

Triggered Shallow Megathrust Slip following the 2020 M_w 7.6 Sand Point (Alaska) Strike-Slip Earthquake Explains its Tsunami

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Abstract On October 19, 2020, the M_w 7.6 Sand Point earthquake occurred near the Shumagin Islands in Alaska. Moment tensor solutions suggest oblique strike-slip faulting, but the earthquake was accompanied by an uncharacteristically large tsunami. Prior models explain this with co-seismic megathrust slip release, or shallow very slow thrust slip in the overriding plate. Here, we analyze the most complete land-based data set available for this earthquake including teleseismic and high-rate GNSS waveforms, static GNSS and InSAR deformation, and relocated aftershock distributions. We test two models: a single strike-slip fault model, and a two-fault model consisting of strike-slip and thrust segments. While both satisfy the observations, neither model reproduces the tsunami well. To locate the tsunami source, we performed a systematic test of thrust-slip models. This suggests that delayed, very long period concentrated aseismic slip on the shallow megathrust caused the tsunami without generating measurable surface displacements. This occurred minutes after the mainshock, at a location with mainshock-induced Coulomb stress increase, suggesting a triggered tsunami earthquake. The partitioning of strike-slip and triggered megathrust slip offers opportunities to better understand tsunami earthquakes, and—if prevalent for megathrust earthquakes—might impact earthquake and tsunami hazards.

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1 Introduction

The M_w 7.6 Sand Point earthquake struck on 19 October 2020, as part of a sequence of large earthquakes in the Shumagin and Semidi segments of the Alaska Aleutian megathrust, offshore of the Alaska Peninsula (Fig. 1). The sequence initiated with the July 2020 M_w 7.8 Simeonof earthquake, a thrust event that released most of its slip between ~20–40 km depth on the plate interface east of the Sand Point earthquake (Crowell and Melgar, 2020; Liu et al., 2020; Ye et al., 2021; Xiao et al., 2021, Fig. 1). Following the October 2020 Sand Point earthquake, the sequence continued with the July 2021 M_w 8.2 Chignik earthquake, releasing megathrust slip at depths similar to and further east from the Simeonof earthquake (Elliott et al., 2022; Liu et al., 2022; Ye et al., 2022), and triggered by stress changes due to that event (Elliott et al., 2022). In 2023, the M_w 7.2 Western Shumagin earthquake, a thrust event with a hypocenter at 25 km depth, released further strain on the deeper parts of the megathrust west of the Sand Point and Simeonof earthquakes. The most recent event in the sequence is the 16 July 2025 M_w 7.3 Sand Point earthquake, which occurred

east of the 2023 Western Shumagin thrust event. In contrast to the three megathrust events of this sequence, the 2020 and 2025 Sand Point earthquakes had oblique strike-slip focal mechanisms with ~N-S or ~E-W striking focal planes (Fig. 1, GCMT, Ekström et al., 2012). Aftershock locations show that the 2020 Sand Point earthquake ruptured a ~N-S plane (Jie et al., 2026), while the 2025 event appears to have ruptured the ~E-W plane at shallower depth. Unlike the 2020 event, the 2025 event did not produce a notably large tsunami. Most of the large ($M_w \geq 4.8$) aftershocks of these three major events share similar focal mechanisms with the mainshocks, with two exceptions: an M_w 5.4 event ruptured a vertical or horizontal fault and an M_w 4.9 normal-fault earthquake occurred nearly 14 and 32 hours after the 2020 M_w 7.8 mainshock, respectively.

The 2020 Sand Point earthquake, which was advanced by stress changes from the Simeonof earthquake and possibly located at a transition from high to low coupling on the megathrust (Fig. 1; Herman and Furlong, 2021; Xiao et al., 2021) (see Supplementary Text 1 for tectonic background), ruptured on a fault located mainly or entirely within the downgoing plate (Herman and Furlong, 2021; Jiang et al., 2022). In addition to its

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unusual location and mechanism, the 2020 Sand Point event was accompanied by a larger tsunami than the Simeonof event (e.g., [Bai et al., 2023](#); [Santallanes et al., 2025](#)), despite being smaller magnitude, strike-slip instead of thrust mechanism, and rupturing a similar depth range.

The unusually large tsunami has been an intriguing puzzle. Strike-slip events at depth, even oblique ones, in general do not generate large tsunami; yet this event did, measuring 5 cm at nearfield DART buoys, and >25 cm in Hawaii ([Bai et al., 2023](#)). One possible explanation for the tsunami could be an undersea landslide triggered by the shaking. The chief problem with this hypothesis is a current lack of evidence for or against the existence of such a landslide. Another possible explanation would be that the Sand Point event also involved slip on the megathrust, perhaps at shallow depth. If such slip occurred, it has the potential of explaining the tsunami, as the megathrust slip would displace the ocean floor and water column more effectively than a purely strike-slip event would. [Santallanes et al. \(2025\)](#) presented a model that suggests that slip on a strike-slip fault was accompanied by 30 s delayed megathrust slip below 20 km depth that propagated at a very slow 1 km/s. [Bai et al. \(2023\)](#) presented a much more complex slip model for the event, involving three faults, none on the megathrust. Their model includes the main oblique strike-slip fault in the subducting plate, and an oblique normal fault and a westward dipping slowly slipping thrust fault in the overriding plate. The first two faults could explain all seismic and geodetic data, while the third fault was a shallow structure added to explain the tsunami. Although they could explain the tsunami by adding the unusually oriented thrust source, their model also required release of 15 m of slip over 5 minutes, which is not easy to explain, especially because of the requirement that the tsunamigenic source could not have a measurable impact on either seismic or geodetic observations. Furthermore, a new machine-learning-based catalog shows no aftershocks in the overriding plate for the first 7 days after the mainshock ([Jie et al., 2026](#)), suggesting large co-seismic deformation in the overriding plate is unlikely.

In this paper, we use the most complete land-based data set to date and apply finite fault inversions, seismic waveform modeling, tsunami modeling, and Coulomb stress change modeling to (a) infer a plausible finite fault model for the 2020 Sand Point earthquake, and (b) exhaustively investigate the cause of the tsunami and its relationship to this earthquake. We leverage three-component co-seismic GNSS (Global Navigation Satellite System) offsets calculated from daily GNSS positioning solutions, coseismic displacements from InSAR (Interferometric Synthetic Aperture Radar) interferograms, high-rate GNSS waveforms from 1 sample-per-second (sps) GNSS observations, and teleseismic waveforms to infer a coseismic finite fault slip model for the 2020 M_w 7.6 Sand Point earthquake. However, this strike-slip-only slip model underpredicts the observed tsunami at nearby DART buoys. Similarly, our inversion for simultaneous strike-slip and megathrust slip results in a model that also cannot constrain the tsunami well,

suggesting land based observations might lack sensitivity to any trench-proximal megathrust slip of sufficient amplitude to generate the tsunami. When we analyze long-period seismic data, we find no signature of the previously hypothesized (e.g., [Santallanes et al., 2025](#)) submarine landslide as cause for the tsunami. Thus, to further assess whether slip release on the megathrust could explain the observations, instead of using tsunami observations in an inversion to determine its source ([Santallanes et al., 2025](#)), or fitting a parameterized seafloor deformation pattern to drive the tsunami model ([Bai et al., 2023](#)), we experiment with distributed unit sources of scaled megathrust slip that drive a tsunami model to identify the location, origin time, and rise time of the tsunami source. We find that the most likely tsunami source resulting from this experiment falls into a narrow, shallow near-trench region on the megathrust, updip of the Sand Point event, but rupturing up to 8 minutes after the Sand Point origin time. Coulomb stress change modeling of the strike-slip-only Sand Point model predicts substantial stress increase for this location, suggesting a causal relationship between mainshock and modeling-supported tsunami slip location.

2 Data

Our data analysis and methodology follow the approaches detailed in [Xiao et al. \(2021\)](#) and [Elliott et al. \(2022\)](#). Raw GPS data were processed in point positioning mode using the GIPSY/OASIS goa-6.4 software and JPL non-fiducial (`_nf`) orbit and clock products, and single station ambiguity resolution ([Bertiger et al., 2010](#); [Zumberge et al., 1997](#)) to estimate daily positions. Daily solutions for all stations across Alaska are then aligned with the ITRF2014 reference frame. We used the VMF1GRID ([Boehm et al., 2006](#)) a priori tropospheric dry and wet path delay values, and estimated a residual correction to the wet zenith path delay (including an azimuthal gradient) using the VMF1GRID tropospheric mapping functions. We applied second order ionospheric corrections using the IGS final ionosphere models (IONEX files). Finally, the individual daily station positions were combined to a network solution, and aligned with the ITRF2014 reference frame using a 7-parameter transformation.

Because the Sand Point earthquake occurred in the midst of postseismic relaxation caused by the 2020 M_w 7.8 Simeonof earthquake, we estimated static GNSS displacements from daily position solutions using a parametric fit to the time series. Two stations have data gaps (AC21, ~6 weeks before Sand Point, and AC41, 9 months with Sand Point in the middle), but we are able to recover coseismic estimates even for these stations. We fit each station's time series from decimal year 2018.2 through 2021.85 with a linear trend, annual and semi-annual seasonal components, coseismic displacements, and logarithmic postseismic relaxation terms. We estimated coseismic displacements for the Simeonof earthquake, the Sand Point earthquake, and the 2021 M_w 8.2 Chignik earthquake. We included logarithmic relaxation terms starting from the time of the

