Coseismic Rupture Geometry and Slip Rupture Process During the 2018 Mw 7.1 Anchorage, South-Central Alaska Earthquake: Intraplate Normal Faulting by Slab Tear Constrained by Geodetic and Teleseismic Data

Ping He1, Yangmao Wen2, Yunguo Chen3, Caijun Xu2, and Kaihua Ding4

1Hubei Subsurface Multi-Scale Imaging Key Laboratory, Institute of Geophysics and Geomatics, China University of Geosciences, Wuhan, China, 2School of Geodesy and Geomatics, Wuhan University, Wuhan, Hubei, China, 3School of Earth and Space Sciences, University of Science and Technology of China, Hefei, China, 4Faculty of Information Engineering, China University of Geosciences, Wuhan, China

Abstract The Mw 7.1 Anchorage earthquake on 30 November 2018 beneath the south-central Alaska is a rare intermediate-depth event larger than Mw 7 that occurred in a complex subduction region, where the young Yakutat oceanic terrane wedges in the continental-oceanic plate collisional region between the Pacific oceanic plate and the North American plate. We use both ascending and descending Sentinel-1 satellite Interferometric Synthetic Aperture Radar (InSAR) images to construct the coseismic displacement associated with this earthquake, which shows a nearly circular deformation pattern with a subsidence of ~4 cm in line of sight direction. Combining coseismic GPS data, we determine the focal mechanism of this event dominated by normal faulting with N-S striking of 186° and westward dipping of 64° by using a uniform slip model. Then we find a preferred slip model with both geodetic data and teleseismic data, suggesting the main slips are concentrated on a depth of 55–75 km. The total released moment of our preferred slip model is 5.32 × 1019 N·m, equivalent to Mw 7.1. The rupture process includes two peaks terminating at about 18 s and indicates a unilateral rupture with its front propagating northwestward direction at an average speed of 2.5 km/s. In comparison with the detailed seismic image in this region, this event just occurred in the Yakutat terrane beneath a low velocity zone, suggesting it was caused by slab tear but not tear boundary breaking and determining the lower boundary of shallow thrust-slip in the Alaska subduction zone.

1. Introduction

Subduction zones account for about 90% of energy released by global historical earthquakes, in particular, of the world’s largest earthquakes (magnitude 9+) and have been deemed to be the home of most seismically active faults on the planet (Hayes et al., 2018). Previous studies demonstrate that the subduction geometry controls most tectonic activities within subducting slabs, including not only the distribution and fragmentation of Earth’s tectonic plates (Maksymowicz, 2015; Mallard et al., 2016) but also the spatial extent, size, and type of associated earthquake (Sarlis et al., 2018; Shillington et al., 2015; Ye et al., 2017). Up to date, seismic image and focal mechanism of large earthquakes have become the two main ways to determine the subduction structure. Note that the seismic image with fine resolution greatly depends on dense station networks, which would greatly increase the economic cost and difficulty of data analysis, resulting in the seismic image usually being analyzed along just one line transect (Eberhart-Phillips et al., 2006). In addition, the robustness of seismic image observation decreases with depth, in particular, when it passes through a low-velocity zone (Kim et al., 2018). With the improvement of Global Seismic Network (GSN) and space geodesy (i.e., Global Positioning System [GPS] and Interferometric Synthetic Aperture Radar [InSAR]) data, the determination of focal mechanism has become easy and a routine work in seismic research (Ye et al., 2016). The proposed Slab2 model, based on focal mechanism, exhibits the link between seismicity, focal depth, and subduction structure worldwide (Hayes et al., 2018). Therefore, a reliable focal mechanism of large earthquake can contribute to understanding the knowledge of subduction zone geometry and predicting the accompanying disaster (e.g., tsunami) (Hayes et al., 2018).
Many major earthquakes/slip deficits associated with occasional thrust faulting have been observed in the subduction zones no deeper than 30–50 km (Lay et al., 2018; Tape et al., 2018). In shallow megathrust interface of subduction zones, they yield to typical intervals of strain accumulation and release driven by plate converge and are capable of >Mw 9 destructive ruptures (e.g., Wetzler et al., 2017; Hayes et al., 2018). Numerous studies related to seismic gap, earthquake recurrence period, and earthquake nucleation process have been carried out for fault slip behavior in these subduction regions (e.g., Freymueller & Beavan, 1999; Li & Freymueller, 2018; Tape et al., 2018). The knowledge of subduction structure in shallow regions is to be well understood by high seismicity and recurrent characteristic earthquake (e.g., He et al., 2017). In comparison with the force driven by plate converge in shallow depth, normal faulting earthquakes in intermediate depth are mainly caused by plate bending and then the denser oceanic lithosphere pulled downward by gravity (e.g., Davies & Richards, 1992; Schellart & Rawlinson, 2013). The potential energy of slab pulling is very important to the subduction recycled back into the Earth’s mantle (Schellart & Rawlinson, 2013). The seismicity and deformation associated with deeper zone along the slab interface are usually relatively small in magnitude or are in the oceanic region, resulting in most focal mechanisms of normal faulting earthquakes in intermediate depth being determined only from teleseismic data (e.g., Sarlis et al., 2018; Ye et al., 2017), which has lower sensitivity to the focal geometry, in particular of focal depth than geodetic data (Christensen & Ruff, 1985; Okuwaki & Yagi, 2017; Silwal & Tape, 2016). Therefore, the subduction structure in intermediate depth is not known as well as in shallow depth. However, the location and geometry of intermediate-depth normal faulting earthquakes are very important to constrain the maximum nucleation/locked width of thrust plate interface and understand the rheology of lower crust and upper mantle (Li & Freymueller, 2018; Ye et al., 2017).

