Subduction earthquakes controlled by incoming plate geometry: The 2020 M>7.5 Shumagin, Alaska, earthquake doublet

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Abstract

In 2020, an earthquake doublet, a M7.8 on July 22nd and a M7.6 on October 19th, struck the Alaska-Aleutian subduction zone beneath the Shumagin Islands. This is the first documented earthquake doublet, of considerable size, involving a megathrust event and a strike-slip event, with both events producing deeply buried ruptures. The first event partially ruptured a seismic gap, which has not hosted large earthquakes since 1917, and the second event was unusual as it broke a trench-perpendicular fault within the incoming oceanic slab. We used an improved Bayesian geodetic inversion method to estimate the fault slip distributions of the major earthquakes using Interferometric Synthetic Aperture Radar (InSAR) wrapped phase and Global Navigation Satellite Systems (GNSS) offsets data. The geodetic inversions reveal that the Shumagin seismic gap is multi-segmented, and the M7.8 earthquake ruptured the eastern segment from 14 km down to 44 km depth. The coseismic slip occurred along a more steeply, 26-degree dipping segment, and was bounded up-dip by a bend of the megathrust interface to a shallower 8-degree dip angle connecting to the trench. The model for the M7.6 event tightly constrained the rupture depth extent to 23-37 km, within the depth range of the M7.8 coseismic rupture area. We find that the M7.6 event ruptured the incoming slab across its full seismogenic thickness, potentially reactivating subducted Kula-Resurrection seafloor-spreading ridge structures. Coulomb stress transfer models suggest that coseismic and/or postseismic slip of the M7.8 event could have triggered the M7.6 event. This unusual intraslab event could have been caused by accumulation and localization of flexural elastic shear stresses at the slab bending region. We conclude that the segmented megathrust structure and the location of intraslab fault structures limited the rupture dimensions of the M7.8 event and are responsible for the segmentation of the Shumagin seismic gap. Our study suggests that the western and shallower up-dip segments of the seismic gap did not fail and remain potential seismic and tsunami hazard sources. The unusual earthquake doublet provides a unique opportunity to improve our understanding of the role of the subducting lithosphere structure in the segmentation of subduction zones.

Keywords: Subduction earthquake doublet, slab geometry, Shumagin seismic gap, Alaska subduction zone

1. Introduction

In 2020, a pair of large and similar sized earthquakes (doublet) occurred along the eastern Aleutian subduction zone off the Alaska Peninsula (Fig. 1(a)). The first and largest earthquake of the doublet, with a magnitude (M) of 7.8, occurred at 06:12:44 UTC (22:12:44 local) on July 22 2020. On Oct 19 2020 at 20:54:39 UTC (12:54:39 local), an anomalously large aftershock (M=7.6) occurred 80 km southwest of the first event. According to the U.S. Geological Survey (USGS) hypocentre catalog, both earthquakes are located on the landward side of the subduction trench. The aftershocks of the first event are distributed parallel to the trench, while those of the second event are aligned perpendicular to the trench. The focal mechanism solutions from the Global Centroid Moment Tensor (GCMT) catalog suggest the mechanism of the M7.8 event is of thrust-faulting type, while the M7.6 event was a strike-slip event. The centroid depths of both earthquakes were estimated as about 30-35 km. This suggests that the M7.8 event ruptured the buried megathrust interface, but the M7.6 event was caused by an unusual strike-slip rupture along an approximately trench-normal fault.

The 2020 Shumagin earthquake sequence is interesting for several reasons. Firstly, the mainshock is located within the Shumagin seismic gap. This portion of the subduction thrust has been identified as a seismic gap since the 1970s (Sykes, 1971, Davies et al., 1981). The seismic gap stretches $\sim$200 km along the Shumagin Islands and is bounded to the west by the 1946 M_w8.6 earthquake (López and Okal, 2006) and to the east by the 1938 M_w8.2 rupture (Freymueller et al., 2021). The last earthquakes that are inferred to have ruptured through part of or the whole Shumagin gap occurred in 1993, 1917, 1788, and possibly 1847 (Estabrook et al., 1994). Over the last century, a few moderate (M6.5 to M7.0) events have occurred in the area at depths greater than 30 km. However, the fault sections ruptured by those earthquakes are relatively small compared to the 200 km-long seismic gap (e.g., the estimated rupture area of the 1993 M_s6.9 earthquake is 40 km-long and 15 km-wide, Lu et al. (1994)). If the whole Shumagin gap were fully locked, the accumulated moment equates to 6.6×10^{19} Nm/year, assuming a plate convergence rate of 64 mm/year and a uniform rigidity of 50 GPa. This would require a 7.5 event every 4 years, or a M8 event every 20 years. The lack of historic M7.5+ earthquakes in the Shumagin region has been explained due to substantial aseismic fault creep at seismogenic depths revealed by model inversions of inter-seismic GNSS velocities (Fournier and Freymueller, 2007). Fournier and Freymueller (2007) suggested that instead of rupturing in large earthquakes, most of the seismic moment in the Shumagin gap is released through steady creep. Thus, a moderate M7 earthquake every $\sim$40 years, as observed in the last century, may be sufficient to accommodate the residual slip deficit. To the west of the Shumagin gap, a recent interseismic coupling model shows that the shallow portion along the Sanak segment, 240 km-long and 115 km-wide, might be partially locked, with 15%-25% coupling (Drooff and Freymueller, 2021). For the shallow portion along the Sanak segment, if the estimated 1946 earthquake rupture area, 180 km-long and 115 km-wide (López and Okal, 2006), was fully locked, and the remaining area is 20% coupled, the seismic moment deficit would be accumulating at around 4.5×10^{18} Nm/year. This calculation suggests that the seismic moment of the 1946 earthquake, 8.5×10^{21} Nm (López and Okal, 2006), releases 1900 years of elastic strain accumulation along the Sanak segment. Large uncertainties are associated with estimating earthquake recurrence intervals, including the poorly constrained estimation of the 1946 earthquake slip and the assumption of the highest slip deficit near the trench in the interseismic coupling model

(Drooff and Freymueller, 2021). Nevertheless, such long recurrence intervals could be the reason for there only being one documented major earthquake in the Sanak segment since 1700 (Estabrook and Boyd, 1992).

Secondly, the M7.6 slab-breaking aftershock had an unusual strike-slip mechanism and was deeply buried. Large oceanic lithosphere strike-slip events have previously occurred in the oceanic plate off subduction zones, such as the 2018 $M_w7.9$ Gulf of Alaska earthquake (Lay et al., 2018) and the 2012 $M_w8.6$ Wharton basin earthquakes off-Sumatra (Wei et al., 2013). In addition, the subduction zone outer-rise region regularly hosts normal-faulting mechanisms events. Outer-rise normal-faulting events are attributed to plate bending stresses from slab pull, and can be modulated by the interplate seismic cycle (Ammon et al., 2008). However, the occurrence of a major intraplate earthquake in the oceanic lithosphere just landward of the trench is rare, with only few reported examples, such as the October 4, 1994 $M_w8.2$ earthquake off Shikotan Island along the Kuril trench (Tanioka et al., 1995). Another notable example was a M_w7 strike-slip intraslab event located beneath Kodiak Island, Alaska down-dip of the locked portion of the Alaska-Aleutian megathrust (Hansen and Ratchkovski, 2001).

Thirdly, to our knowledge, these earthquakes are the first documented sizable earthquake doublet to involve a megathrust earthquake rupture, followed by an intraplate strike-slip earthquake tearing the subducting incoming slab. Earthquake doublets are pairs of events with comparable size and likely occur due to earthquake triggering interactions. Subduction earthquake doublets have been studied in the 2006-2007 Kuril and 2009 Samoa doublets (Lay, 2015). In the 2006-2007 Kuril earthquake doublet, a $M_w8.4$

megathrust earthquake was followed by a $M_w8.1$ earthquake rupturing an outer trench-slope normal fault, while in the 2009 Samoa earthquakes, a normal-faulting earthquake ($M_w8.1$) in the outer-rise region triggered a similarly sized thrust-faulting earthquake ($M_w8.0$) on the plate interface (Beavan et al., 2010). Therefore, detailed documentation of this doublet might contribute to the general understanding of the triggering mechanisms during doublets.

