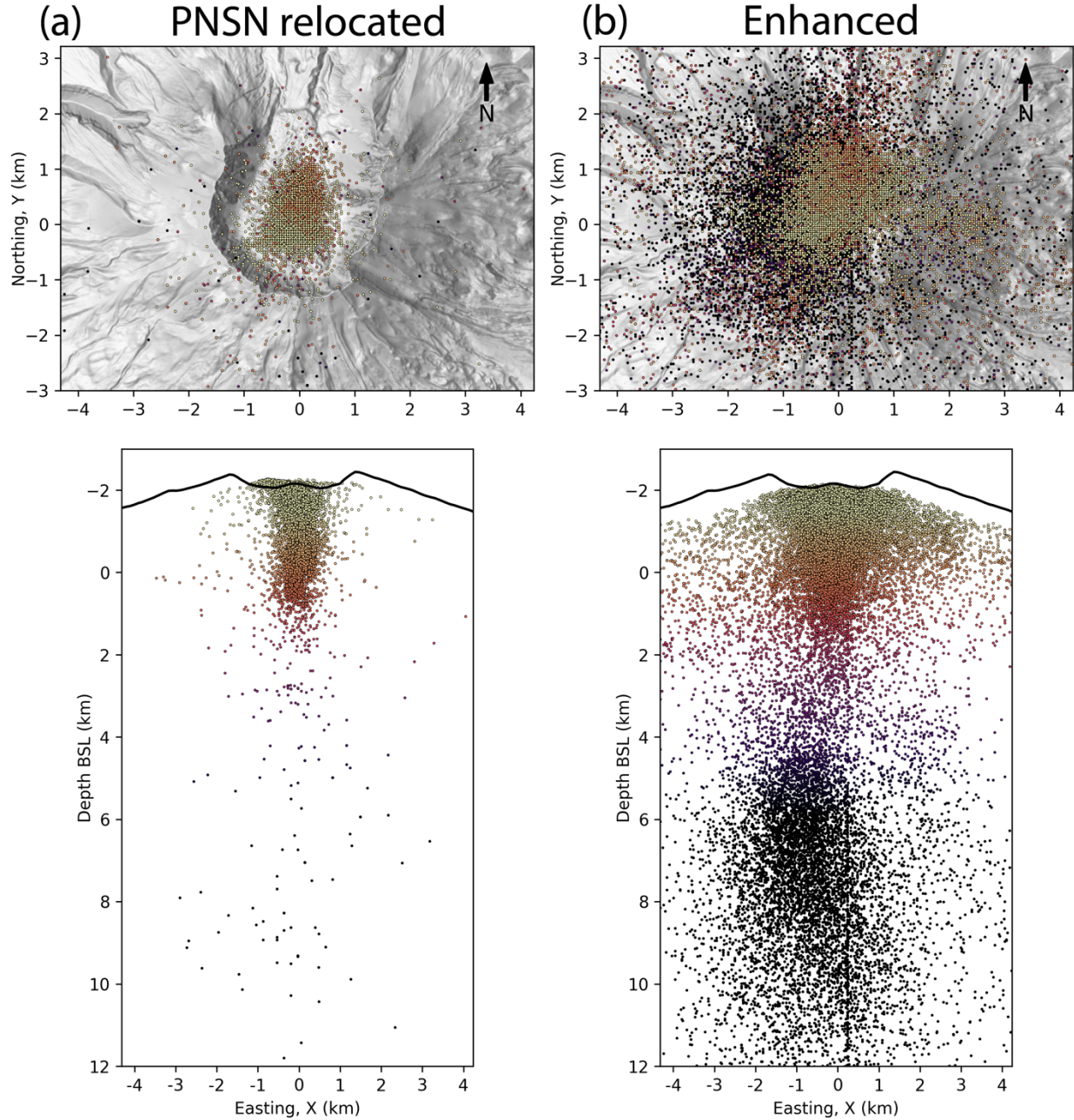


Figure 4.9: Events at Mount St. Helens from the 2004 to 2008 eruption. Map and along-depth comparisons of the (a) PNSN relocated and (b) Enhanced catalogs. In (a), we show all 9,677 PNSN events from this period, without filtering for maximum error or azimuthal gap. In (b), we show 40,184 events from the Enhanced catalog which have errors of less than 3 km and azimuthal gap of 200 degrees. It is questionable if the deep events beneath 4 km BSL in the Enhanced catalog are real.