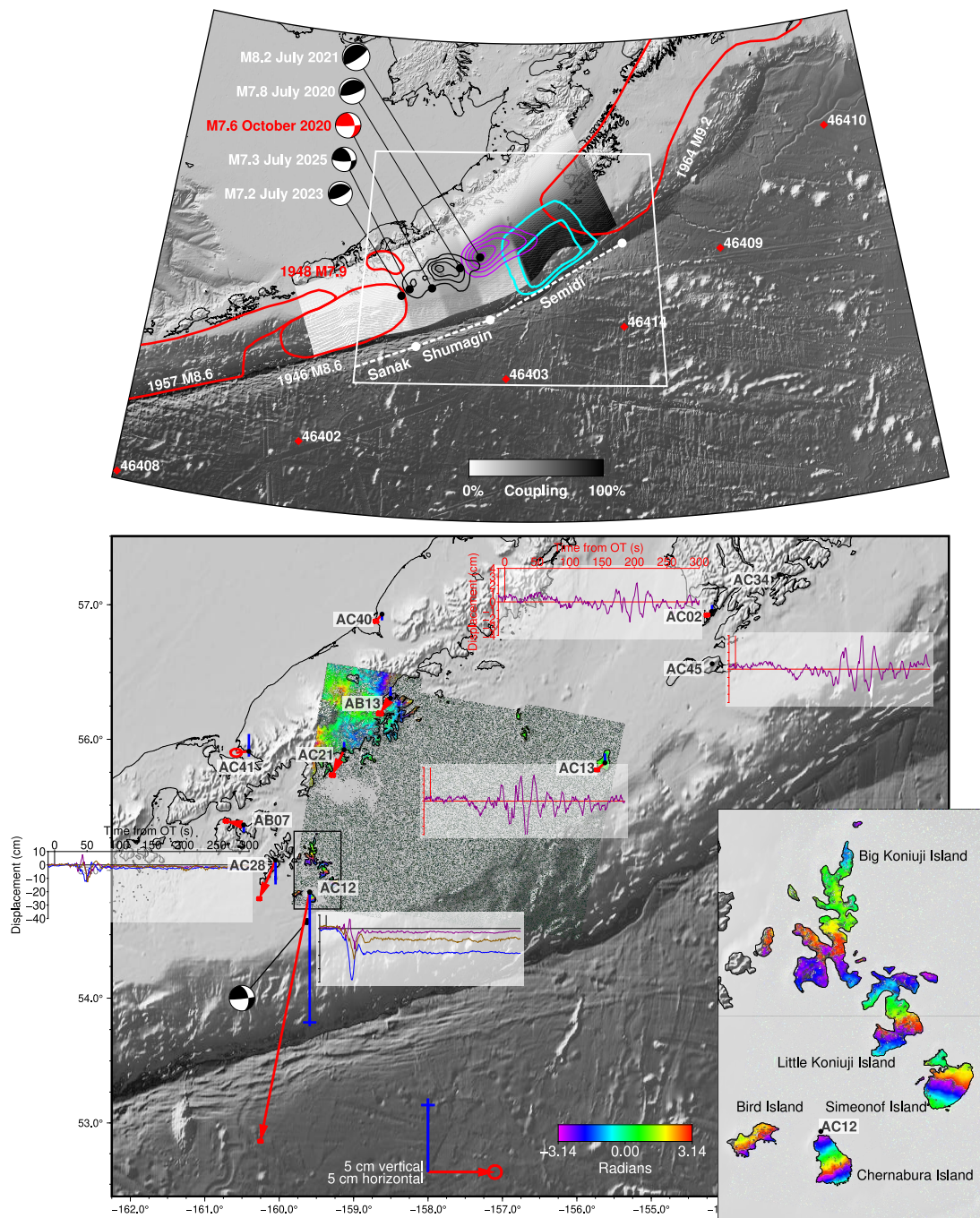


Figure 1 Overview of tectonics and data used. **Top** map shows the focal mechanisms of the five events of the 2020-2025 Alaska earthquake sequence: M_w 7.8 July 2020 Simeonof earthquake, M_w 7.6 October 2020 Sand Point earthquake, M_w 8.2 July 2021 Chignik earthquake, M_w 7.2 July 2023 Western Shumagin earthquake, and M_w 7.3 July 2025 Sand Point earthquake. Red contours show outlines of historic ruptures (Davies et al., 1981; López and Okal, 2006). The colored contours show: black - 0.5 m slip contours from 0.5 m to 2 m for the Simeonof earthquake (Xiao et al., 2021); purple - 1 m slip contours from 1-5 m for the Chignik earthquake (Elliott et al., 2022); cyan - eastern model 1-2 m contours of slip estimates for the 1938 M_w 8.3 earthquake (Freymueller et al., 2021). White dashed lines and annotations identify segments with different properties of the megathrust. DART buoys marked by red diamonds. Shaded coupling values are from Xiao et al. (2021). **Bottom** zooms into white box showing co-seismic GNSS offsets with red arrow-tipped vectors (horizontal) and blue bar-tipped vectors (vertical), where tips represent 95% confidence ellipses. The time series near GNSS stations show 300 seconds of high rate 1-Hz position solutions. We show two sets of these time series that share the same time scaling, but differ in the displacement scaling: black coordinate systems show 10 cm of displacement (AC12, AC28; for figure clarity only the latter is labeled), while red coordinate systems indicate 1 cm of displacement (AC13, AC45, AC02; for figure clarity only AC02 axes are labeled); for these we only show the east component as it shows the surface wave most clearly. Colors of time series traces, magenta, blue, and orange, respectively, show east, north, vertical displacements. The vertical bar along the time series marks the earthquake origin time. The Sentinel-1 interferogram spans 2020-10-04 through 2020-10-28. Lower right inset shows zoomed-in InSAR phase gradients for the eastern Shumagin Islands (location indicated by the black box in the main map).

Simeonof and Chignik earthquakes, with estimated relaxation times of 0.025 years for Simeonof and 0.005 years for Chignik. The displacements estimated for the Sand Point earthquake were not sensitive to reasonable changes in the relaxation times for the postseismic transients. We assumed a white noise error model (using the formal uncertainties for each solution), and rescaled the estimated parameter uncertainties so that the reduced chi-square statistic for the solution was equal to 1.0. Final horizontal displacement uncertainties were on the order of 0.5–1 mm, and vertical displacement uncertainties were ~1.5–2 mm.

We generated high-rate GNSS waveforms in a local north-east-vertical coordinate system from 1 sps observations using GipsyX (Bertiger et al., 2020) following the same approach as in Xiao et al. (2021). The obtained waveforms were filtered with a second order lowpass filter with 0.4 Hz cutoff frequency to reduce the high frequency noise. In the inversion, the vertical GNSS waveforms at AC40 and AC42 sites were not utilized, and the vertical waveform of the AC13 was assigned a 33% weight of the horizontal components in the slip inversion due to their lower signal-to-noise ratios (SNR). The horizontal and vertical waveforms of other sites were assigned the same unit weight.

We produced two-pass interferograms using GMT-SAR (Sandwell et al., 2011) from Sentinel-1 (Torres et al., 2012) observations. The interferogram on path 102 frame 407 spanning 2020-10-04 to 2020-10-28 is subject to the least atmosphere disturbances and captures the co-seismic offsets on the Shumagin Islands best (Fig. 1). We also processed the neighboring path 73 frame 407 that covers the remaining islands to the west, but found no clear signal. Although the ascending path 153 frame 177 covers the entire island group, we did not find a pair of SAR acquisitions (respectively testing two SAR scenes prior to and following the earthquake) that yielded a clear signal in the interferogram. Due to the above mentioned postseismic deformation, we did not search over longer time intervals to avoid biasing estimates with components of that signal. As the islands are disconnected patches of coherent phase, we unwrap each separately and create line-of-sight displacements using SNAPHU (Chen and Zebker, 2002). Since Chernabura Island hosts the continuous GNSS site AC12 (Fig. 1), we projected the estimated co-seismic displacement at that site into the satellite LOS and aligned the InSAR-inferred LOS displacement with the GNSS measurement to use in the inversion. All other islands are treated as “floating” and we estimated a phase ambiguity parameter in the slip inversion to align the observed and model-predicted phases (Xiao et al., 2021). All LOS observations were downsampled with a uniform sampling algorithm.

We accessed 39 P and 27 SH global seismic waveforms with good azimuthal coverage and high signal-to-noise ratio, between 30° to 90° epicentral distance. Raw waveforms are band-filtered between 0.001 and 5 Hz to retain the frequency range of interest and suppress high-frequency noise above 5 Hz. The filtered waveforms were then transformed to ground displacements with a narrower bandpass filter of 0.01 to 1 Hz to remove

long-period oscillations and further reduce noise. P and SH waves were resampled to 5 sps to improve computational efficiency.

We used the first 3 hours of water level observations from DART buoys 46402, 46409, and 46414 (Fig. 1) for the time after the earthquake. A 3-hour moving average filter was applied to DART buoy records to remove low-frequency variations associated with tides and isolate the short-period tsunami signal (e.g., Nicolsky et al., 2016).

We relocated more than 12,000 earthquakes from January 2020 to February 2022 in this region to investigate aftershocks and background seismicity. These events were detected and initially located by the Alaska Earthquake Center (AEC). Following Wei et al. (2021), we use tomoDD (Zhang and Thurber, 2003), a double-difference relocation package with 3-D ray tracing, to relocate all the events. The 3-D velocity model, which is fixed during the relocation inversion, is adopted from a regional tomography model (Wang et al., 2024) constrained by seismic data from the Alaska Amphibious Community Seismic Experiment (AACSE), EarthScope USArray TA stations, and the Alaska regional network. Compared to previous regional and global models, this model significantly improves the fore-arc structure thanks to the data from AACSE ocean-bottom seismographs. Consequently, the earthquake hypocenters, particularly the absolute depths, are better determined in this study.

3 Modeling Methods

3.1 Coseismic Slip Inversion

In this study, we developed two distinct finite fault slip models with different seismogenic faults to explore the tsunami source. For both models, the earthquake initiated from our relocated hypocenter (54.565°N, 159.666°W, at a depth of 34 km). However, in the first model, the rupture was assumed to only propagate on a strike-slip fault with no coseismic slip occurring along the megathrust interface, while the second model allowed the rupture to propagate onto the megathrust interface with some delay (these models are hereafter referred to as the “strike-slip-only” and the “strike-slip and slab interface” finite fault models).

To determine the strike-slip fault geometry, we tested three different 3D fault geometry parameterizations (USGS/GCMT, Jiang et al., 2022, and our own relocation) with the full dataset described above. Slip inversions with these geometries yield the respective misfits presented in Table 1. The Jiang et al. (2022) fault geometry is based on static geodetic observations, which seem to call for a lower dip angle, while the teleseismic waveforms call for a steeper dip. Our geometry, based on mainshock relocation, achieved the best teleseismic waveform fits. However, the static GNSS offsets are not well predicted using this geometry, with a maximum GNSS misfit of -3.7 cm, which is ~6 times larger than those resulting from the USGS/GCMT geometry. Thus, we choose the USGS/GCMT geometry to balance geodetic and seismic observations. This geometry is consistent with our aftershock relocation, which sug-

gests most of the early aftershocks within a day of the mainshock occur along a similarly striking and dipping plane (Fig. 2). We use local averages of Slab2 (Hayes et al., 2018) for the megathrust geometry without curvature.

With the above fault geometries, we explored the kinematic slip models with the same nonlinear wavelet-based inversion as Xiao et al. (2021) and Elliott et al. (2022), including a modification to accommodate “floating” InSAR patches by estimating parameter values for missing phase cycles. The strike-slip fault and the slab interface were divided into 135 and 195 patches, respectively, each with a size of 10 km along-strike by 10 km along dip. For each subfault, we simultaneously estimated slip amplitude, rake angle, rupture velocity, and an asymmetric rise time function (Ji et al., 2003; Shao et al., 2011; Xiao et al., 2021) by matching the dynamic waveforms and static geodetic offsets. The maximum slip amplitude on each subfault was set to be 4 m for the strike-slip fault and 1.5 m on the slab interface. Rake angles ranged from 240° to 120° in increments of 3° on the strike-slip fault and from 70° to 110° in 2° intervals on the slab interface. The rupture velocity was constrained to between 1.0 to 4.5 km/s. Since we constrained the start and end times of the asymmetric rise time function from 0 to 4.8 s in 0.6 s intervals, the slip duration on each subfault was therefore limited to 0 to 9.6 s. Green's functions for each subfault were calculated with the 1-D layered velocity model from LITHO 1.0 model (Pasyanos et al., 2014), potentially impacting teleseismic waveforms fits for an earthquake this close to the shelf edge as the dipping sea floor introduces large long-period late signals (Wiens, 1989).