In this paper, we use geodetic observations to determine kinematic coseismic fault slip models of the M7.8, M7.6 earthquakes and the postseismic afterslip between two events. We investigate the major controls for the 2020 Shumagin earthquake doublets. We analyze the static fault slip distribution of both events using static GNSS offsets and InSAR surface displacement measurements. The earthquakes ruptured an area off the Alaska Peninsula covered with scattered islands, and incoherence due to water channels makes it challenging to estimate phase ambiguities during the InSAR phase unwrapping process. Hence, we take advantage of an improved Bayesian inversion of wrapped interferometric phase change observations (Jiang and González, 2020) to estimate the fault geometry and slip distribution. Our coseismic geodetic inversion results reveal that the Alaska megathrust has a complex down-dip segmentation. We propose a slab bend structure, which represents a major factor controlling the occurrence and interaction during this doublet, and contributes to the understanding of the mechanics of the subducting oceanic lithosphere in the central Alaska subduction zone.

2. Datasets

2.1. GNSS dataset

We used three-component coseismic offsets and postseismic time series from GNSS stations computed by the Nevada Geodesy Laboratory (Blewitt et al., 2018). The estimated coseismic offsets of the M7.8 event were derived from 5-minute sample rate time series of 108 GNSS stations, using 48 hours of data before and after the mainshock. The coseismic displacements were estimated by subtracting the median position after the mainshock from the median position before. The coseismic displacements of the M7.6 event were derived by subtracting the 24-hour final solutions of 97 GNSS station on October 19 from those on October 20. For the postseismic displacements between the M7.8 and M7.6 events, the displacements can be observed in the GNSS time series of daily solutions. Taking station AC12 as an example, the M7.8 postseismic horizontal displacements in the first day, 2 days, 24 days and 48 days are 0%, 4%, 19%, and 25% of the coseismic horizontal displacements of the M7.8 event. Therefore, we model the 89-day postseismic deformation signal between July 22 to October 19 in the following way: (1) three-component daily solution of 21 GNSS stations less than 500 km away from the epicenter are downloaded; (2) a parametric model is fit to the time series from July 22 to October 19 with an exponential transient decay function (Hearn, 2003), to obtain estimates for the displacement magnitude for each station and a relaxation time (see Fig. S8); (3) we compute a parametric model of postseismic GNSS displacements between July 22 and October 19, and subtract the displacements on July 22 from those on October 19.

2.2. InSAR dataset

We imaged the ground surface displacement caused by the 2020 Shumagin earthquake doublet using InSAR. Satellite radar interferograms capture line-of-sight (LOS) motion away or towards the satellite. We used 12 European Space Agency Sentinel-1 satellite interferometric wide swath mode images to make six interferograms from three different satellite tracks from July 10 to November 7, 2020. We used data from two parallel descending tracks to cover the epicentral area around the Shumagin and neighboring islands (track 73 and track 102). The ascending track 153 fully images the epicentral area. We processed the coseismic interferograms using the TopsApp module of the ISCE software. We removed the topographic phase contribution in the interferograms using SRTM 30-m resolution digital elevation model.

Our interferograms (Table 1) spanning the M7.8 event are dominated by the coseismic deformation signals (Fig. 1(c)-(f)). We generate a preseismic interferogram (Fig. S1) which is dominated by the turbulent atmosphere phase delays. The two descending-track coseismic interferograms span less than 2 days of early postseismic deformation. However, for the coseismic interferogram in ascending track 153, the first available secondary image was acquired 48 days after the earthquake. Hence it could be affected by post-seismic deformation. We use the same strategy as described in Section 2.1 to model the 48-day postseismic deformation signal (Fig. S9), and then we forward-simulate and remove the line-of-sight phase change from the ascending interferogram (track 152, Jul 22-Sep 08) during the first 48 days of the postseismic period. Although, the correction is relatively small, our approach reduces the leakage of postseismic deformation into our coseismic models.

For the interferograms covering the M7.6 event, the interferometric phase observations are also dominated by the coseismic deformation signals. The estimated postseismic relaxation time for the M7.8 event is approximately 40 days, while the acquisition dates of the primary images of the interferograms for the M7.6 event are 84 and 86 days after the M7.8 event, that is 5 and 3 days before the M7.6. Therefore, any M7.8 postseismic deformation signal can be considered negligible. The interferograms spanning the M7.6 event might contain 7, 9 and 19 days of postseismic deformation of this event. However, we did not find clear transient displacement signals either in the postseismic interferograms or GNSS time series during the M7.6 early-postseismic period, so we performed no corrections on the coseismic interferograms.

3. Methodology

3.1. Fault geometry: Non-linear surface displacement inversion

To determine the fault geometry of the ruptures, we invert for an elastic uniform slip rectangular dislocation model. First, we solve for the fault parameters using only the coseismic horizontal and vertical GNSS offsets using the GBIS package (Bagnardi and Hooper, 2018). However, the sparse spatial distribution of the GNSS stations does not allow to tightly constrain the fault geometry (Fig. S2). Thus, we take advantage of independent high-spatial-resolution InSAR observations over the Shumagin Islands and the imaged far-field surface deformation over the Alaska Peninsula to refine the rupture fault geometry.

Standard modeling approaches of InSAR observations require unwrapping the wrapped phase from [-π,π] to the absolute unwrapped LOS displacements. However, phase unwrapping is an ill-posed problem requiring integration along a path connecting pixels. In the Shumagin islands case, the incoherence due to water channels between islands makes the phase unwrapping especially challenging. Any phase unwrapping of coseismic interferograms might contain unknown multiples of 2π between islands (see Fig. S12) due to the dense gradient of fringes. Instead, our method skips the phase unwrapping step, and directly inverts for fault source parameters by applying the WGBIS method, a Bayesian algorithm that minimizes the weighted wrapped phase residuals (Jiang and González, 2020). Now, using the wrapped InSAR phase and GNSS offsets, we can constrain more tightly the fault geometry parameters (Fig. 2).

3.2. Distributed slip models

Next, we propose an extension to the WGBIS method to estimate distributed fault slip directly from InSAR wrapped phase observations applying a novel physics-based fault slip regularization. Traditional kinematic fault slip inversion method used static observations to solve for the slip displacements but neglected to consider the driving forces or stresses that cause these motions. Recently, a laboratory-derived crack model was introduced to describe the relationship between stress and slip on the fault (Ke et al., 2020). Instead of a uniform stress drop across the whole fault plane, this model allows a constant stress drop in the crack center while keeping the stress concentration at the rupture tip finite, and it retains a smooth transition in between. The preferred shape of the crack model, an ellipse, is supported by mechanical considerations (Sendeckyj, 1970). Ke et al. (2020) proposed an analytical model of the slip profile from the centre of the crack to the rupture tip, and we expand this one-dimensional model into a two-dimensional model with an elliptical shape, by assuming one of the focal points of the ellipse to be the crack centre and the elliptical perimeter to be the crack tip. Therefore, the slip distribution s on the fault plane is controlled by a very reduced set of parameters, our crack model contains only seven parameters $\mathbf{m}$, $s = \mathbf{f}(\mathbf{m})$.

$$\mathbf{m} = \{x_0, y_0, a, e, \lambda, d_{max}, \theta\} \tag{1}$$

where x_0, y_0 are the locations of the focal point, and e is the eccentricity of ellipse, λ is the ratio controlling the displacement transition from the center to the edge of the elliptical crack, d_{max} is the maximum slip, and θ is the rake angle. We design synthetic tests (see Fig. S13) to validate our approach, and compare the performance with respect other slip-inversion methods (Amey et al., 2018).

We name our method, the Geodetic fault-slip Inversion using a physics-based Crack MOdel, hereafter referred to as GICMo. The forward model proceeds as follows: (1) the crack model parameters are provided and slips for all fault patches are determined based on the two-dimensional crack model discussed above; (2) the surface displacements are computed by integration over the fault slip distribution set; (3) for the inversion, we follow Jiang and González (2020), using a misfit function based on the wrapped phase residuals and the weighting matrix of observations. This misfit function is then regarded as the likelihood function, and used to retrieve the posterior distribution of crack model parameters by a Bayesian sampling process.