We detect and locate 66,204 earthquakes within 5 km of the middle of the Mount St.

Helens crater between 2002 and 2024—a roughly 4.8-fold increase over the 13,783 events in the

PNSN relocated catalog. A total of 54,426 of the newly detected events occur during the 2004-2008 eruption. The 2004-2008 eruption is significant for being the first eruption at the Mount St. Helens volcano to have been fully captured by continuous seismic and geodetic monitoring equipment, prompting numerous geophysical studies that contribute to the understanding of eruptive dynamics (e.g. Sherrod et al., 2008; Mastin et al., 2008; Thelen et al., 2008; Dzurisin, 2015; Dzurisin, 2018; Wong and Segall, 2020). Intense and anomalous seismicity remain defining features of the 2004-2008 eruption—including over one million shallow drumbeat earthquakes recorded throughout the eruption, in addition to, thousands of shallow and sporadic volcano tectonic events during the vent clearing and shutting phases (Thelen et al., 2008; Dzurisin, 2015; 2018). Due to sheer quantity, most seismicity of the eruption remains uncatalogued by the PNSN analysts, who instead process a representative subset of events. An advanced automated detection and location pipeline, such as the one developed here, may be useful to collect observations which would remain uncatalogued by typical network methods.

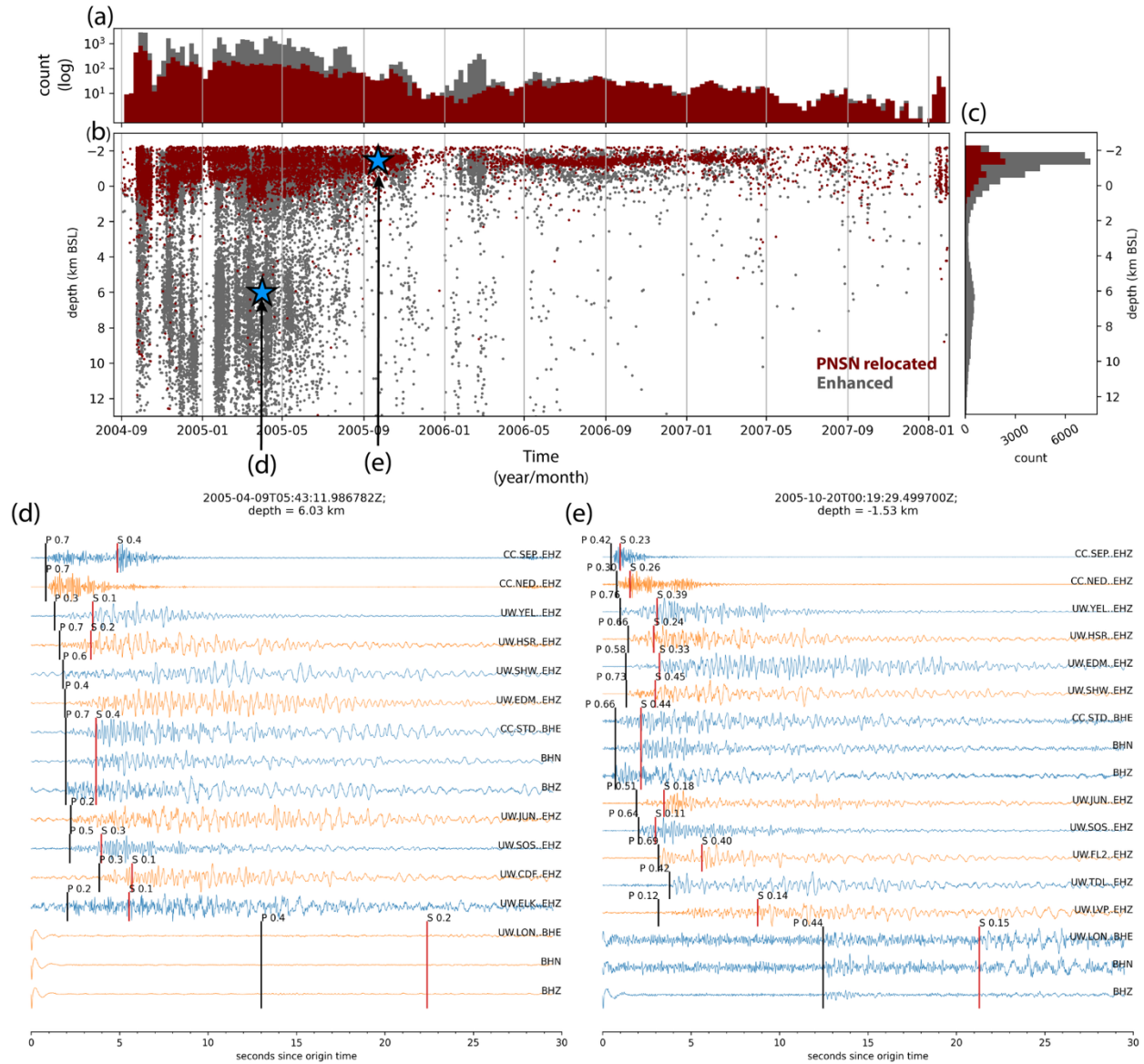


Figure 10: Timeline of depths for events at Mount St. Helens from the 2004 to 2008 eruption from both PNSN relocated and Enhanced catalogs. Weekly event rates in (a), depth of all events in (b), and depth distributions in (c). Example waveforms, phase picks, and their respective pick probabilities of a deep event at 6 km BSL in (d), and a shallow event at -1.53 km BSL in (e). The x axes of (d) and (e) are in seconds with respect to their event origin times. Stations are arranged from nearest to furthest from the event epicenter. Both events have very small uncertainties (less than 0.3 km in x, y, and z). All waveforms shown are bandpass detrended and filtered between 1 to 20 Hz. Their locations are shown with blue stars in (b).

We inspect the geographic, along depth, and temporal distributions of events from the eruption in both the PNSN relocated and Enhanced catalogs (Fig. 4.9, Fig. 4.10). Nearly all