3.2 Tsunami Modeling

To model tsunami propagation due to seafloor deformation, we use a numerical model that solves the nonlinear shallow water equations through finite difference methods which has been validated against analytical benchmarks as well as laboratory and field data (Nicol-sky, 2012; Nicolsky et al., 2011). Our main assumption is that we neglect the timing of the rupture propagation, assuming that the displacement of the seafloor occurs instantaneously (Nicol-sky et al., 2016), but we experiment with the rupture onset to account for unmodeled water propagation effects and match the observed tsunami waveforms. Given the presumed strike-slip nature of the main rupture we do, however, consider the contributions of both vertical displacements of the sea floor and horizontal displacements of bathymetry as contributors to the tsunami (Tanioka and Satake, 1996). The vertical and horizontal ocean floor displacements are calculated on a 0.0083 x 0.0083 degree grid using the formulas by Okada (1985). We find that in the strike-slip case resolved below, the contribution of the horizontal displacement to the modeled tsunami dynamics is relatively small, ~10% of the total tsunamigenic ocean floor displacements.

3.3 Coulomb Stress Change Modeling

To calculate the static stress changes caused by the Sand Point earthquake on the subduction slab interface, assuming the 2021 Chignik geometry as the receiver fault, we used the Coulomb 3 deformation and stress change software (Lin and Stein, 2004; Toda et al., 2011, 2005). We did so by assuming an elastic half-space with uniform isotropic elastic properties (Okada, 1992), and computing stress changes based on the finite-fault model developed for the Sand Point earthquake in this study. These stress changes were calculated for the receiver fault geometry (strike = 235.2°, dip = 15.1°, and rake = 95°), which is based on finite-fault models from Elliott et al. (2022) for the Chignik earthquake and is a flat approximation of the shallow portion of Slab2 in this region. The Chignik fault parameters are associated with the geometry of the subducting slab derived from Slab2 (Hayes et al., 2018), and therefore it is reasonable to interpolate the stress changes due to the Sand Point earthquake computed on a Chignik-like receiver fault onto the slab interface. To do so, we calculated stress changes at 1 km intervals for depths from 1 to 60 km, as this is the computational resolution limit along depth for the software used. At each slice, the stress changes are calculated with respect to the receiver fault geometry. Stress change values in each of these depth-slices were interpolated onto the model slab surface using the `scipy.spatial.KDTree` Python module, which uses an algorithm described by Maneewongvatana and Mount (2002). We tested two values for the coefficient of friction (0.1 and 0.4) to account for and assess the uncertainty in friction and its impact on the results (e.g., Elliott et al., 2022).

We also analyzed the stress-changes caused by the Sand Point fault model from this study on a normal fault dipping north (strike = 260°, dip = 35°, and rake = 225°), and a thrust fault dipping east (strike = 190°, dip = 30°, and rake = 90°). These faults were invoked in the Bai et al. (2023) model (they refer to them as F2 and F3). We calculated stress-changes for rigidity moduli 32 GPa as well as 3.2 GPa. Bai et al. (2023) attribute the fast slip on F3 to a lower rigidity modulus, and therefore we present results for that. We used a coefficient of friction $\mu = 0.4$ since a lower value did not cause a significant difference in values.

4 Results

4.1 Strike-Slip-only Finite Fault Model

Our inversion (temporal and spatial smoothing based on L-curves, see Supplementary Fig. S1) places the hypocenter slightly deeper than the USGS solution, at 34 km, which is similar to the Global CMT depth of 37.7 km (Ekström et al., 2012). Our spatio-temporal slip model resolves the bulk of the right-lateral strike-slip release about 20-30 km south of the hypocenter, between 15-45 km depth (Fig. 3a,b). From 30-50 km depth, within the downgoing slab, the slip ranges from about 1.5-2.0 m over an approximately 25 km long segment. In total, the bulk of the rupture was released on a ~90 km long fault segment between 10 and 50 km depth. The rupture

Model	USGS/GCMT	Jiang et al. (2022)	Our relocation (Fig. 2)
	350° strike, 49° dip	358° strike, 45° dip	1° strike, 58° dip
static GNSS rms	0.29 cm	0.31 cm	0.76 cm
static GNSS max. misfit	-0.60 cm	-0.82 cm	-3.71 cm
InSAR LOS offsets rms	0.34 cm	0.33 cm	0.34 cm
InSAR LOS offsets max. misfit	1.60 cm	1.64 cm	1.75 cm
teleseismic waveforms error in wavelets domain	0.412	0.442	0.403
high-rate GNSS error in wavelet domain	0.633	0.625	0.648
Teleseismic waveforms normalized variance reduction	66.4%	56.1%	55.0%
High-rate GNSS waveforms variance reduction	90.6%	89.5%	87.2%
High-rate GNSS horizontal components variance reduction	91.0%	90.6%	87.0%
High-rate GNSS vertical components variance reduction	89.3%	86.1%	87.8%

Table 1 Data misfit for different fault geometries

propagated south, initially slowly between 1-1.5 km/s at the hypocenter and eventually speeding up to 2.0-2.5 km/s (Fig. 3c). The moment was released in one narrow pulse from about 17-25 s after the origin time, when the high slip asperity in the center broke (Fig. 3a,d).

This finite fault slip model predicts all the data well (Fig. 4). The azimuth and magnitude of the co-seismic GNSS displacements are captured within the 95% uncertainties for horizontal and vertical offsets (Fig. 4a,b). While AC12 and AC28 are certainly the sites showing the largest offsets due to their proximity to the slip, it is notable that more distant sites such as AC21 and AB13 are also well predicted (Fig. 4a). The main characteristics of the high-rate GNSS waveforms are captured at almost all sites in both phase and amplitude (Fig. 4e). The notable exceptions are the vertical components at AC40 and AC42, both located ~150 km from the epicenter, which we assume is due to site effects as observed elsewhere (e.g., Parameswaran et al., 2023). Both the InSAR phase gradients as well as the line-of-sight (LOS) deformation are reproduced well by the model predictions (Fig. 4d). Misfits in the LOS are within about +/- 1-1.5 cm and largest on Big and Little Koniuji Islands, mostly because of small differences in phase gradients which are perhaps best attributed to slight differences in strike and dip of the model fault. The teleseismic waveforms (Fig. 4c) are also generally well reproduced in phase and amplitude, albeit the model is not reproducing some of the higher frequency signals, perhaps due to site effects. Amplitudes at some sites with azimuths perpendicular to the fault strike (e.g., sites in the middle column in Fig. 4c) are underpredicted, likely due to topographic effects as these phases might be affected by the steep forearc slope (Wiens, 1989).

The tide-free tsunami records for DART buoys 46402, 46409, and 46414 (Fig. 1) are shown as red traces in Fig. 5. We can clearly see the arrival of the seismic waves followed by small tsunami waves of ~2 cm at DART 46402 and 46414, and ~1 cm at DART 46409. The tsunami that our preferred, 49 degree dipping strike-slip model (see Fig. 3a for rake) would induce does not reproduce these observations well (blue trace). While the positive

phase of the modeled tsunami roughly matches in amplitude, it arrives 5 minutes too early. Following this, the modeled tsunami lacks the negative phases observed at DARTs 46402 and 46414.

4.2 Strike-Slip and Slab Interface Finite Fault Model

To test whether the tsunami can be explained with megathrust slip, we set up a combined strike-slip and slab interface model, in which the rupture was assumed to initially propagate on a strike-slip fault and trigger a rupture of the slab interface from the relocated main-shock hypocenter's projection on the slab interface with several seconds delay. The strike-slip fault has the same geometry as previously used. The geometry of the slab interface was 239° strike and 16.5° dip, as specified by the Slab2 model (Hayes et al., 2018). Both fault planes cross at 54.91°N, 159.77°W at 34 km depth. To determine the time delay for the rupture on the slab interface, we tested some candidates ranging between 2 and 15 s in 0.5 s intervals and found 9.5 s as the preferred delay, based on minimized waveform misfits (Fig. S2).

The resulting model (Fig. 6) resolves the bulk of the right-lateral strike-slip release on the strike fault between 15 and 40 km depth and some notable thrust slip on the slab interface at shallower depths (12 - 20 km depth). The maximum slip on the strike-slip fault and the slab interface are ~1.7 m and ~0.8 m, respectively. Notably, the maximum slip on the strike-slip fault decreased and was substituted with slip on the megathrust in the same area. The released seismic moment is $\sim 2.9 \times 10^{20}$ Nm, corresponding to M_w7.6. Of this, 79.5% of the seismic moment ($\sim M_w 7.5$) are released on the strike-slip fault, and 20.5% of the seismic moment ($\sim M_w 7.1$) are released on the thrust fault.

This model fits the static and high-rate GNSS data as well as the strike-slip model (Fig. 7a,b,e) with only subtle differences (both improvements and deteriorations) in the later parts of the high-rate GNSS waveforms. The normalized variance reduction of teleseismic waveforms improves slightly from 66.4% in the strike-slip model to 74.4%, especially with the amplitudes in the