We rationalize our choice for a simple elliptical crack model, firstly because the resolution power of InSAR and GNSS data to constrain the subsurface slip distribution decrease with the fault depth and off-shore distance. Deeper earthquake sources will produce less surface deformation than shallower events of the same size, and hence the detailed distributions of fault slip of deep sources are not well resolved. Second, the published M7.8 earthquake coseismic slip distributions agree on the most notable feature: a high fault slip area with rather smooth slip distribution on the plate interface beneath the Shumagin Islands (Crowell and Melgar, 2020, Ye et al., 2021). This first-order pattern is well resolved by our GICMo model. Third, a simple circular crack is also a widely accepted model to estimate the stress drop of earthquakes using the observed seismic spectra (Madariaga, 1976). In addition to the desirable physics-based properties (finite shear stress at the crack tip), another advantage of this method is its low dimensionality. The model is parametrized using fewer parameters than usually needed to describe the spatial pattern of slip distributions. Previous inversion algorithms using deterministic or Bayesian approaches allow for highly complex patterns of slip distributions by allowing unconstrained or regularized slip distributions (Fukahata and Wright, 2008). However, those methods are solving very high dimensional problems with larger associated null-spaces, and are also computationally more intensive.

3.3. Coulomb stress models

The Coulomb stress theory has been extensively applied to study the interaction between earthquakes. Coulomb stress change induced by fault slip is a quantitative measure that has been correlated with the aftershock distribution, seismicity rate changes and earthquake triggering. Usually, more aftershocks occur in the high stress-change region. It is thought that increases in Coulomb stress of 0.01 MPa are sufficient to trigger events (King et al., 1994). In our study, we calculate the Coulomb stress changes due to the M7.8 event and investigate whether the M7.8 earthquake and its afterslip promoted failure of the subsequent M7.6 event. We use the Coulomb 3.3 program to carry out the stress calculations, which is based on the dislocation model algorithms (`https://www.usgs.gov/software/coulomb-3`).

4. Results

4.1. Coseismic model for the M7.8 earthquake

The Shumagin earthquake nucleated near the eastern edge of the Shumagin seismic gap (Davies et al., 1981). Our static surface displacement inversions suggest that the coseismic rupture extended for 112$\pm$2 km to the WSW from the location of the USGS hypocentre (red rectangles in Fig. 2), with an average pure thrust slip of 1.5$\pm$0.1 m, corresponding to an estimated M7.8. The buried rupture extended down-dip to 44$\pm$2 km and up-dip up to 14$\pm$2 km depth and did not break the seafloor at the Alaska-Aleutian trench. This is consistent with reports of a minor tsunami (Ye et al., 2021). A remarkable feature of our inversion results is that the inferred fault geometry requires a relatively steep dip angle (26$\pm$0.5 degrees, Fig. 2(b)), steeper than the widely used Slab2 subduction model ($\sim$15 degrees, Hayes et al. (2018)). We further investigate this feature by separately inverting for the fault geometry using GNSS coseismic offsets only, InSAR wrapped phase only and both observations. In all cases, the obtained fault geometries are consistent with a 25-to-28 degree fault rupture plane. The GNSS only inversion suggests a slightly steeper fault dip angle (Fig. S2), than the 26$\pm$0.5-degrees dip angle, obtained using only the InSAR wrapped phase or both datasets (Fig. S3-S4).

Next, we use our estimated fault geometry model to refine the location and pattern of coseismic slip during the earthquake. We tested two different 3D fault geometry parameterizations. The first 3D fault geometry, based on the estimated fault geometry, contains two segments. A deeper segment dipping 26-degree from 14 km to 44 km depth using the optimal rectangular dislocation plane estimated by the non-linear inversion, and then a shallower segment connecting the top edge of the rectangular plane to the trench. These fault planes were then discretized into a triangular mesh with patch dimensions of $\sim$5 km. A second geometry was obtained based on the Slab2 model for the Alaska megathrust, which has an average dip of 15 degrees from 20 km to 50 km (Hayes et al., 2018).

We solve for the slip distribution of our elliptical rupture model on the Slab2 and our proposed fault geometry (Fig. 2). Fig. 3 shows the observed and modeled GNSS displacements and the wrapped interferometric phase, as well as the residuals using the proposed down-dip structure. The modeled phase is consistent with the observed phase. The root-mean-square (RMS) of the GNSS residuals in the east, north and vertical directions are 0.3, 0.3 and 0.6 cm, and corresponding to data variance reductions of 98%, 99%, and 97%. The GNSS offsets can be fit comparably well with both interface geometries (see Fig. S5-S7). However, the distributed slip model on the Slab2 geometry cannot reproduce the InSAR surface displacement patterns as well as those with the optimized, steeper fault geometry (Fig. 3, S5-S6).

Moreover, the posterior probability distribution functions on the elliptical rupture model parameters are less well resolved for the Slab2 fault parameterization (Fig. S7). Our final slip model (Fig. 4(a)) shows a patch of large slip near the hypocenter and below the Shumagin Islands, consistent with kinematic coseismic slip models constrained using near-field high-rate GNSS and strong-motion data showing a more broadly distributed slip (Crowell and Melgar, 2020), and the finite-fault slip model using joint inversion of teleseismic P and SH waves and static displacements from regional GPS stations (Ye et al., 2021). The peak slip is 1.7 m, and the average slip is 0.7 m. The fault slip distribution inverted from GNSS and three interferograms is shown in Fig. 4(a). The total geodetic moment is 6.12×10^{20} Nm, which is equivalent to $M_w7.79$, a value consistent with the seismic moment magnitude of $M_w7.8$. The estimated rupture centroid is located at [158.834°W, 55.130°N] and the centroid focal depth is 32 km, which is deeper than the 28 km estimated by USGS and 19 km away from the USGS-estimated hypocenter, [158.596°W, 55.072°N], in the northwest direction. Furthermore, as shown in Fig. 4(b), few aftershocks are located close to the slip peak, and most seismic events occurred near the edges of the estimated rupture area.

4.2. Postseismic model for the M7.8 earthquake

The M7.8 postseismic phase is important to study the whole doublet sequence, so we quantify the amount and distribution of early postseismic slip caused by the M7.8 event. As afterslip is unlikely to be compact, and may fully surround the coseismic rupture, the spatial distribution of postseismic slip is resolved by using the slip inversion package, slipBERI (Amey et al., 2018). This method incorporates the fractal properties of fault slip to regularize the slip distribution. We assume that afterslip dominates the observed surface deformation during the 89-day-long period between the M7.8 and the M7.6 events. Afterslip describes postseismic aseismic fault motions occurring near the mainshock rupture regions over several months to several years. Postseismic offsets are estimated by fitting the daily GNSS data from July 22 to October 19 with a simple exponential model and then inverted for the postseismic slip distribution. Compared with the coseismic model 3D fault discretization, the subduction zone interface is extended along strike and down-dip to investigate the distributed postseismic slip over a wider area of the plate interface. The model predictions agree well with GNSS observations (Fig. S8), and the RMS of the GNSS residuals in the east, north and vertical directions are 0.6, 0.6 and 0.7 cm, respectively.

We find the postseismic afterslip region mainly covered the deep portion (>50km depth) of the plate interface (Fig. S10 and orange lines in Fig. 4(a)). A small patch, 60km-long and 40km-wide, is inferred to have slipped aseismically in the very shallow portion (6-9 km depth). In the depth range of 14-44 km, where the M7.8 earthquake ruptured, no strong afterslip is revealed. The 3-month postseismic slip has a cumulative geodetic moment of 10^{20} Nm, corresponding to M_w 7.27, assuming a variable crustal shear modulus with depth from CRUST 1.0. We try different slip variance values and rupture dimensions in slipBERI and it does not change those spatial characteristics substantially.

Crowell and Melgar (2020) estimated the first 10 days of the postseismic afterslip, finding that the majority of afterslip is concentrated downdip of the mainshock between 40-60 km depth. Their model is generally consistent with our finding of afterslip dominantly occurring at greater depth. They also argued that although afterslip might occur up-dip of the M7.8 earthquake, the current configuration of GNSS stations is insensitive to the afterslip at shallow depth. Recently, Zhao et al. (*pers. comm., 2021*) applied additional constraints to regularize the afterslip distribution, where they considered both stress-driven frictional models and kinematic inversions in which no slip is allowed within the coseismic peak slip zone. The models of Zhao et al. (*pers. comm., 2021*) suggest possible afterslip in the up-dip area of the M7.8 earthquake.