events in the PNSN relocated catalog collapse to a shallow, roughly cylindrical source volume between -2.2 and 0.5 km BSL directly beneath the lava dome complex (Fig. 4.9). By comparison, events in the Enhanced catalog are considerably more scattered. Surprisingly, we find over 15,783 events from the eruption in the Enhanced catalog located at depths of greater than 4 km BSL (~6.6 km below the crater floor). Most of the deep earthquakes located between 4 and 12 km BSL occur within a diffuse cluster that is consistent with the location of upper crustal magma reservoir, which is thought to have fed both the 1980-1986 and 2004-2008 eruption (e.g. Kiser et al., 2020) (Fig. 4.9B, Fig. 4.10).

4.4) Discussion

The Enhanced earthquake catalog presented here contains a vast number of observations relevant to a wide breadth of geologic contexts. The results presented here represent an early iteration of what will become the CRESCENT seismicity catalog. Our preliminary results show that the automated detection and location workflow we have developed here is useful for cataloguing previously undiscovered events in regions near the subducting slab, within volcanic magma transport systems, and in upper-crustal tectonic faults. Here, we discuss some features of the new Enhanced catalog and comment on some potential directions for future improvements and scientific inquiry.

4.4.1) Seismicity near the subducting plate interface

Although we observe a significant relative increase in the number of detected and located events near the plate interface, these events make up less than 5 percent of all seismicity in the Enhanced catalog. Due to the lack of offshore seismic stations and azimuthal gap limitations, seismicity in the updip region of the plate boundary may be underrepresented. In addition, it is unclear whether the newly detected events occur along the plate interface, within the slab, or in

the nearby overriding crust. The lack of recorded seismicity along the plate interface is not significantly improved by our workflow. This result provides further evidence that the plate interface is largely aseismic and does not produce many small magnitude events, despite an abundance of evidence for M9+ rupture (e.g. Heaton and Kanamori, 1984; Atwater 1987; Satake et al., 1996; Small et al., 2025).

Nonetheless, newly recorded seismicity near the plate interface provides some possibilities for future research, especially when combined with high-precision relocation, cross-correlation, and focal mechanisms. A few examples include deformation near the source regions of the 1949, 1965, and 2001 M6+ intraslab earthquakes and searching for repeating earthquakes with highly similar waveforms.