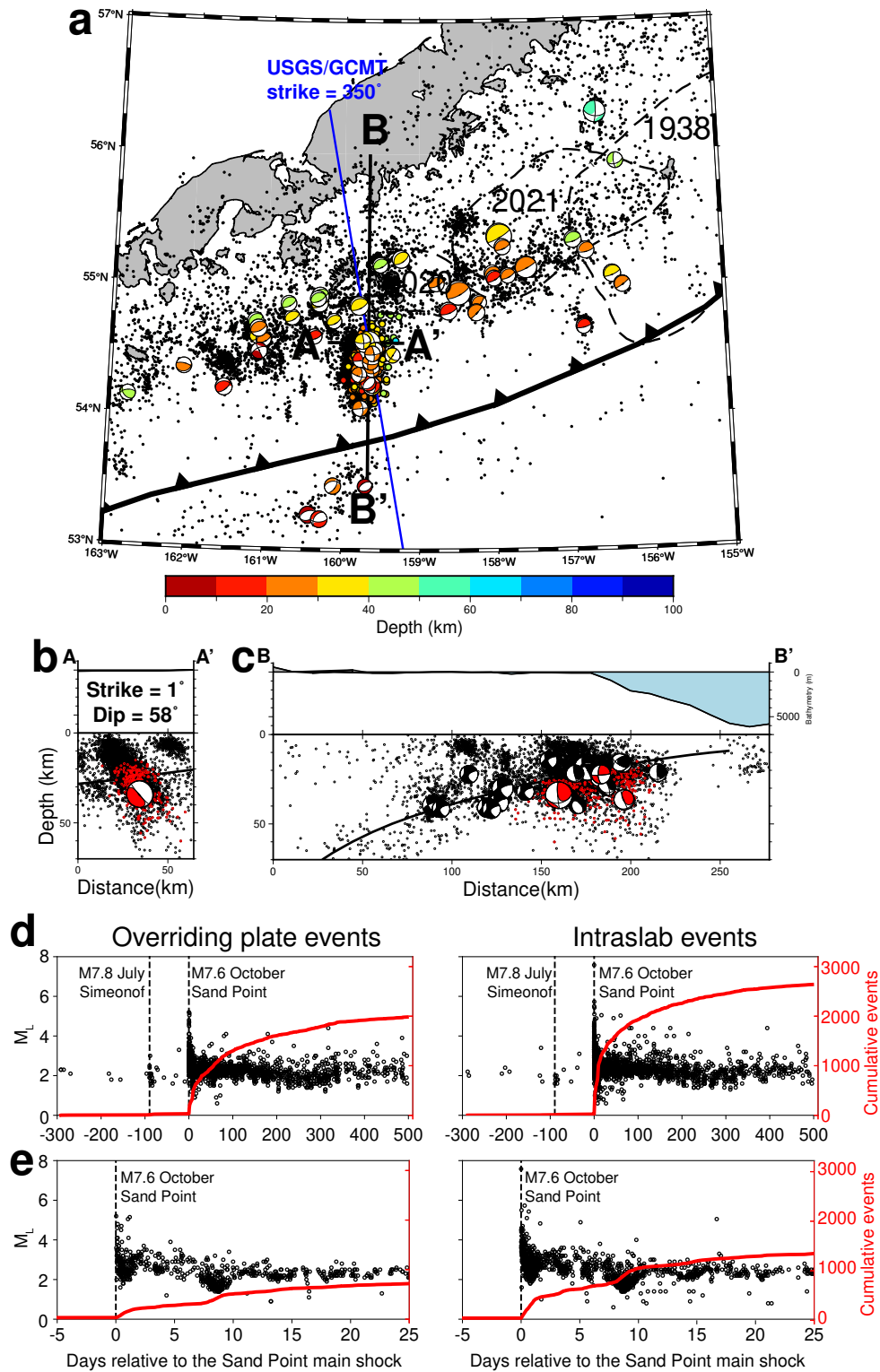
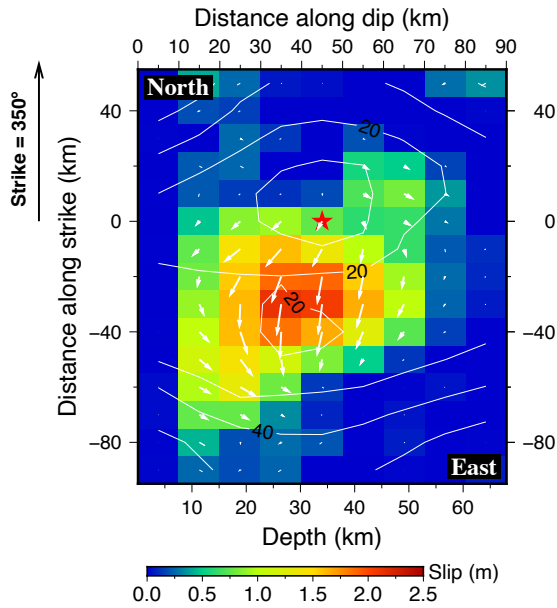
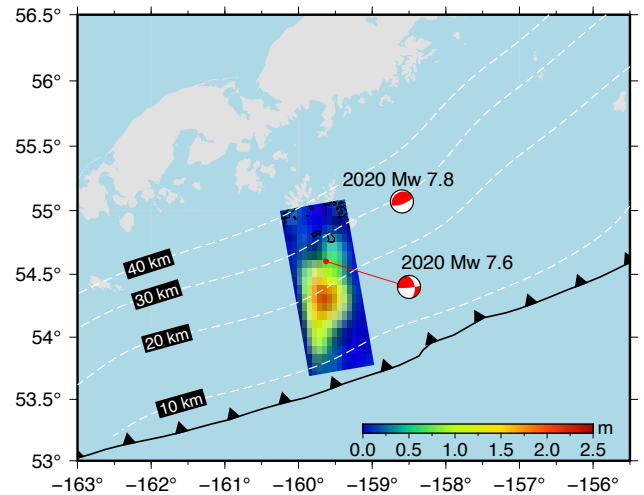


Figure 2 Map and cross-sections of relocated earthquakes from 1 Jan. 2020 to 28 Feb. 2022. **(a)** Map view. Dashed contours outline the rupture areas of the 1938 M_w 8.3 Alaska Peninsula earthquake (Freymueller et al., 2021), 2020 M_w 7.8 Simeonof earthquake (Xiao et al., 2021), and 2021 M_w 8.2 Chignik earthquake (Elliott et al., 2022). Earthquakes 24 hours after the 2020 Sand Point mainshock are highlighted as larger dots. Labeled black straight lines indicate the cross-sections in other panels. Blue straight line shows the fault strike of the 2020 Sand Point earthquake determined by USGS and Global CMT (Ekström et al., 2012). Cross-section B-B' is parallel to the strike determined by 1-day aftershocks and A-A' is perpendicular to B-B'. **(b-c)** Cross-sections. Earthquakes 24 hours after the 2020 Sand Point mainshock are highlighted in red. Black curve indicates the Slab2 model (Hayes et al., 2018). Focal mechanisms plotted as back-hemisphere projection. The bathymetry is plotted on top with vertical exaggeration. **(d)** Temporal distribution of 2020 Sand Point aftershocks in the overriding plate (2 km above the plate interface) and in the slab (2 km below the plate interface). **(e)** Zoom-in of (d) 5 days before and 25 days after the Sand Point mainshock.

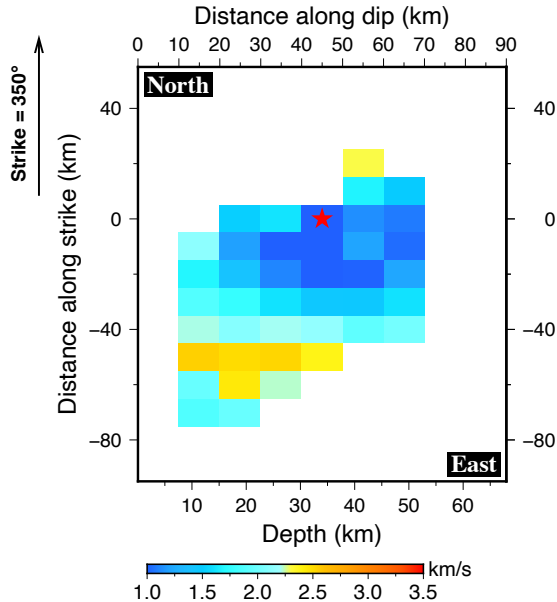
(a) Slip distribution



(b) Slip mapview



(c) Rupture velocity



(d) Moment rate function

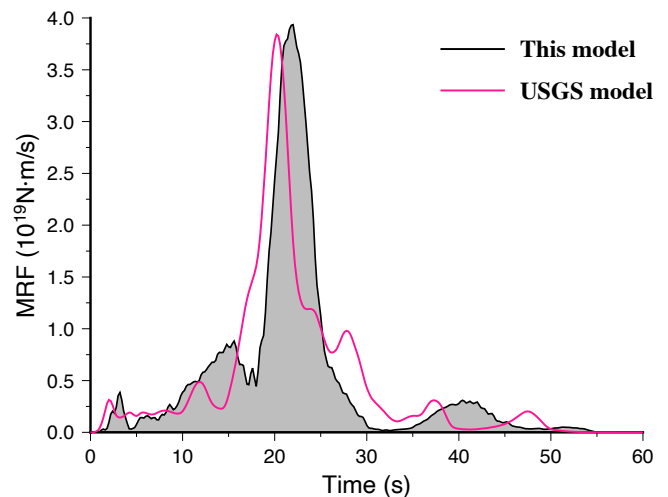


Figure 3 Strike-slip-only finite fault model. **(a)** Fault normal view (depth on horizontal axis) of preferred slip distribution and time propagation contours achieved in the strike-slip-only finite fault model, **(b)** map view of slip distribution, dip of fault plane is 49 degrees, slab depth contours shown for reference only; slip is not on the megathrust, **(c)** Fault normal view of rupture velocity distribution, **(d)** Resolved moment rate function (shaded area) and comparison to USGS moment rate function (magenta curve).

second column of Fig. 7c being better fit than for the strike-slip model (Fig. 4c). A notable difference, however, is in the InSAR misfit (Fig. 7d), which is significantly worse in the line-of-sight panel compared to the strike-slip model, especially on islands (such as the Koniuji islands) that are not also constrained by static GNSS offsets (Fig. 4d).

Fig. 8 shows the tsunami this model (blue trace) would generate and the wave height observations (red trace). While the combined model creates a slightly larger tsunami than the single-fault model (Fig. 5), it arrives even earlier, and still does not match the negative phases in the DART 46402 record, but begins to ex-

press the negative phase observed at DART 46414. The black line shows that much of the characteristics of the tsunami are already produced by the strike-slip component of this two-fault model.