4.3. Coseismic model for the M7.6 earthquake

To parameterize the geometry of the October 19 2020 M7.6 Shumagin earthquake, we consider the spatial distribution of its aftershocks. Most of these aftershocks occurred at the western edge of the coseismic slip area of the M7.8 event. Aftershocks aligned in a north-south direction, parallel to the plate convergence direction. We first approximate the dimensions of the rupture area using the aftershock locations in the first two days after the M7.6 event. The estimated rupture area dimensions from the aftershocks are 100-150 km long and 50-60 km wide and dipping 38 degrees to the east. Those parameters are consistent with the focal mechanism from GCMT catalog (dip angle=49°, strike angle=350°) and the inverted parameters for a rectangular dislocation source (length=50 km, width=20 km, top depth=23 km, bottom depth=37 km, dip angle=44.5°, strike angle=358.5°, strike slip=3.2 m). Our slip model reproduces well the coseismic deformation observed by GNSS and InSAR (Fig. 3 and Fig. S11).

The coseismic slip model shows right-lateral strike-slip motion on a fault plane perpendicular to the Alaska subduction zone, consistent with the distribution of the aftershocks. The aftershocks following the M7.6 event occurred at the periphery of the coseismic rupture (Fig. 1(a)), effectively extending the latter farther to the south and were dominated by strike-slip rupture mechanisms with east-dipping north-south-striking nodal planes. The total moment release from the coseismic slip was 2.1×10^{19} Nm, assuming a variable crustal shear modulus with depth based on the CRUST 1.0 model. The corresponding moment magnitude is M_w=7.5, in reasonable agreement with the seismically determined value. We also estimated the stress drop to be 6.6 MPa, which is within the usual bounds of intraplate earthquakes (Allmann and Shearer, 2009).

Our model suggests that the rupture zone is located from 23 km to 37 km in depth, beneath the slab interface (Fig. 2(c)). This reveals that the M7.6 strike-slip earthquake ruptured the subducting oceanic slab, rather than the forearc. This is also confirmed from the focal depth range of the aftershocks. 70% of the M2.5+ aftershocks in the first 2 days after the mainshock occurred at 20-40 km depth. A significant non-double-couple component in the moment tensor, the substantial tsunami and the residuals of the GNSS vertical component (Fig. 3(d)) indicate that another shallow rupture segment parallel to the trench might exist (Lay, 2021), but our geodetic inversion cannot resolve a second segment at shallow depth. Our inversions for two segments using the geodetic data are not stable, which might be limited by the minor deformation signals on the islands.

5. Discussion

5.1. Influence of slab geometry on the rupture characteristics of the M7.8 earthquake

Our preferred coseismic rupture model constrains the deep structure of the Alaska megathrust along the Shumagin segment. It reveals a ∼26 degrees dipping interface from 14 to 44 km depth. The megathrust interface at shallower depths is a gentler dipping segment (∼8±4 degrees) of 90 km width, connecting the up-dip edge of the rupture to the trench (Fig. 2). This plate-interface geometry substantially deviates from the most recent subduction interface model Slab2, which is based on regionally and globally located seismic events (Hayes et al., 2018). The Slab2 model suggests a 15-degrees dip in the depth range from 20 to 50 km. This might indicate that the steeper segment could be a relatively localized structural feature along this section of the subduction zone. The discrepancy with the Slab2 model might be due to smoothness constraints applied to the subduction zone model, which might not resolve length-scales similar or smaller than those of the Shumagin gap (100-200 km). This highlights the need to create additional regional models that capture finer spatial structural details to improve subduction zone seismic hazard assessment.

Seismic reflection imaging along profiles across the Shumagin segment suggests a geometry similar to our inversion results (Li et al., 2015). In Fig. 2(b), we show the interpreted seismic reflection data from Line 4 of Li et al. (2015). Line 4 is located in the proximity of the M7.8 rupture area, at the boundary of the Semidi segment and Shumagin seismic gap. The seismic reflectors are consistent with our inferred fault geometry. Our plate interface geometry also agrees with a fault geometry grid search using GNSS vertical coseismic offsets caused by the M7.8 earthquake by Crowell and Melgar (2020). Their dislocation models also support a 25-degree dip fault geometry, with an up-dip edge at 21$\pm$2 km and extending down to 45$\pm$5 km depth.

We also note that the M7.8 down-dip rupture limit approximately coincides with the depth of the continental Moho, imaged by the seismic reflection data at 39-41 km depth. This is consistent with a first-order correlation of the base of the seismogenic zone and the base of the continental crust (e.g., Oleskevich et al. (1999)), but exceptions to this pattern have been noted (Simoes et al., 2004). A zone of low-frequency tremor sources (Brown et al., 2013) is located at $\sim$50km depth, and there is a gap between the seismogenic zone and the area hosting tremor, which is also observed in the Nankai and Cascadia subduction zones (Gao and Wang, 2017). The bottom of the rupture likely reached the down-dip limit of the locked seismogenic zone. Recently, Shillington et al. (2021) analyzed the seismic reflection data from nearby Line 5 (Fig. 2(a)) and found the continental Moho depth at 35 km, with less uncertainties than Line 4 (Li et al., 2015). If this is confirmed, it might suggest that part of the coseismic slip extended downdip of the continental Moho (or mantle wedge corner). This coseismic slip feature was previously observed in very large megathrust events, e.g. the 2010 M8.8 Maule, Chile, earthquake (Weiss et al., 2019). One of the explanations could be that hydrated materials (e.g., serpentinites) along the base of the mantle wedge control the frictional properties of the megathrust, and allow the propagation of large ruptures, even though the megathrust downdip of the mantle wedge corner is predominantly velocity strengthening (Wang et al., 2020, Kohli et al., 2011).

Our findings suggest that the fault geometry controls the rupture size and extent. A similarly large buried rupture was observed during the 2015 Gorkha, Nepal earthquake on a continent-continent subduction zone (Elliott et al., 2016). Hubbard et al. (2016) developed a fault morphological model consisting of two ramps and found that the location and shape of coseismic fault slip (>1m) match well with the location and shape of the middle decollement bounded on both sized by ramps. Therefore, they proposed that the variations in fault dip angle controlled the shape and size of the mainshock rupture in this continental megathrust earthquake. Decollement-ramp structures formed in subducting sediments are not rare in global subducting zones (Seno, 2017). About 1-km-thick subducting sediments were inferred from seismic reflection data beneath the eastern Shumagin gap (Li et al., 2018) and clear variations of the megathrust dipping angle were revealed at 7 km and 17 km (Li et al., 2015), which is consistent with the top rupture depth at 14 km. In summary, the variation in fault orientation with depth was likely a controlling factor limiting the extent of the Shumagin rupture.

5.2. The M7.6 slab-tear earthquake source region

If we assume that the M7.6 earthquake occurred on a pre-existing fault plane, this fault had a 16-degree strike and 60-degree dip prior to being subducted. This strike angle is consistent with the strike of the Kula-Resurrection ridge (Fig. 6(a), Fuston and Wu (2020)). These ridges were active from 60 to 40 Ma, producing north-south striking faults through the seafloor spreading, and have been inactive since $\sim$40 Ma. The inferred dipping angle of the pre-existing fault is consistent with the dip angle of mid-ocean-ridge normal faults. The pre-existing faults are unlikely to be formed in the outer-rise region because the outer-rise bending faults are parallel to the trench with approximately east-west strike directions (Shillington et al., 2015). The pre-existing faults are unlikely to have formed along the Pacific-Kula ridge or the Pacific-Farallon ridge, because the orientation of the magnetic anomalies (east-west and northwest-southeast) are inconsistent with the eventual strike of the M7.6 ruptured fault.