4.4.2) Upper crustal seismicity

Roughly 95 percent of all events both in the PNSN Relocated and Enhanced catalogs occur within the overriding North American plate from crustal tectonic or volcanic sources. Additionally, many upper crustal faults can host damaging earthquakes (M 5+) near densely populated areas (e.g. 1981 M 5.5 near Elk Lake, WA; 900 AD M 7+ near Seattle-Tacoma, WA; 1872 M 7+ near Entiat, WA; Weaver et al. (1986)). Despite high rates of seismicity from the upper crust and high potential for seismic hazard, upper crustal faults are poorly characterized owing to limited surficial exposure and long recurrence intervals of moderate to high-magnitude earthquakes. Clusters of precisely located hypocenters from enhanced seismicity catalogs can be helpful to identify and characterize active faults that lack subaerial exposure (e.g. Alongi et al., 2025) or are too small to be detected by tomographic imaging methods.

A simple inspection of our early results indicates that upper crustal seismicity, particularly near Puget Sound, can be used to identify fault structures at depth (Fig. 4.8).

Improved location precision through a refinement step in addition to fault plane solutions could be very helpful to characterize the structure and modes of strain partitioning of the upper crust.

4.4.3) Volcanic seismicity

The machine learning-based workflow can detect substantially more events during volcanic eruptions than typical network methods. However, due to unconventional waveform character, phase pick quality is questionable, and may translate to poor location quality and erroneous patterns when viewed collectively. One such example are the high levels of deep events from 4 to 12 km BSL between during the first two years of the eruption: 2004 to 2006. At Mount St. Helens, deep seismicity between is thought to reflect pressure changes within the upper crustal reservoir due to either evacuation during, or recharge during or after an eruption (e.g. Moran, 1994; Musumeci et al., 2002; Moran et al., 2008; Dzurisin, 2015; Hirao et al., 2025).

Due to the potential significance of deep seismicity during the eruption, it is useful to closely inspect the accuracy of the phase picks and location uncertainties. We inspect the waveforms and phase picks of a selection of events from the eruption which have a location uncertainty of less than 0.3 km and azimuthal gap less than 150 degrees—two examples including deep event on April 9, 2005, at 6 km BSL and a very shallow event on October 10, 2005, at 1.5 km ASL are shown in Figure 10. Both events feature waveforms characteristics that are visibly different from each other and of those from conventional tectonic earthquakes—note the emergent P-wave arrivals, long duration of coda phases, and lack of easily identifiable S-wave arrivals. Phase pick quality is mixed--P-wave arrivals appear to be accurately picked; however, the quality of most S-wave arrival picks is questionable. S-wave arrivals are difficult to identify visually. Moreover, most stations that recorded the eruption were only equipped with a

single vertical channel, so it is difficult to accurately identify and shear waves since horizontal polarization of ground motion is not fully captured. In both events, the S-P time difference is nearly identical on stations YEL, HSR, and STD. In the October 10th example, the S-P time differences are consistent with moveout from a shallow source. However, in the April 9th example, S-P time differences on station SEP, which captures a near-vertical ray path, reflects a source that is much deeper. The S-wave pick on SEP could simply have been triggered by a high-amplitude phase that followed the P-wave arrival. A comparison of these two events suggests that dubious S-wave arrival picks on as few as one or two stations can drastically change the depth solution and that small uncertainties indicate that the observations are explained by a small source volume. Importantly, if this style of mis-picking is systematic, it could suggest that the deep cluster near the proposed magma reservoir at 4-12 km BSL is not real.

Our simple checks highlight the fact that the earthquake location process, and therefore interpretation, can be sensitive to poor quality phase picks. In addition, both events have small uncertainties, suggesting that the quality of an event location is not always reflected in the magnitude of its uncertainties or misfit. However, the quantity of detected events and relatively good quality of P-wave picks suggests that our automated workflow may still provide useful results. Due to the uniqueness and variation of volcanic seismicity, we suggest that the user apply the workflow to volcanic environments with carefully selected criteria determined on a case-by-case basis.