4.3 Submarine landslide as a tsunami source

To investigate other potential tsunami sources, we assessed the possibility of a submarine landslide generating the tsunami. Vertical seismograms for stations within 600 km of the 2020 M_w 7.6 Sand Point epicenter (Fig. S3), filtered at 3.3 - 20 mHz (50 - 300 s period), are shown in blue in Fig. S4. We compare this

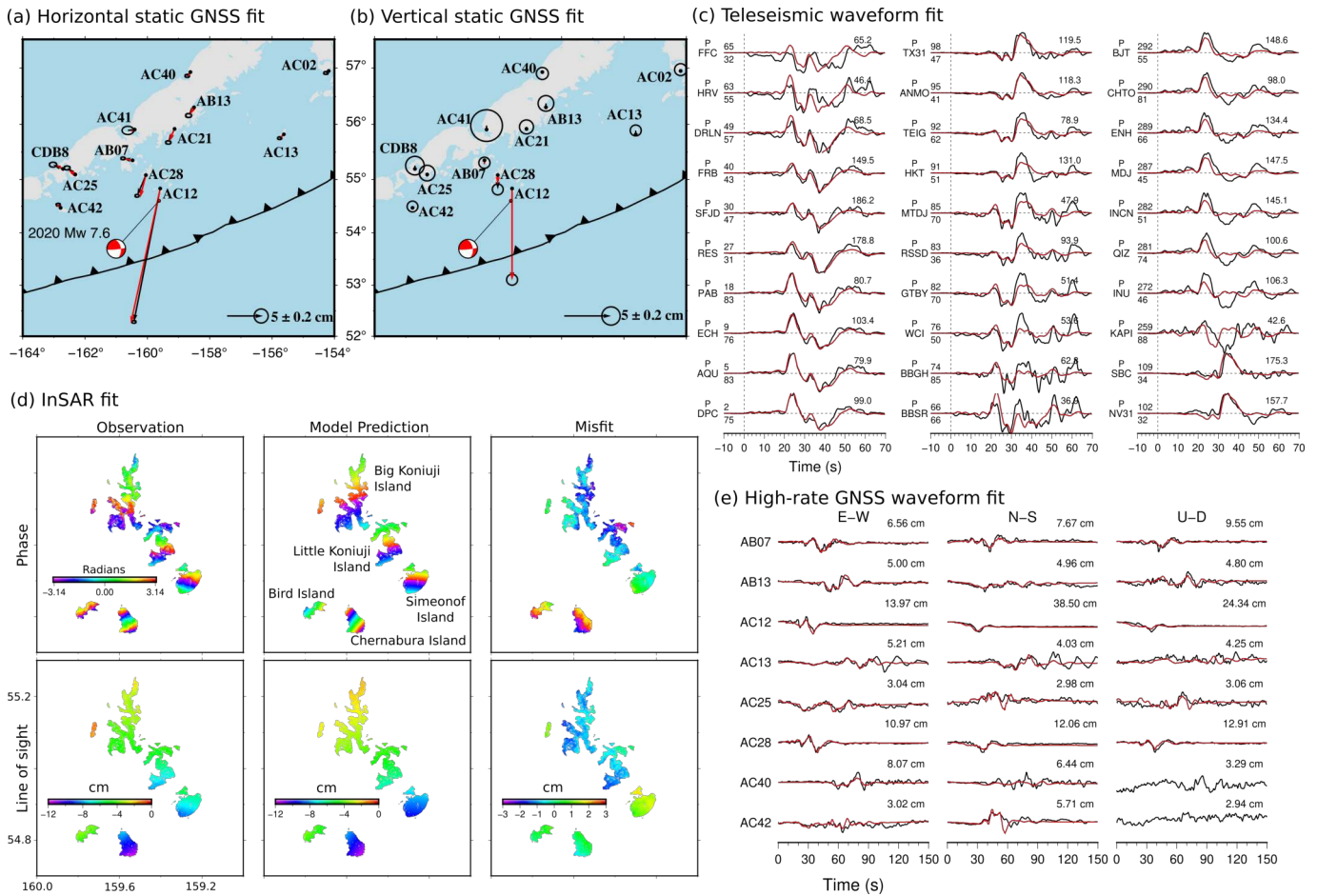


Figure 4 Fit of strike-slip-only finite fault model predictions to data that constrain the inversion. **(a)** horizontal and **(b)** vertical static co-seismic GNSS displacement observations (black) and respective model predictions (red); **(c)** subset of teleseismic waveform observations (black) and predictions (red); the annotations of each record are at the start: the type of seismic wave (top left), station name (bottom left), azimuth (top right) and epicenter distance (bottom right, in degrees) for each seismograph and at the end: the peak displacement in micrometers; **(d)** top: InSAR phase observations, phase prediction, and difference, bottom: Line of sight displacement (LOS) observation, prediction, and difference; **(e)** high-rate GNSS time series observations (black) and model predictions (red), no model predictions for AC40, AC42 vertical, because they were not used in the inversion.

to a long-period synthetic assuming the velocity structure of the Preliminary Reference Earth Model (PREM) (Dziewonski and Anderson, 1981) and using the GCMT moment tensor as input. This shows that observations at most stations are well reproduced by the model data. Small phase shifts can be explained by the finiteness of the rupture propagation and other source details that are not considered here. For some of the station observations with mismatching synthetics the sensor was affected by non-linear effects due to strong shaking, as evident in the raw seismograms for AK.CHN on Chernabura and AT.SDPT at Sand Point.

The dotted lines in Fig. S4 are the same model predictions as in the red line, but shifted by a 5 minute delay. We introduce this to demonstrate that the long period data (blue lines) show no appreciable signal at this point, which is when we would expect a submarine landslide signal to manifest in the records if the Sand Point tsunami was caused by such an event (Ekström et al., 2003), thus ruling out this hypothesis.

4.4 Shallow, focused megathrust slip as a tsunami source

Having ruled out megathrust slip that would be sensed by land-based observations in the finite fault inversion (Section 4.2) and the submarine landslide (Section 4.3) as potential tsunami sources, we further investigated where, when, and how much slip would need to be released on the megathrust to create the observed tsunami. This requires low amplitude or very concentrated shallow slip, undetected by the land-based geodetic observations, and slow enough to not excite seismic waves. We conducted an experiment, in which we distributed unit sources of pure thrust-slip along the megathrust interface on a 150 km (along-strike) x 180 km (along-dip) grid, with a spacing of 30 km along strike and 10 km along dip (Freymueller et al., 2021; Nicolsky et al., 2016). Each unit source is made of nine 10x10 km rectangular subfaults arranged in 3 rows and 3 columns. The central subfault has unit slip (1 m); subfaults sharing an edge have half of the unit slip and corner neighbors have a quarter of the unit slip to achieve some tapering. We used these to predict the vertical sea floor

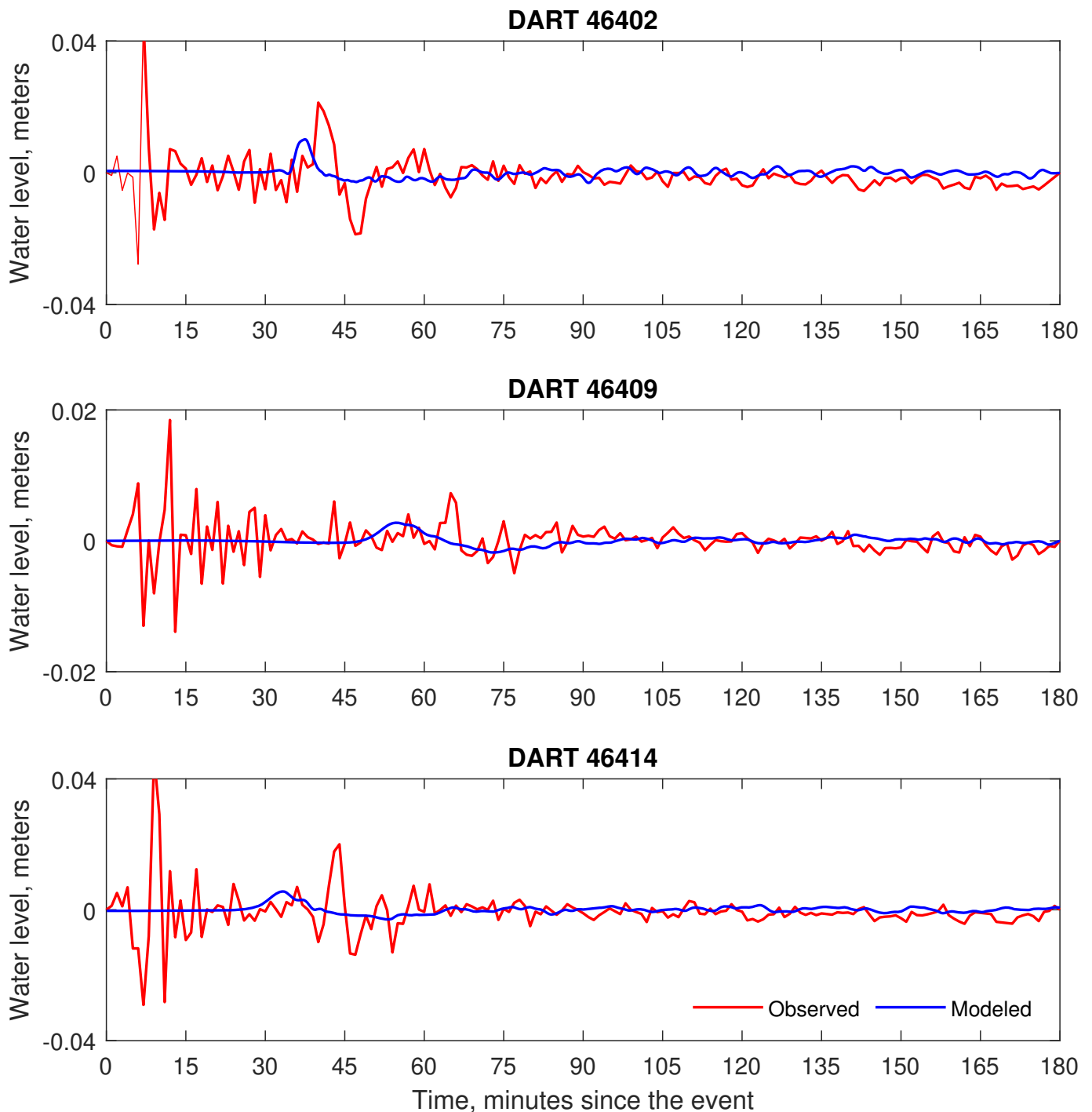


Figure 5 Tide-removed DART tsunami observations (red) at DART buoys 46402 (top), 46409 (middle), and 46414 (bottom). The tsunami model (blue line) driven by the strike-slip-only model sea floor displacements predicts the arrival about 5 minutes before observation and lacks a distinct negative phase at DART buoys 46402, 46414, and cannot reproduce the second phases for any buoy.

displacements using elastic Green's functions (Okada, 1985). Given the linear relationship between slip amplitude and surface displacement, we scale the vertical displacement to generate a tsunami of appropriate magnitude.

Because it is clear from the tsunami models above (Figs. 5, 8) that the timing of the modeled tsunami might not agree with the observed tsunami, we also experimented with delaying the strike-slip-only (0 s, 30 s, 60 s) and unit source sea floor deformation (300 s, 360 s, 420 s, 480 s, 540 s) relative to the earthquake origin time. This

is justified, because the tsunami model assumes instantaneous sea floor deformation, but a large earthquake rupture takes finite time, and the surface displacements will occur when the S waves reach the ocean bottom (Grapenthin and Freymueller, 2011). The moment rate functions in Figs. 3 and 6, for instance, suggest about 50 s for this earthquake.