In addition, our M7.6 fault model is correlated with the location of a low seismic-velocity anomaly, which has been attributed to higher slab hydration (Li et al., 2020). Li et al. (2020) imaged the crust and uppermost mantle structure of the Alaska subduction zone using ocean bottom seismographs and broadband seismic stations. They constructed a 3-D shear velocity model, where one trench-normal profile (TT1) is just $<$5 km away from the M7.6 rupture area. They found upper mantle shear-velocity reductions along this profile of about 15% (from $\sim$4.6 to $\sim$3.9 km/s), which extends more than 12 km beneath the Moho. In other regions along the Alaska subduction zone (e.g., Semidi segment), the upper mantle velocity reduction is only about 11% (from $\sim$4.6 to $\sim$3.9 km/s). They interpret this feature as evidence of stronger hydration of the incoming plate along the Shumagin seismic gap. Furthermore, in the outer-rise region of the Shumagin Islands (Line 5 in Fig. 2(a) and Fig. 4(a)), Shillington et al. (2015) found a P-wave velocity reduction in the upper mantle from 8.25 to 7.75 km/s, associated with abundant bending faults. These observations lend further support to the existence of faults in the subducted slab beneath the Shumagin Islands, which might have played a major role in the location of seismogenic ruptures.

Here, we propose a simple mechanical model that partially explain the location of the M7.6 event. As previously shown, the M7.6 strike-slip event ruptured the incoming slab near a bend in the down-dip geometry of the plate interface. This bend could localize deformation. Knowing that subducting lithosphere is subject to flexural bending shear stresses, which are large enough to break the crust in the outer-rise region, we propose that the M7.6 could have partially been caused by accumulated flexural bending shear stresses in addition to lateral stress loading variations along the trench. Here, we assume that the rheological behaviour of the oceanic lithosphere can be approximated by that of an elastic beam. The deflection of the oceanic lithosphere is, to the first order, controlled by the gravitational body forces and bending moment acting on the descending plate (Fig. 5). So, we can compute the shear stress rate dV acting on the elastic lithosphere as a function of the distance from the trench, X (Turcotte and Schubert, 2014):

$$dV = \frac{\sqrt{2}\pi^3 e^{\pi/4}}{32A} \frac{Dw_b}{(X_b - X_0)^3} \left[\cos\left\{ \frac{\pi (X - X_0)}{4 (X_b - X_0)} \right\} + \sin\left\{ \frac{\pi (X - X_0)}{4 (X_b - X_0)} \right\} \right] \exp\left[-\frac{\pi (X - X_0)}{4 (X_b - X_0)} \right] \quad (2)$$

where X_0 is the location of the trench, X_b is the location deflection forebulge with height w_b, and A is the slab age. The flexural rigidity parameter, D is given by the expression $D = \frac{ET_e^3}{12(1-\nu^2)}$, which is a function of the effective elastic thickness (T_e), the Young's modulus (E) and the Poisson's ratio (ν). To simulate the shear stress rate dV acting on the Shumagin segment, we use parameters from Zhang et al. (2018) as listed in Table 2. The estimated shear stress rate at the location and along the length of the M7.6 rupture fault varies from 0.006 to 0.05 MPa/year. If we compare these values with respect to the coseismic stress drop, 6.6 MPa, the M7.6 event could have released 130∼1100 years of accumulated bending shear stress. In addition to flexural bending shear stress, shear stress directions could be controlled by the slab geometry variations along trench-parallel direction. For example, the downdip plate geometry from 10-50km depth along Line 5 (Shillington et al., 2021) is smoother than that along Line 4 (Li et al., 2015).

Alternatively, shear stresses could be caused by spatial variations of elastic coupling along the megathrust interface. Herman and Furlong (2021) present models that simulate the effect of laterally variable coupling. The preferred models represent the Semidi segment to be highly coupled while the Shumagin segment has low coupling. The lateral displacement variations can impose large-magnitude, right-lateral shear stresses on the M7.6 rupture plane geometry, assuming the target fault plane was north-south striking and east dipping with a dip angle 50°. However, we note that the available geodetic observations infer only 30%±10% coupling in the western portion of Semidi segment (Drooff and Freymueller, 2021) which is much lower than the 100% assumed by Herman and Furlong (2021), for the whole Semidi segment. Therefore, interseismic coupling variation between the Semidi and Shumagin segments may contribute to the shear stress accumulation on the M7.6 rupture plane, but geodetic evidence suggests this contribution may be more modest in magnitude. Hence, lateral variations of coupling, the existence of structural weaknesses and long-accumulated bending flexural shear stress could explain the occurrence of the M7.6 slab breaking event, which broke the entire seismogenic thickness.

5.3. Mechanisms for the interaction between the two earthquakes

Earthquake doublets are not uncommon and suggest short-term fault interactions and triggering. Lay (2015) compiled 7 pairs of earthquake doublets in subduction zones, where he proposed that stress transfer and triggering interactions are clearly demonstrated by several doublet sequences and the complexity of faulting of many of the events. To investigate the possible relationship between these events, we calculate the stress perturbations on the M7.6 event associated with the Jul 22 2020 M7.8 coseismic and postseismic slips (Fig. 4(d)). We utilize the inferred slip distribution from our inversion model for the M7.8 event. Then, we compute the stress change on the estimated fault plane of the M7.6 event. We extend the M7.6 rupture fault plane along dip from the surface down to 60 km depth, and compute the stress change on a regular grid with 5 km-wide patches. The M7.8 earthquake caused a shear stress increase of 0.07 MPa and tensile normal stress increase of 0.27 MPa around the hypocenter, while the contributions from the postseismic slip are almost neutral. Our Coulomb stress models suggest that the second, M7.6 intraslab, earthquake was likely triggered by the elastic stress changes transferred by the slip during the M7.8 coseismic slip on the megathrust interface, with postseismic deformation processes possibly explaining the ~3-month delay in the occurrence of the large intraslab event.

6. Conclusions

We conclude that the 2020 Shumagin earthquake doublet represents a rare example of two deeply buried ruptures on a subduction megathrust and an oceanic intraplate strike-slip fault (Fig. 6(b)). The first M7.8 earthquake partially ruptured the Shumagin seismic gap, along a 112 km-long, 65 km-wide section, extending from 14 km to 44 km depth. The second M7.6 event was likely triggered by static stress changes due to the M7.8 coseismic slip and could have released 130$\sim$1100 years of accumulated flexural bending shear stresses. The M7.6 broke the incoming oceanic plate at moderate depths from 23 km to 37 km along a north-south striking and east-dipping, right-lateral strike-slip fault. We propose that the Shumagin gap is segmented and has variable mechanical characteristics. The M7.8 earthquake ruptured a distinct eastern segment of the Shumagin gap, while the western segment and shallow portions remain unruptured. We highlight that the inferred rupture geometry of the M7.8 event is substantially steeper compared to the Slab2 model. The variations of down-dip megathrust structure of the Shumagin segment might have implications for seismo-tectonics and tsunami hazard of this segment of the Alaska-Aleutian subduction zone, e.g., by controlling the degree of coupling and seismic segmentation of the megathrust interface (Fournier and Freymueller, 2007, Hayes et al., 2018), and influencing coseismic and postseismic slip distributions (Crowell and Melgar, 2020). In addition, we identify Kula-Resurrection ridge fault structures imprinted in the oceanic lithosphere as the likely earthquake source plane reactivated during the M7.6 event. Our study highlights that the reactivation of such oceanic lithospheric structures might pose an important seismic hazard in subduction zones, and might represent favorable pathways for fluid flow and dehydration of the subducting slab.

7. Acknowledgments

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The bathymetry data is retrieved from SRTM30_PLUS (`https://topex.ucsd.edu/WWW_html/srtm30_plus.html`). The earthquake catalog is retrieved from USGS (`https://earthquake.usgs.gov/earthquakes/search/`). The coastal data is retrieved from NOAA (`https://www.ngdc.noaa.gov/`

mgg/shorelines/). This is a contribution of the CSIC Thematic Platform PTI-Teledect (https://pti-teledetect.csic.es/). CRUST 1.0 model is retrieved from https://igppweb.ucsd.edu/~gabi/crust1.html.

Table 1: Details of Sentinel-1 interferograms for 2020 Shumagin earthquake doublet, M7.8 and M7.6 event.