4.4.4) Future improvements

To improve the quality of the catalog and its contents, further optimizations to the workflow may be applied towards the detection, association, and location steps. The quality of a

phase pick from the PNW-EQTransformer is generally proportional to its output probability. Phase picks of low probability, whether due to ambiguous waveform character or low signal to noise ratio, have a higher chance of having incorrect phase classification, arrival time error, or being a false detection. In addition, it is likely that pick quality can vary across different geologic settings and sensor types. Thus, in future versions, it is worth considering minimum phase-pick quality criteria taking these factors into account. The effects of outlier phase picks on location solutions can partially be mitigated through use of the Equal Differential Time (EDT) weighting scheme in NonLinLoc; however, this assumes a relatively high ratio of good-quality picks that are easy to fit, relative to outlier mis-picks. As demonstrated in Section 4.4.3, the EDT system is not always effective at dealing with outliers. Thus, other forms of quality criteria should be considered.

Inspection of event locations contained in the PNSN catalog demonstrate that the NonLinLoc solver and parameter search of 3-D travel time grids from the CRESCENT CVM can solve for precise locations and meaningful uncertainties in most environments—including very challenging volcanic settings such as Mount St. Helens, WA. We find that our catalog is relatively free of coarse gridding and constant-depth striping artifacts, which are often found in network catalogs that use a 1-dimensional velocity model. Nonetheless, improvements to event precision can allow events to better cluster, which can improve the ability to interpret and characterize fault and volcanic plumbing system architectures. In a future version, we intend to improve event precision and clustering by including in our workflow the NonLinLoc Source-Specific-Station correction terms (SSST) algorithm to help refine location solutions by reducing the effects of unmodelled velocity structure (Lomax and Savvaidis, 2021). The SSST algorithm

is useful for our application, since it can generate corrected travel-time grids which can then be applied towards future updates of the catalog.

Earthquake magnitudes are a key piece of information for assessing seismic hazard potential in a particular region. In addition, magnitudes are important for understanding the nature of earthquake spatiotemporal clustering—whether they are swarm-like or contain clear mainshock-aftershock sequences. Estimating event magnitudes with a consistent scale over a multidecadal timespan and broad geographic area is challenging, owing to varied attenuation and instrumental types. A key challenge is finding a magnitude scale which can accommodate amplitude data measured from stations which may lack horizontal components, which are especially prevalent in the first few years of the study term. A solution may be to solve for region-specific attenuation and station-correction terms, such as the approach described by Trugman (2023).

In the current iteration, the geographic extent of the Enhanced seismicity catalog covers the states of Washington and Oregon. Due to the high computational and financial expense to run PNW-EQTransformer model, our pick database is limited to southwestern Canada south of the 50-degree latitude, Washington, and Oregon. In addition, our velocity model does not extend north of the 50-degree latitude. Seismicity in northern California and Nevada is underrepresented because we do not have yet a comprehensive dataset of phase arrivals from stations south of the Oregon-California and Oregon-Nevada borders. Enhanced seismicity catalogs in northern Cascadia and northern California (e.g. the Mendocino Triple Junction) are of high value for basic scientific research and hazard assessment. We intend to include these regions into later versions of the catalog by implementing a combination of different velocity models for their specific

regions and outsourced phase arrival picks from different ML models (i.e. the global pick database of Ni et al., 2025).

4.4.5) Enhanced seismicity catalogs for the end user

The scientific impact of an earthquake catalog dataset is determined by how it is used in future studies by other researchers. We have yet to decide how our Enhanced catalog will be distributed to the public and how it may be updated. Moreover, we think it is useful to allow the end user to modify the workflow with the phase detector, associator, velocity model, and location solver of their choice. A modular workflow can be helpful since each step of the enhanced catalog pipeline has its strengths and weakness, which are likely context specific. Because automated seismicity cataloguing workflows lack analyst verification, comparison between different permutations of each method and redundancy can be helpful for validation. We may derive inspiration from the USGS earthquake catalog. One of the most important features is its ease of access and continuous updates. The USGS catalog can be easily navigated through a graphical user-interface which can be used to filter and select for events of specific criteria. Moreover, it contains both essential and complimentary information including event timing, locations, uncertainties, phase picks, magnitudes, and for many events, focal mechanism and moment tensor solutions. These origin products can be applied towards any study and can be complemented with relocations, different velocity models, and even be applied towards workflows for template matching and machine-learning model training