Figs. S5-S17 show a suite of candidate megathrust unit source models that create well-fitting tsunami, their necessary time delays, and respective fits to DART buoy observations. All of them place slip in the shallow

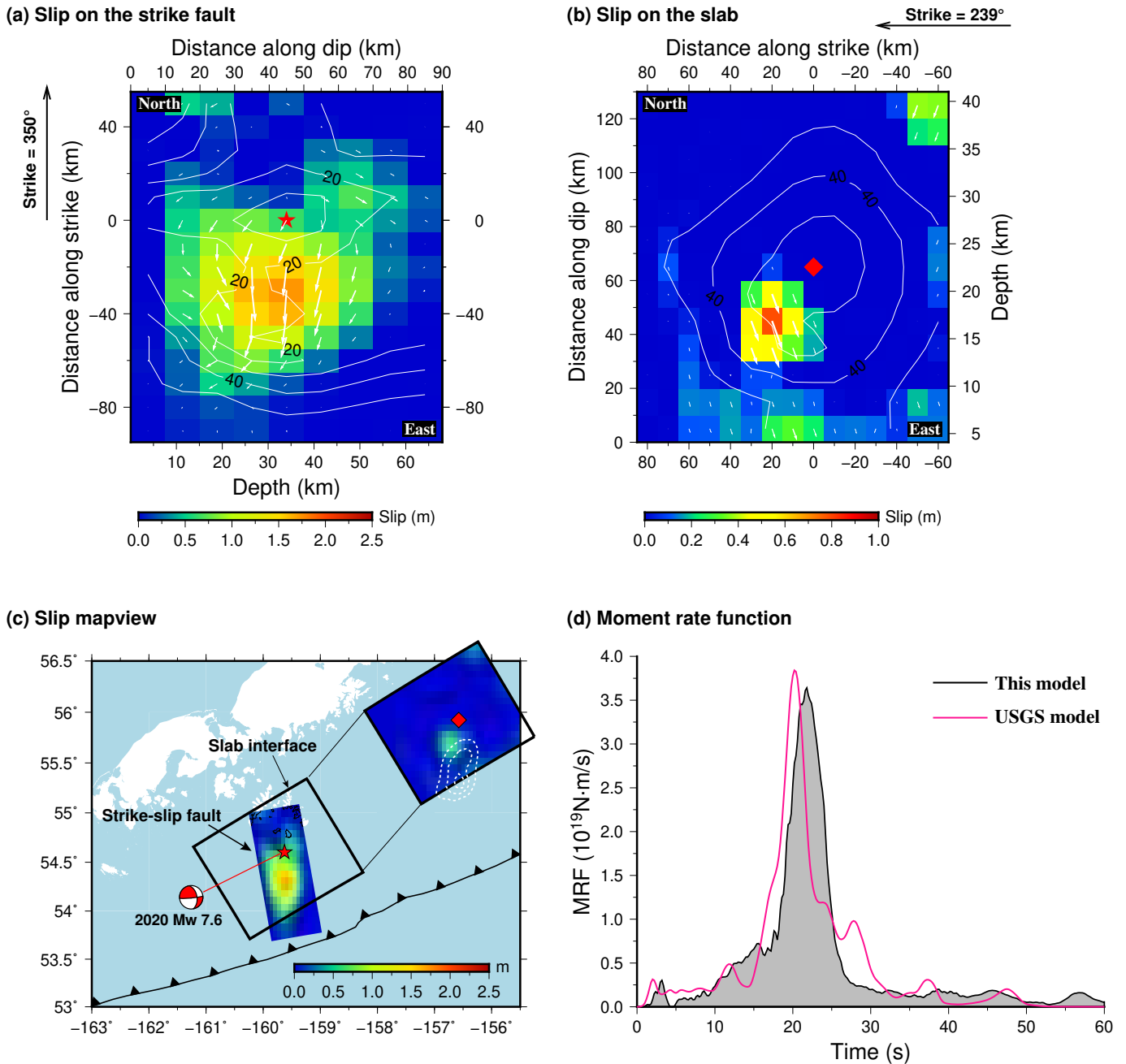


Figure 6 Strike-slip and slab interface finite fault model (note different color scales). **(a)** Fault normal view of preferred slip distribution and time propagation contours for the strike-slip fault component of the model, strike direction of 350° indicated by arrow on the left, red star indicates the relocated hypocenter on the strike-slip fault. **(b)** Same as (a) but for the slab component of the model, strike direction is indicated as 239°; red diamond indicates the relocated hypocenter projection onto the slab interface, **(c)** Map view of the slip distribution of the 2 fault model. The location of the slab interface slip, and symbols relative to this (red diamond, white contours) are shown in top right inset (actual model location indicated by thick black rectangle outline) to ensure the strike-slip fault model component is visible, red star and red diamond are the relocated hypocenter locations as in (a) and (b), respectively. The white contours with a 0.5 m contour interval indicate the unit slip source locations estimated from fitting the tsunami recorded by DART buoys (see below). Note that this extends to shallower depths than the slab interface slip model. **(d)** Resolved moment rate function of this model (shaded area) and comparison to USGS moment rate function (magenta curve).

western model domain. Based merely on the fit of predicted and observed tsunami, and requiring relatively low maximum slip, we find that a model (Fig. S5, Fig. 9) with slip of up to ~1.7 m on the megathrust close to the trench (~2.5–10 km depth), updip of our two-fault inversion estimation (Fig. 6c), could generate tsunami waves that reasonably approximate the observations (Fig. 10c–

e).

4.5 Stress Change

To better understand what stress changes the strike-slip-only finite fault slip induced on the megathrust and whether there is a relation to the proposed tsunami source, we model the Coulomb stress change (Fig. 9).

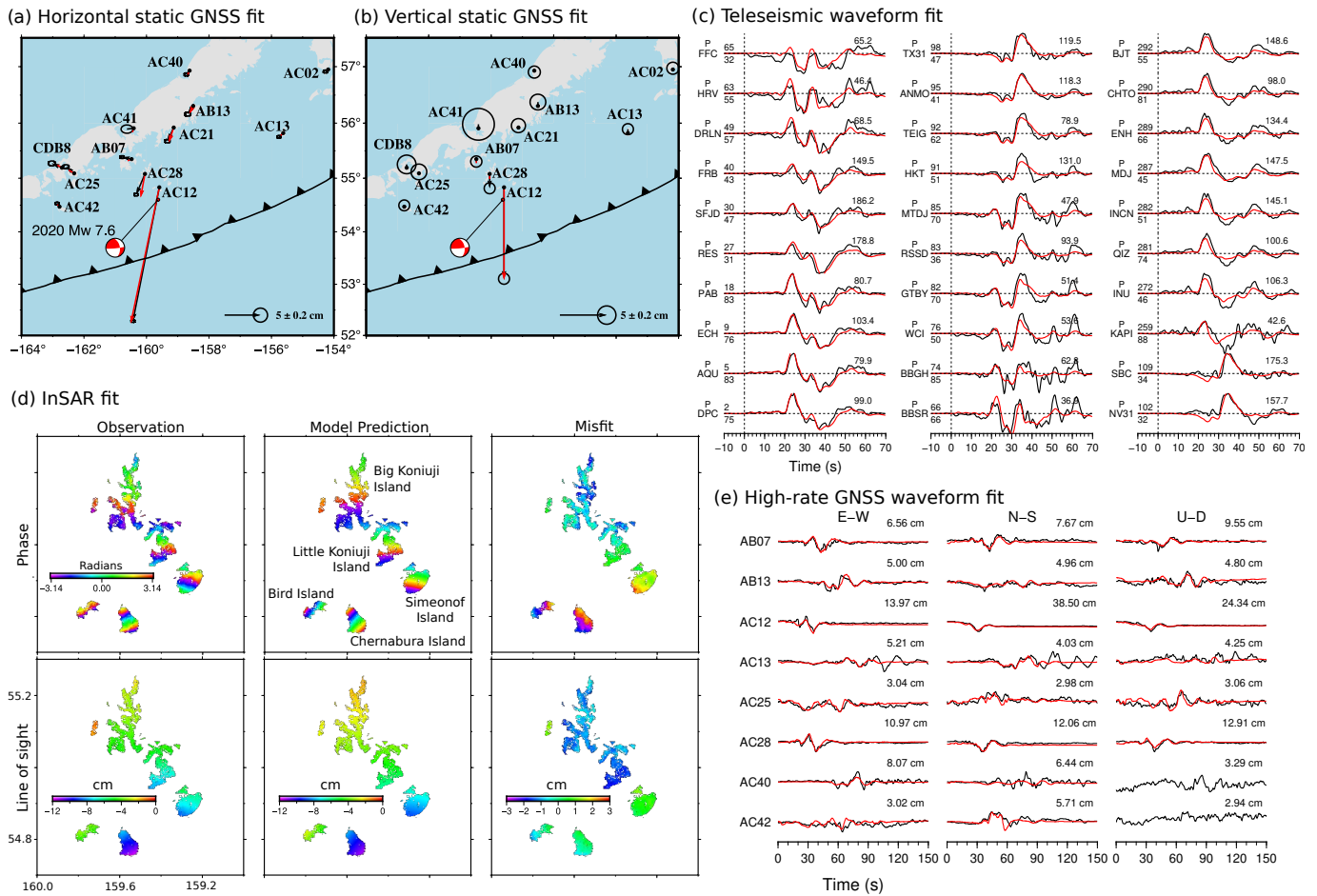


Figure 7 Similar to Figure 4, but for the combined strike-slip fault and slab interface model. The differences between both models are subtle and most obvious in increased LOS misfit, while the teleseismic waveform fit improves. The GNSS waveforms degrade subtly, but overall misfit measures and static offset fits remain the same.

Stress changes due to a Sand Point strike-slip-only model on the slab interface with Chignik receiver fault geometry (see Elliott et al., 2022) have the hypothetical megathrust slip maximum in the up-dip end of the Coulomb stress increase (Fig. 9c). While the preferred megathrust slip model was chosen without considering stress change, most of our adequate tsunami models (Figs. S5-S17) place the slip in this general region; some more focused with higher slip amplitudes (e.g., Fig. S12) necessary to minimize GNSS misfit.

The modeled stress changes for the lower coefficients of friction ($\mu=0.1$) are as high as ~ 36 bar (3.6 MPa) and stress-shadows as low as ~ 21 bar (-2.1 MPa). For this frictional value, we note that the Sand Point earthquake caused only small stress changes of ~ 0.01 bar (-0.001 MPa) near the source region of the 2020 Simeonof event, and ~ 0.004 bar (0.0004 MPa) near the source of the 2021 Chignik event. For $\mu = 0.4$, the maxima of positive and negative stress changes due to the earthquake are ~ 49 bar (4.9 MPa) and ~ 30 bar (-3 MPa), respectively. In this case, the stress changes observed at the Simeonof and the Chignik earthquake sources were ~ 0.04 bar (-0.004 MPa) and ~ 0.005 bar (0.0005 MPa), respectively. The contribution of static stress change from the Sand Point earthquake to the Chignik earthquake source remains, thus, minimal compared to those that the Simeonof earthquake induced at the Chignik location as pre-

viously suggested by Elliott et al. (2022). Meanwhile, the Simeonof source region, which already slipped in the 2020 M_w 7.8 event, seems to have been pushed further into a stress-shadow by the Sand Point earthquake. Notably, two M_w 5.7 aftershocks of the Sand Point event are located at the transition from stress increase to stress drop on the megathrust (Fig. 9), showing that these were caused by the strike-slip mainshock.