Earthquake date and magnitude	Track no.	Direction (asc/des)	Incidence (degree)	Primary image (yyyy/mm/dd hh:mm:ss)	Secondary image (yyyy/mm/dd hh:mm:ss)
2020/07/22 06:12:44 M7.8	73	des	30-33	2020/07/10 17:03:59	2020/07/22 17:04:00
	102	des	43-46	2020/07/12 16:47:32	2020/07/24 16:47:33
	153	asc	36-41	2020/07/22 04:23:36	2020/09/08 04:23:39
2020/10/19 20:54:38 M7.6	73	des	30-35	2020/10/14 17:04:04	2020/10/26 17:04:04
	102	des	43-46	2020/10/16 16:47:36	2020/10/28 16:47:36
	153	asc	34-41	2020/10/14 04:23:40	2020/11/07 04:23:40

Table 2: Variables for shear stress rate calculation (Zhang et al., 2018)

Symbol	**Variables**	**Value**	**Unit**
E	Young's modulus	7×10^{10}	Pa
ν	Poission's ratio	0.25	-
T_e	Effective thickness of oceanic lithosphere	18.2	km
w_b	Height of the forebulge	0.18	km
x_0	Location of the trench	0	km
x_b	Location where the deflection is w_b	42.7	km
A	Plate age	54	Ma

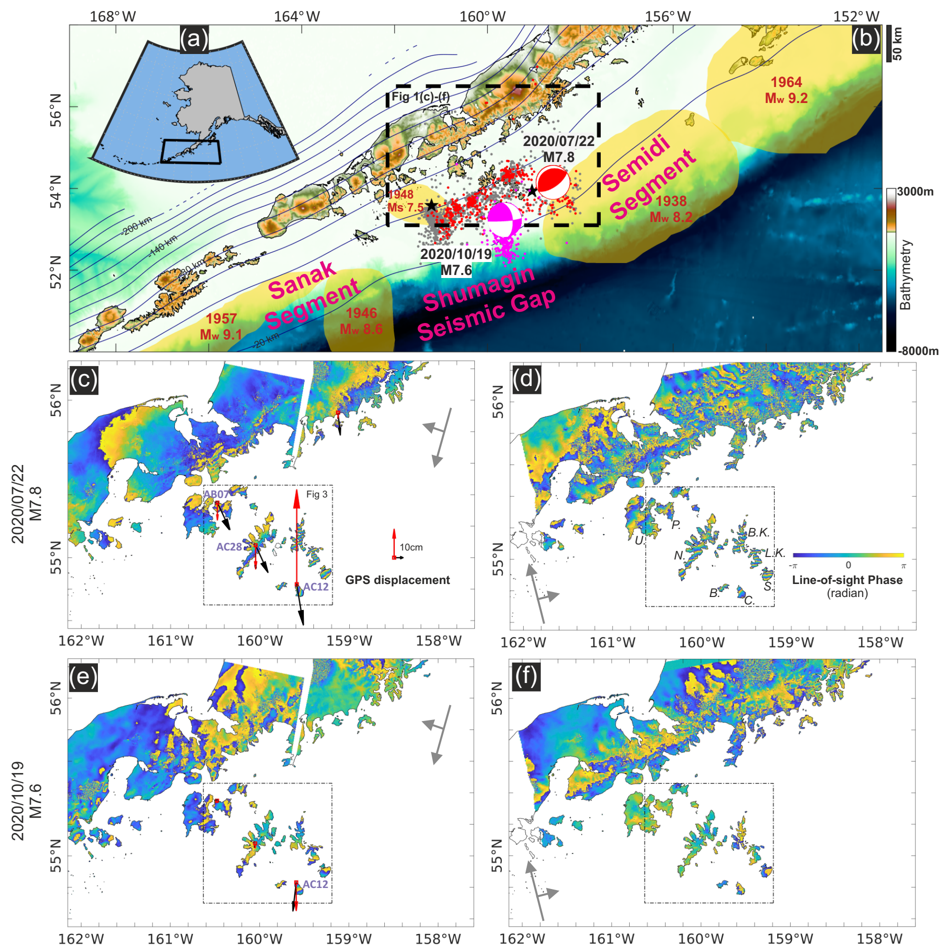

Figure 1: Tectonic background and geodetic observations of the Shumagin earthquake doublet, 2020/07/22 M7.8 earthquake and 2020/10/19 M7.6 earthquake. (a) inset shows the Aleutian subduction zone. Panel (b) shows historic ruptures as shaded yellow areas, on top of the bathymetry as the background. The Shumagin seismic gap is the 200 km-long region between the 1946 M_w8.6 and the 1938 M_w8.2 earthquakes. The M7.8 and the M7.6 events are plotted as red and magenta beachballs. The first 2-day and 3-month aftershocks following the M7.8 event are plotted as red and gray dots, where two M6+ events are plotted as little black stars. The first 2-day aftershocks following the M7.6 event are plotted as magenta dots. The dashed box shows the boundary for images (c)-(f). Panel (c) shows the wrapped phase of two descending interferograms, 2020/07/10-2020/07/22 (Track 73) and 2020/07/12-2020/07/24 (Track 102). The arrows show the GNSS

Figure 1 (continued): displacements retrieved from Nevada Geodesy Laboratory, red for vertical, and black for horizontal displacements. AB07, AC28 and AC12 are three GNSS stations with the most significant movement. GNSS displacement at [158.5W,55N] is the unit displacement vector for 10cm vertical and horizontal displacement. The dotted-dashed box marks area in Fig. 3. Panel (d) shows the ascending interferogram 2020/07/22-2020/09/08 from track 153. Panel (e) and (f) is same with Panel (c) and (d), but three interferograms covering the M7.6 event, of two descending interferograms, 2020/10/14-2020/10/26 (Track 73) and 2020/10/16-2020/10/28 (Track 102), and one ascending interferogram, 2020/10/14-2020/11/07 (Track 153). Island abbreviations: U.: Unga; P.: Popof; N.: Nagai; B.K.: Big Koniuji; L.K.: Little Koniuji; B.: Bird; C.: Chernabura; S.: Simeonof.

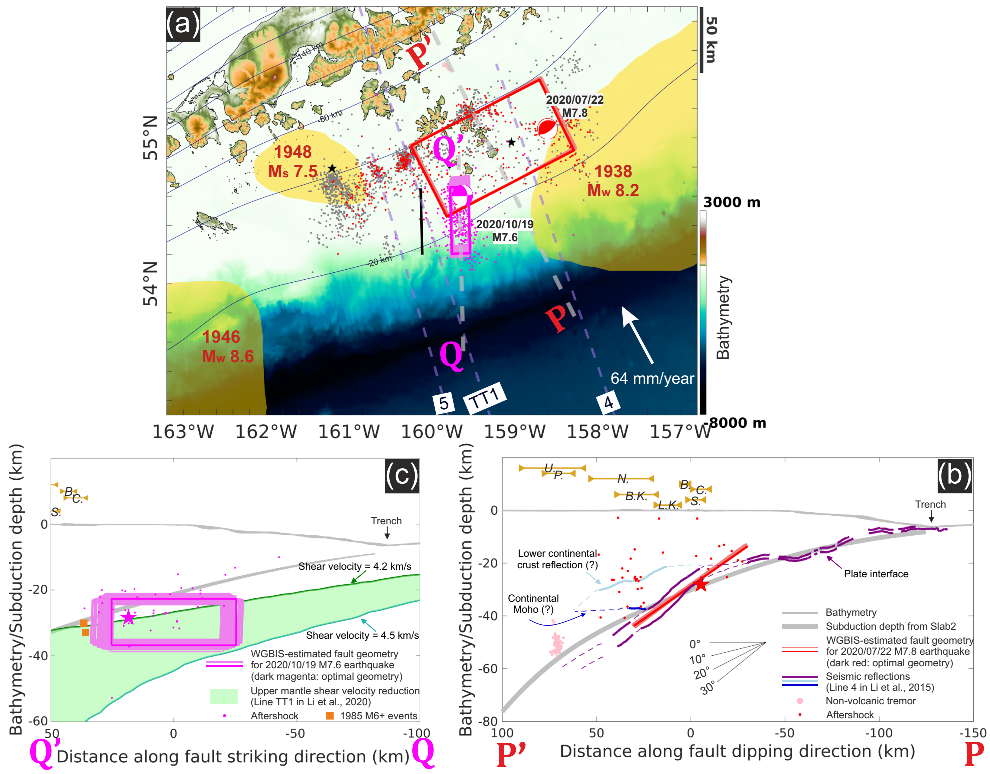

Figure 2: Preferred geodetic fault model constrained using InSAR wrapped phase and GNSS. In panel (a), red and magenta rectangles outline the ensemble of inverse bayesian fault geometry models for the M7.8 and the M7.6 earthquakes. The black line west of the magenta rectangle indicates its projection to the surface. The dashed purple lines 4, 5 and line TT1 indicate the position of a seismic reflection line from Li et al. (2015) and a shear velocity profile from Li et al. (2020); the dashed gray lines are profiles PP' and QQ' shown in (b) and (c). Panel (b) shows a cross-section of the inferred fault geometry models of the M7.8 earthquake projected to profile PP'. We also show the geophysical interpretation of the reflection lines (Line 4 and Fig. 5 in Li et al. (2015)), and locations of tremor (Brown et al., 2013). The cross-section also shows Slab2 model (depth to the top of subducting plate) and the bathymetry along profile PP'. Brown lines show the projected location of islands with the same abbreviations as Fig. 1(d). Panel (c) shows a cross-section of the inferred fault geometry models of the M7.6 earthquake projected to profile QQ'. We also show the shear velocity reduction zone within upper mantle, constrained by the shear