4.5) Conclusion

We develop an automated workflow to detect and locate previously undiscovered seismicity throughout the Pacific Northwest, USA between the 2002 and 2024 study period. Here, we obtain phase picks from a specially trained variant of the EQTransformer, which are

then associated to events with the PyOcto associator. The resulting events are then located using the NonLinLoc package to search for optimal location parameters within 3-dimensional travel-time grids that are solved using the CRESCENT Community Velocity Model. With our automated workflow, we detect and locate to a reasonable quality, over 3 times as many events as contained in the PNSN catalog during this same period. Most newly detected seismicity is of upper-crustal or volcanic source origin. Despite improved detection thresholds, the subducting slab remains relatively aseismic. To broaden the utility of the enhanced earthquake catalog as a dataset, future work should be directed towards expanding the study region into Canada and Northern California, refinement of location solutions, and magnitude attenuation term calibration which will allow for accurate magnitude estimation.

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CHAPTER V

CONCLUSIONS AND FUTURE DIRECTIONS

In this dissertation, I produce a series of enhanced earthquake catalogs which contain thousands of previously hidden small-amplitude earthquakes in the Pacific Northwest, which are not included in the catalog from the Pacific Northwest Seismic Network. We then leverage these catalogs to observe and characterize the evolution of volcanic and faulting structures across several decades at an unprecedented resolution.

In Chapter 2, I inspect volcano-tectonic earthquakes at Mount St. Helens after the end of the last eruption in 2008 until 2024. I find tens of thousands of shallow volcano-tectonic events at shallow depths (0-2 km BSL; 2.2-4.2 km below the surface), beneath the historical lava dome complex. Elevated levels of shallow and deep earthquake rates in periodic swarms, beginning in 2016, suggest that the Mount St. Helens magmatic plumbing system is repressurizing 8 years after the last eruption. Perhaps the most striking discovery are the repetitive, shallow burst-like swarms, which may indicate valve-like response of shallow conduit to accumulation of volatile pressure. Future work may be directed towards a more comprehensive understanding of the source mechanisms of volcano-tectonic earthquakes, perhaps through waveform modelling. Lastly, it may be helpful to develop a model of volcano-tectonic seismicity that can be more directly related to the vast amount of petrologic and experimental literature.

In Chapter 3, I analyze non-volcanic seismicity in the faulting systems adjacent to the Mount St. Helens between 2008 and 2024. Here, we find that non-volcanic seismicity clusters into four distinct groups of roughly equal distance from volcano. I do not find evidence for direct interaction between each of the four non-volcanic seismicity groups or with volcano-tectonic events. However, mapped geologic features and fault-plane solutions from the four seismicity

groups suggest that the Mount St. Helens volcano may be underlain by a sub-orthogonal fracture zone. Further work could be directed towards modelling of the stress field in consideration of background tectonic stresses, a viscoelastic magmatic reservoir, and a lithologic boundary (i.e. Siletzia basalt and Spirit Lake Batholith granodiorite).

In Chapter 4, I create an enhanced seismicity catalog that spans the Pacific Northwest states of Oregon and Washington between the years 2002 and 2024, which is intended to be an initial version of the authoritative CRESCENT Earthquake Catalog. I create an automated workflow to use phase picks from a specially trained variant of the EQTransformer phase detector. These are then associated with a new high throughput associator, Pyocto. I then locate all events using the new CRESCENT Community Velocity Model. The new workflow can detect and locate nearly four times as many earthquakes across the Pacific Northwest. The CRESCENT Earthquake Catalog is specifically designed to enable future work towards understanding the nature of seismicity across the Pacific Northwest. To maximize the usefulness of later versions of this catalog, future work should be directed towards more deliberate selection of phase picks, a location refinement step, region-specific magnitude scaling, and a way to distribute the catalog to individual users.

In total, I detect, locate, and catalog over 100,000 previously undiscovered events across the Pacific Northwest at an unprecedented precision. To our knowledge, these three catalogs represent the first enhanced seismicity catalogs in the Pacific Northwest to span several decades in duration. All three chapters required enormous computing resources to complete. These studies, through implementation of big data processing with high precision detection and location procedures, reveal numerous avenues for future research, and highlight the vast amount of information that may be hidden in archived seismic data.