5 Discussion

5.1 Finite Fault Models

While our strike-slip-only model reproduces the geodetic and seismic observations well, it cannot generate the first-order features of the tsunami observed at the DART buoys, particularly the negative initial phase of the tsunami or its later phases. This deficiency is not unique to our model, but has been a feature of nearly all previous rupture models, resulting in inverse models that suggest slab interface slip in addition to strike-slip (Santallanes et al., 2025), or a complex combination of intraplate strike-slip faulting and enigmatic slow thrust slip (Bai et al., 2023). While our earthquake relocations show many aftershocks in the overriding plate, and strike-slip faults intersecting both subducting and overriding plate have been documented in, for example,

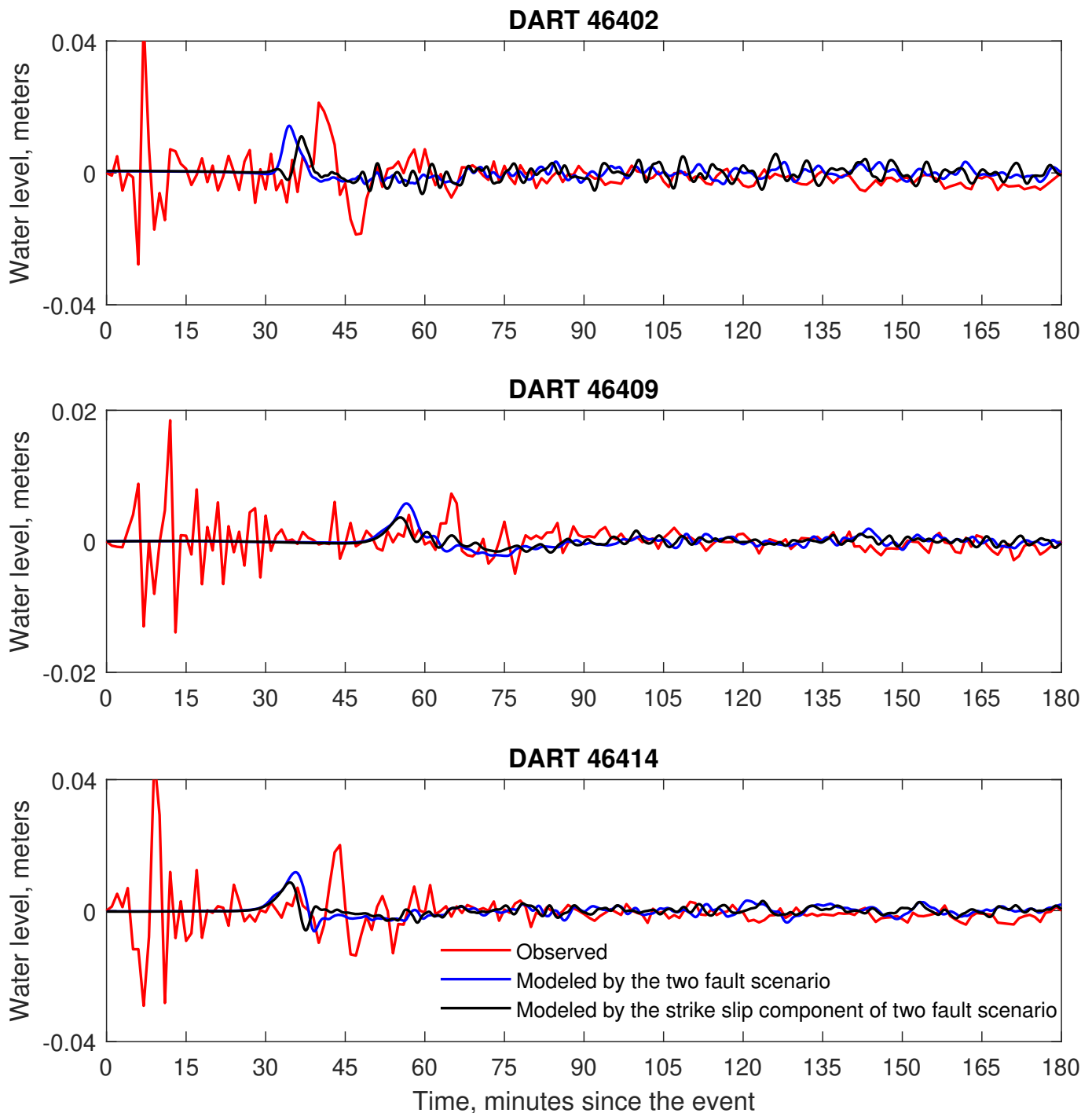


Figure 8 Like Figure 5, but showing the fit of the two-fault model (strike-slip and slab interface) tsunami prediction to the tide-removed observations (red) at DART buoys 46402 (top), 46409 (middle), and 46414 (bottom). The blue line considers vertical ocean floor displacements for both faults, the black line considers the vertical ocean floor displacements from the strike-slip fault of the two fault model only. Both arrive about 5-10 minutes early and neither predicts the initial negative phase or the amplitude of the second phase of the tsunami.

Cascadia (Applegate et al., 1992; Goldfinger et al., 1992), the existing geodetic data could not uniquely resolve whether this region slipped co-seismically, nor would slip on an upward continuation of the modeled strike-slip fault affect the timing of the tsunami and its later phases. Hence, we tested whether a combination of strike-slip and megathrust slip would explain the geodetic and seismic data and reproduce the tsunami.

The slip distribution on a two-fault model that inverts for strike-slip and megathrust slip achieves an ac-

ceptable fit to the geodetic and seismic data. Notably, the slip assigned to the megathrust aligns roughly with the region of slip reduction in the main slip area of the strike-slip fault when compared to the single fault model (compare Figs. 3 and 6). Some aspects of the data are fit slightly better with the two-fault model, for instance the later parts of the teleseismic waveforms, while the fit to the GNSS data remains mostly the same. This highlights classic non-uniqueness issues of inverse problems. The fit to the InSAR data, however, is notably

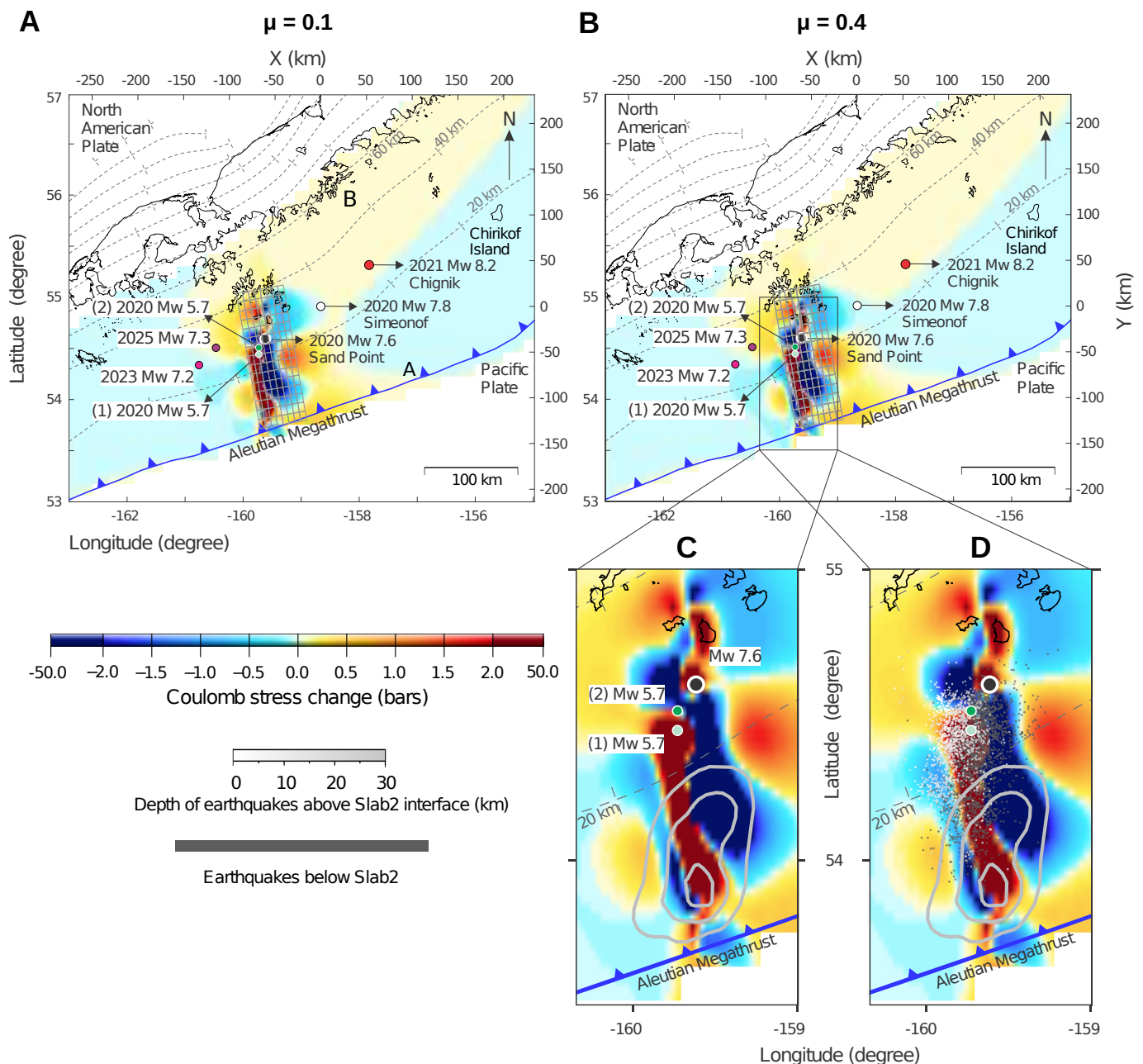


Figure 9 Coulomb stress changes caused by the 2020 Sand Point earthquake on the subduction slab interface, assuming a receiver fault geometry similar to the 2021 Chignik earthquake (see Elliott et al., 2022). The gray grid represents the strike-slip-only finite fault model from this study. The 2020 M_w 7.6 Sand Point, the 2020 M_w 7.8 Simeonof, the 2021 M_w 8.2 Chignik, the 2023 M_w 7.2, and the 2025 M_w 7.3 earthquakes are labeled and indicated by the black, white, red, magenta, and purple circles respectively. The two 2020 M_w 5.7 (labeled as (1) and (2)) aftershocks are indicated by circles in two shades of green. The first M_w 5.7 aftershock (light green, label (1)) occurred ~ 11.3 minutes after the Sand Point mainshock, and was soon followed by the second M_w 5.7 aftershock (dark green, label (2)) just ~ 23 seconds later. Stress changes on the slab interface assuming coefficient of friction, (a) $\mu = 0.1$, and (b) $\mu = 0.4$; rectangle shows location of panels (c), (d). (c) Detail plot of Coulomb stress change on subduction slab interface, showing the two M_w 5.7 aftershocks approximately in the region of largest stress gradient due to the M_w 7.6 mainshock. The gray 0.5 m contours show the unit slip source locations, independently estimated from fitting the tsunami recorded by DART buoys as shown in Figs. 6 and 10. (d) same as (c), but with the grid removed and smaller relocated aftershocks from Fig. 2 plotted. Aftershocks above the slab are plotted in light gray gradients down to 30 km depth; earthquakes below the Slab2 interface are plotted in dark gray. See Fig. 2c for cross section.