Figure 2 (continued): velocity 4.2 km/s and 4.5 km/s and digitized from Li et al. (2020). Two orange blocks present the M6+ subducting events in 1985 <5 km from the northern end of the fault model (https://earthquake.usgs.gov/earthquakes/eventpage/usp0002kmh/executive, https://earthquake.usgs.gov/earthquakes/eventpage/usp0002msj/executive).

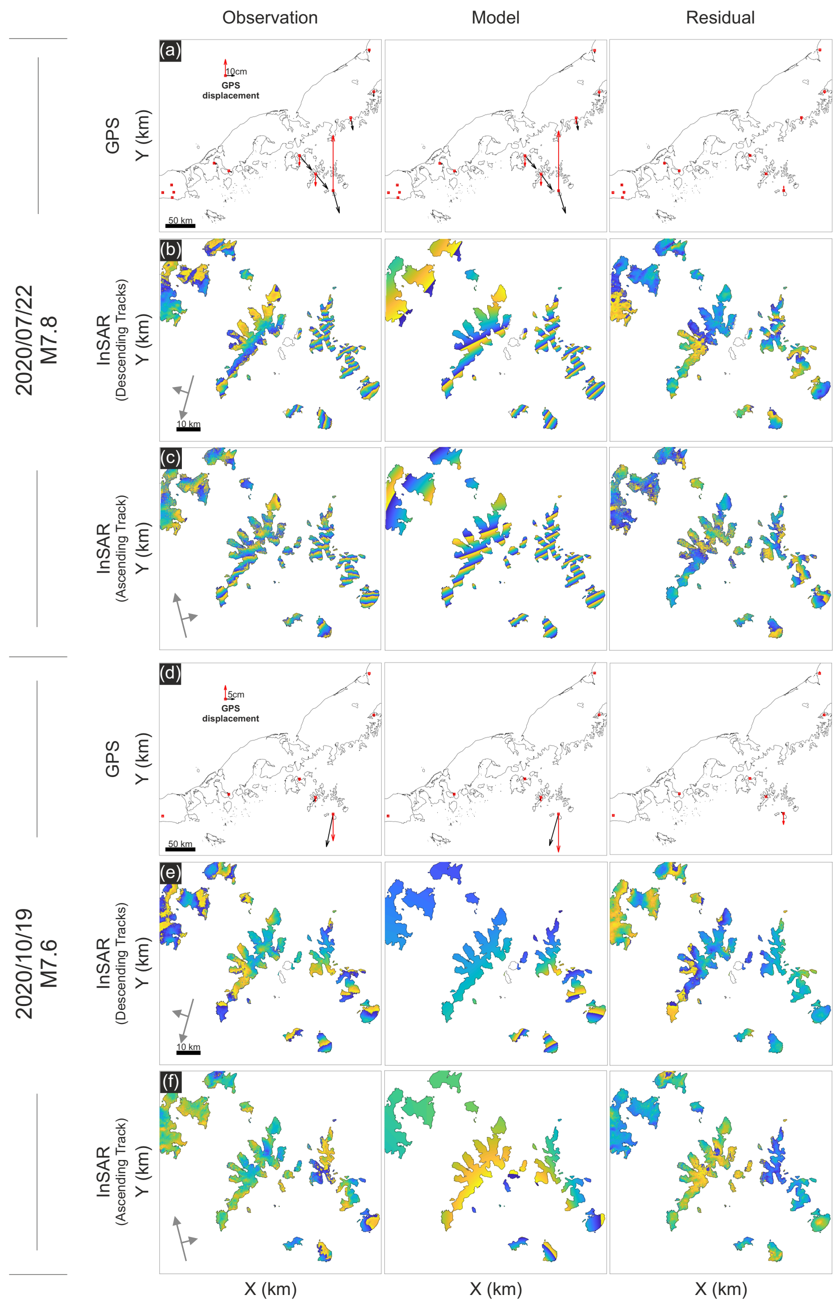

Figure 3: The observed and modelled GNSS displacements and wrapped interferometric phase. Images in the left column present the GNSS observations and the observed wrapped phase for the interferograms along 2 descending tracks, as shown in the dotted-dashed box in Fig. 1(c)-(f). Images in the middle column are the modelled GNSS and wrapped phase based on the optimal slip distributions. Images in the right column are the residual between observations and model.

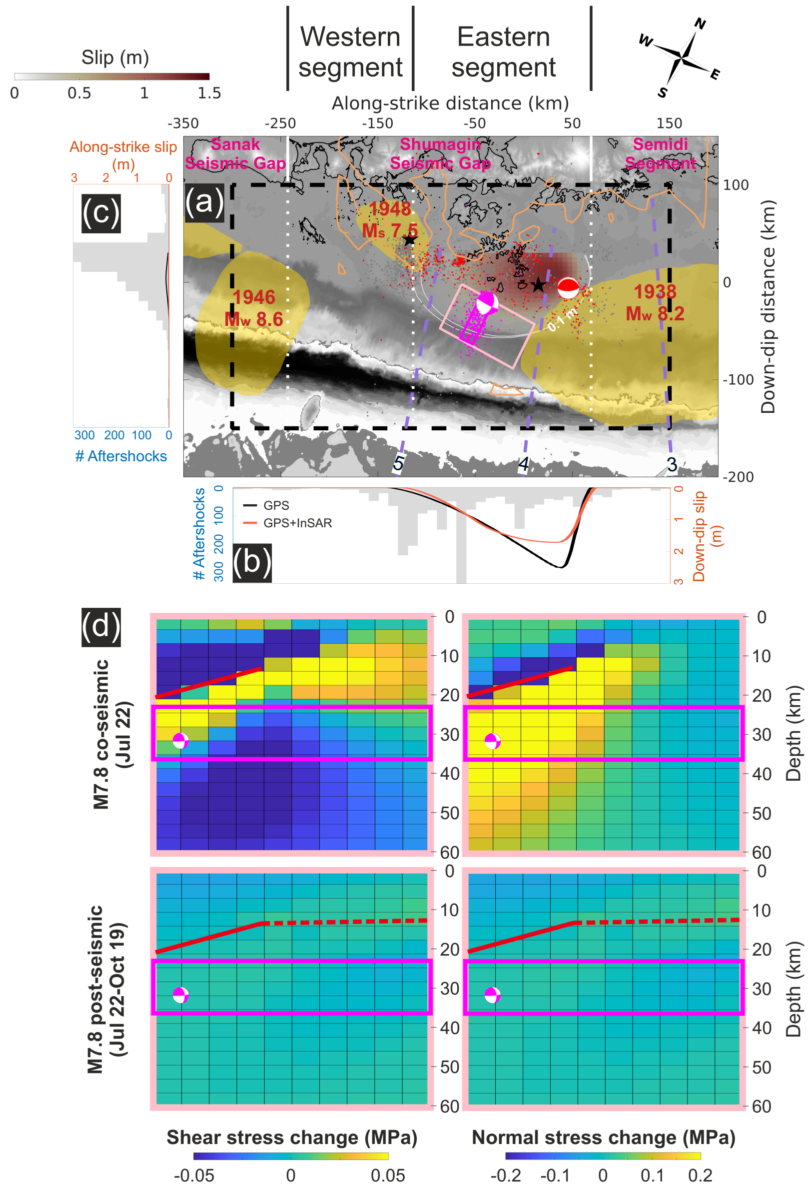

Figure 4: 2020 Shumagin earthquake doublet inferred fault slip and aftershock distribution. Panel (a) presents the coseismic slip distribution estimated from GNSS offsets and interferograms. The map is rotated so the along-strike slip is parallel to the x-axis and along-dip slip to the y-axis. White thin lines correspond to the 0.5m slip contours for minimum and maximum acceptable model parameters illustrating the uncertainties in the estimated coseismic rupture models of the Jul 22 2020 M7.8 event, and white thick lines