worse, especially on the islands that are not also constrained by static GNSS offsets. This could perhaps be improved by further optimizing the relative weights of data sources in the inversion. We find, however, that the InSAR data (not used in previous studies) provides observations in regions that are sensitive to megathrust

slip and thus help resolve some of the model non-uniqueness. This leads us to reject the hypothesis that co-seismic slab interface slip in addition to slip on the main strike-slip fault created the tsunami. The modeled tsunami for the two-fault model (Fig. 8) further supports this as it cannot match the observations in either timing

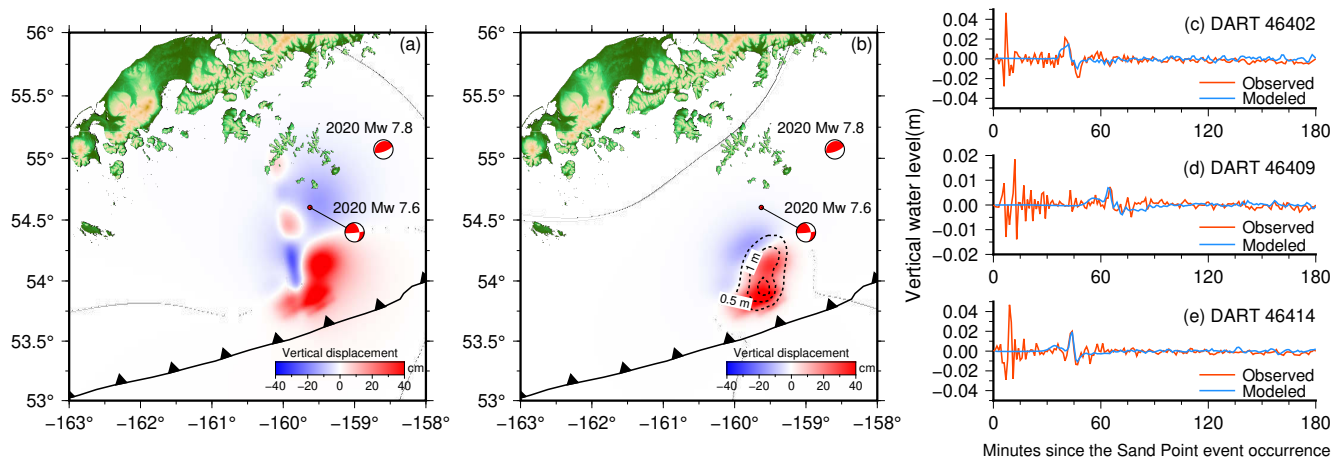


Figure 10 (a) Vertical displacement field resulting from the combined contribution of strike-slip and unit megathrust slip sources on the slab, derived through inversion of tsunami observations recorded by DART buoys, black line shows trench location. (b) Vertical displacement field created from just the unit sources of megathrust slip on the slab. The black contours indicate locations of the megathrust slip. (c-e) Comparison of observed and predicted tsunami data at three DART stations for a tsunami model that assumes a delay of 60 s in the strike-slip rupture and 480 s for the megathrust component of the rupture compared to the earthquake origin time ($D_{\text{strike}} = 60$ s; $D_{\text{slab}} = 480$ s model in Supplement).

or shape.

5.2 Tsunami Source

The sea floor deformation due to our inferred strike-slip-only and unit source model components has to be delayed by 60 s and 480 s, respectively (Fig. 10a,b). Some of this shift could be due to incomplete physics of the shallow water equation, which has been observed in numerical simulations (Nicolisky et al., 2016; Watada et al., 2014). Fig. S5 (bottom) separates the contributions of the strike-slip and megathrust models to the synthetic tsunami, illustrating that the bulk of the high amplitude phases of the tsunami depends on the delayed megathrust component, while the strike-slip component drives earlier, lower amplitude elements. The seismic moment released by such a slip scenario is about 3.8×10^{20} Nm, equivalent to $M_w 7.6$. Since the water column likely exceeds 1 km depth at the candidate location, waves would quickly propagate away from the source, limiting how slow the slip can be released before noticeable wave diffusion. Dynamic slip-release experiments (Figs. S18-S21) show that a rise time of about 200 seconds, between 400-600 seconds after the mainshock, produces an acceptable tsunami fit if we quadruple the maximum slip. Compared to previously proposed megathrust slip (Santallanes et al., 2025), all our candidate sources are closer to the trench and thus much shallower (2.5-10 km compared to >20 km), more compact (10-20 km compared to >50 km width), and have similar maximum slip amplitude (1.7-6.8 m compared to ~4.5 m). Notably, the Santallanes et al. (2025) megathrust and strike-slip model components predict deformation not observed in the InSAR data (Fig. S22).

Given such a large moment release on the megathrust that had to be shifted by about 6.5-8 minutes, depending on the rupture velocity, seismic observations should capture such energy release as a discrete event. How-

ever, we could find no such earthquake (Fig. S23). Similar to a landslide (see text above), very slow slip events should generate long-period energy (Ekström et al., 2003), which also is absent from the seismic records (Figs. S4, S23). When we compare (Supplementary Text S2) observed long-period seismicity to predictions for our strike-slip and tsunami sources, the model predicts a second Rayleigh wave that is not observed when assuming fast rupture (Fig. S24). However, assuming slow rupture between 50-150 s, common to many tsunami earthquakes (Bilek and Lay, 2002), the unobserved phase predictions disappear (Fig. S25). Lastly, while the hypothetical slip scenario could cause unobserved deformation at the nearest site AC12 with a maximum offset of 3 cm, concentrating the slip in the updip region would eliminate this.

5.3 Tsunami Source and Mainshock Stress Change

Given the geometric assumptions of the unit source modeling, the match between proposed megathrust slip and stress change is not perfect and some slip falls into the stress shadow. These regions are also closer to GNSS site AC12. Since it is unlikely that a delayed rupture would occur here, a more concentrated rupture of the shallow, high-stress change megathrust seems very plausible. This would have to be of higher slip than modeled in our scenario to ensure the necessary seafloor motion due to rupture of a smaller asperity, and is substantiated by a lack of aftershocks in that updip region (Fig. 9d). While we cannot provide a unique answer for the full spatial extent of this slow megathrust rupture, two $M_w 5.7$ aftershocks occurred 11 min 19 s and 11 min 42 s after the mainshock at depths of 22.7 km and 26.2 km, respectively, approximately the slab depth in this area (Hayes et al., 2018, Fig. 9). While speculative, it seems plausible that these might backstop the slip on

the megathrust that unfolded in the minutes following the 2020 Sand Point rupture and represent aftershocks of this very long-period event. Because of its simplicity, this model is a preferable alternative to other models that have previously been put forward, despite its non-uniqueness and some residual waveform complexities. For instance, if we model the stress change due to our single-fault model, we cannot exclude the suggested contribution (Herman and Furlong, 2021) of a normal fault in the sediment wedge (Bécel et al., 2017), but aftershocks in the overriding plate did not begin until 7 days after the event (Jie et al., 2026), and it is clear that the westward-dipping tsunamigenic thrust fault from Bai et al. (2023) falls in a stress shadow (Fig. S26), and is thus less likely to rupture here.

6 Conclusions

We model the most complete seismo-geodetic dataset so far for the 19 October 2020 M_w7.6 Sand Point, Alaska, earthquake, composed of high-rate and static GNSS displacements, InSAR line-of-sight displacements, and teleseismic waveforms. Our preferred model consists of a single strike-slip fault, striking ~350 degrees and dipping 49 degrees, with a maximum slip of 2.1 m and dominant slip released between 22–38 km depth. This puts the fault below the megathrust into the downgoing Pacific Plate. This rupture triggered slow release of strain on the shallow parts of the megathrust in the minutes following the mainshock in a narrow region of Coulomb stress increase, potentially at very long periods and not geodetically detected, which resulted in the uncharacteristically large tsunami for a strike-slip earthquake.

The slowness of slip and distance offshore means that it remains hidden from land-based geophysical observations. This suggests that substantial, spatially concentrated tsunamigenic megathrust slip may occur co-seismically or delayed at very long periods, undetectable with land-based observations, creating small but in our case significant tsunami. It is conceivable that such unresolved strain release is entirely missed when analyzing large megathrust earthquakes and their tsunami, perhaps requiring repeat bathymetry or other direct seafloor observations to become adequately resolved. If common, such slip affects the seismic potential of the most tsunamigenic part of the megathrust, changing the impact of future earthquakes and the tsunami hazards of a region.

Acknowledgments

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Data and code availability

This study uses seismic data from the following networks: (Albuquerque Seismological Laboratory (ASL)/USGS, 2006, 1992; Albuquerque Seismological Laboratory/USGS, 1988; GEOFON Data Centre, 1993; MedNet Project Partner Institutions, 1990; Natural Resources Canada, 1975; California Institute of Technology and United States Geological Survey Pasadena, 1926; Charles University in Prague et al., 1973; Institut de physique du globe de Paris (IPGP) and École et Observatoire des sciences de la Terre de Strasbourg (EOST), 1982; EarthScope Consortium, 1986) and data from the following GNSS sites: AB07 (UNAVCO Community, 2004), AB13 (UNAVCO Community, 2006a), AC02 (UNAVCO Community, 2006b), AC12 (UNAVCO Community, 2008a), AC13 (UNAVCO Community, 2008b), AC21 (UNAVCO Community, 2006c), AC25 (UNAVCO Community, 2005), AC28 (UNAVCO Community, 2008c), AC40 (UNAVCO Community, 2007a), AC42 (UNAVCO Community, 2007b), CDB8 (<https://geodesy.noaa.gov/corsdata>).

This study uses GIPSY/OASIS goa-6.4 (Bertiger et al., 2010; Zumberge et al., 1997) licensed to JF and GipsyX (Bertiger et al., 2020) licensed to RG for the GNSS analysis, available via JPL. Some figures are made with GMT (Wessel et al., 2019) or matplotlib (Hunter, 2007). Earthquake relocations are done with tomoDD, available through its authors (Zhang and Thurber, 2003). The slip inversions are accomplished with a version of code that originated from Chen Ji and was modified by Z. Xiao (Ji et al., 2003; Shao et al., 2011; Xiao et al., 2021), available through its authors. We use open source GMTSAR (Sandwell et al., 2011) for the InSAR processing, which is available at <https://github.com/gmtsar/gmtsar/wiki>. Matlab scripts and data for the tsunami modeling are available through Zenodo with doi 10.5281/zenodo.13395583.

Competing interests

The authors have no competing interests.

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