Figure 4 (continued): are the 0.25m slip contours for the postseismic slip models of this event (Jul 22- Oct 19 2020). Dashed white lines divide the Shumagin gap into two segments, discussed in the main text. Dashed purple lines mark seismic reflection lines 3-5 in Li et al. (2015) and Shillington et al. (2015), located in the Semidi Segment and eastern and western segments of Shumagin gap. The M7.8 and the M7.6 events are plotted as red and magenta beachballs. The first 2-day and 3-month aftershocks following the M7.8 event are plotted as red and gray dots, where two M6+ events are plotted as little black stars. The first 2-day aftershocks following the M7.6 event are plotted as magenta dots. Magenta rectangle outlines the ensemble of inverse Bayesian fault geometry model for the M7.6 earthquake, and the pink rectangle is the extended model from the surface (depth=0) to depth=60 km. Panels (b) and (c) show the slip and aftershock distributions in the along-strike and along-dip directions, with gray bars showing the number of aftershocks in 10 km-wide intervals. Black and orange lines show slip profiles from GNSS only and GNSS and InSAR data inversions, respectively. Image (d) presents the stress change on the extended geometry of the M7.6 fault model, caused by the M7.8 coseismic and postseismic slip distributions shown in (a). Solid red lines from 14-20 km present the intersection of two fault planes.

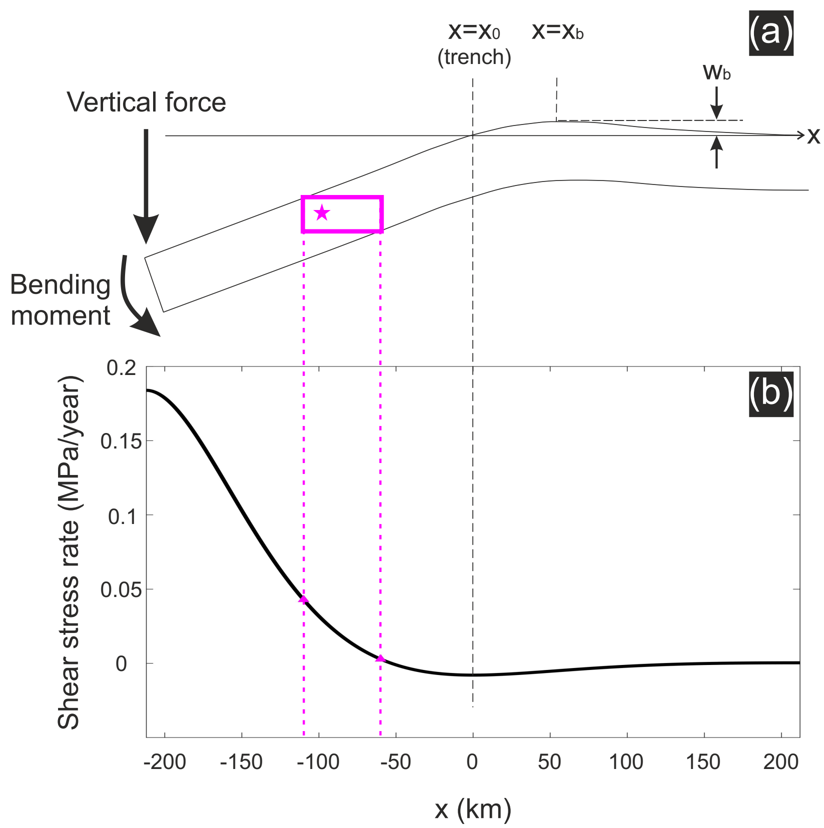

Figure 5: Bending of the lithosphere at an ocean trench due to an applied vertical load and bending moment. Image (a) is the conceptual model lithosphere bending, which is modified from Fig. 3.33 in Turcotte and Schubert (2014). Image (b) is the estimated shear stress rate along the trenching-normal profile.

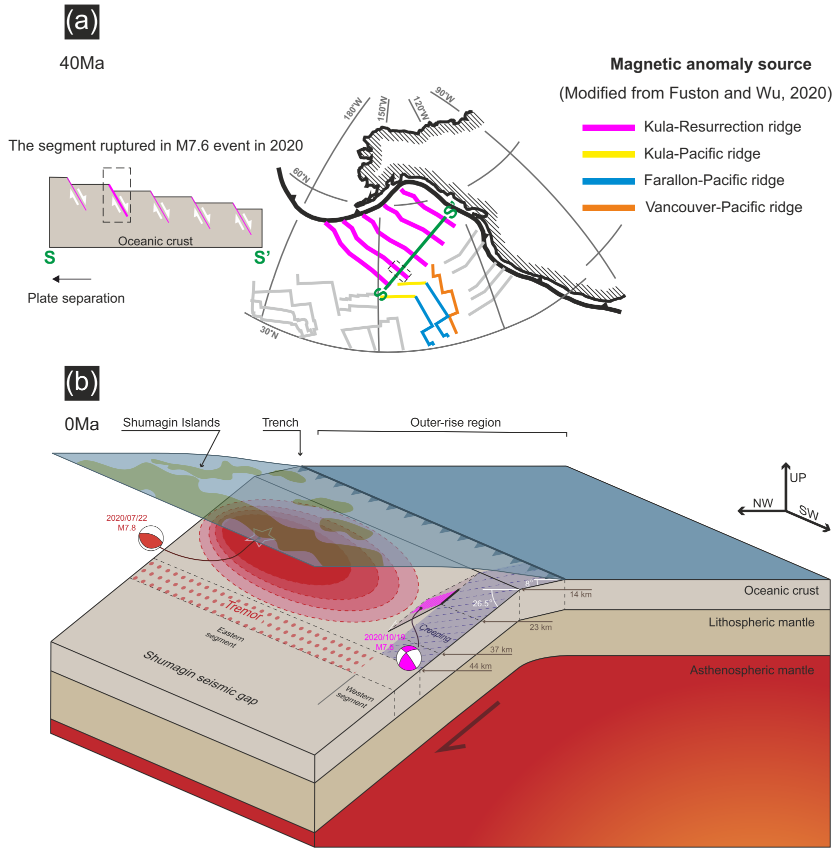

Figure 6: Conceptual model of the main subduction zone characteristics where the megathrust earthquake occurs in the plate interface and the triggered strike-slip earthquake tears the incoming oceanic lithosphere. Image (a) shows the slab plate tectonic reconstruction of western North America at 40Ma, modified from Fuston and Wu (2020). The magenta lines indicated the magnetic anomalies caused by the Kula-Resurrection ridge, which was subducted beneath the Aleutian Islands in the present days. A cross-section of this series of magnetic anomalies are plotted in profile SS'. We propose that a normal fault associated with this ridge system was reactivated in the Oct 19 2020 M7.6 earthquake. Yellow, blue and orange lines present the magnetic anomalies caused by Kula-Pacific ridge, Farallon-Pacific ridge, and Vancouver-Pacific ridge. This Pacific-Farallon-Kula triple junction moved to the north with the subducting Pacific plate and is located in the outer-rise region close to Shumagin Islands now. In image (b), the red shaded region is the rupture area at depth 14-44 km caused by the Jul 22 2020 M7.8 event. The magenta shaded region

Figure 6 (continued): is the rupture area caused by the Oct 19 2020 M7.6 event. The latter fault might be a reactivated pre-existing fault before subducting, caused by the seafloor spreading of Kula-Resurrection ridge, as shown in image (a).

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