On 30 November 2018, at 17:29 UTC (8:29 a.m. local time), a Mw 7.1 earthquake struck the south-central Alaska (USA), as shown in Figure 1. The mainshock was located about 16 km north of Anchorage, which is the most populous city in Alaska and has been subjected to regular earthquakes, including the great 1964 Mw 9.2 Alaska earthquake. It was followed by a maximum aftershock with magnitude of 5.7 just 6 minutes later. The 2018 event, as the largest one since the great 1964 event, caused significant damage to some buildings, bridges, pipes, roads, and highways; fortunately, no casualties were reported (Alaska Earthquake Center). The south-central Alaska experiences a complex tectonic activity, including both north-westward motion of 50.9 mm/yr and 50.3 mm/yr from the Pacific plate and the suboceanic Yakutat terrane beneath the North American plate, respectively (Bird, 2003; Elliott et al., 2010). The primary focal mechanisms from both U.S. Geological Survey (USGS) and Global Centroid Moment Tensor (GCMT) agree this event occurred as the result of normal faulting beneath a depth of about 40 km, suggesting it ruptured on an intraslab fault rather than on the shallower thrust-faulting interface between these three plates. After this event, the coseismic displacements derived from 96 GPS sites can be accessed from the Nevada Geodetic Laboratory (NGL) website (http://geodesy.unr.edu/index.php). Regardless of the severe damages, the 2018 event provides a rare chance for an insight into the subduction structure in intermediate depth.

Space geodetic data (InSAR and GPS) have been widely used to investigate the coseismic ground deformation due to earthquakes, which allow a complete overview for the coseismic displacement field and provide valuable constraints to determine the fault geometry and slip distribution. In contrast, the teleseismic waveform data have low sensitivity to slip fault geometry but have high time resolution, suggesting better details of the earthquake slip heterogeneity, that is, slip temporal evolution. In addition, geodetic inversion of fault slip can be oversmoothed (particularly for small, deep events) such that they are too smooth to explain the teleseismic data (Pritchard et al., 2006). Therefore, the complimentary nature of geodetic and teleseismic data has been simultaneously considered in most recent publications (e.g., Liu et al., 2018, 2019; Yue, 2014), resulting in tight constraint for the slip distribution. However, given that the teleseismic data and geodetic data have different sensitivities to coseismic rupture properties (Pritchard et al., 2006, 2007), a mature strategy to determine the reliable focal mechanism is fault geometry inverted by uniform model with geodetic data first, then the spatial distribution of fault slip tightly constrained by integrating teleseismic and geodetic data (Ding et al., 2018). In this study, we image the coseismic displacement associated with this 2018 event using C-band Sentinel-1 satellite InSAR measurement. Combined with GPS data, we invert the fault geometry with a uniform model in elastic half space (Okada, 1992), then estimate the slip rupture with both teleseismic and geodetic data, and finally discuss the intraslab tectonic and related seismic behavior.
2. Tectonic Setting

South-central Alaska is located at the northeastern elbow (also known as “corner geometry”) of the Pacific plate, where the Pacific plate is subducting beneath the North American plate (Bird, 2003; Eberhart-Phillips et al., 2006). As a famous subduction zone worldwide, it is predominantly controlled by continental-oceanic plate collisional interactions (Fuis et al., 2008; Koons et al., 2010). The situation has become particularly complex with the accretion of the Yakutat terrane, which is composed of anomalously thick and buoyant oceanic crust (Gomberg & Prejean, 2013). The presence of the Yakutat terrane flattens the dip angle of subduction slab with <3°, resulting in exceptionally wide seismogenic and transition zones (Freymüller et al., 2000; Peterson & Christensen, 2009), as shown in Slab 2 model (Hayes et al., 2018) (Figure 1). Relative to the North American plate, the corner region undergoing fast convergence, revealed by present GPS observations, which suggests that the Pacific plate and Yakutat terrane move northwestward with a normal dip at 50.9 and 50.3 mm/yr, respectively (Bird, 2003; Elliott et al., 2010). Seismic image exhibits that the ongoing subduction initiates from north of Alaska-Aleutian trench and then propagates south of the Denali volcanic gap where a paucity of volcanism is observed (Martin-Short et al., 2016). From the Mesozoic to the present,
multiple stages of terrace accretion, arc magmatism, and inboard propagation of crustal deformation are driven by these plate boundary interaction dynamics (Jiang et al., 2018).

Near Anchorage, two slow slip event (SSE) regions have been reported in Upper Cook Inlet and Lower Cook Inlet (respectively known as SSE-1 and SSE-2 hereinafter) (Fu et al., 2015; Ohta et al., 2006; Wei et al., 2012). Although SSE impedes the buildup of large slip deficit, it possibly triggers large shallow megathrust earthquakes (Fu et al., 2015). In addition, a nonvolcanic tremor, which is typically known to occur during episodes of SSEs and located on fault segments downdip of the seismogenic zone (Li & Ghosh, 2017), was observed to accompany with the SSE in the Upper Cook Inlet (Fu et al., 2015). The SSE1 and associated tremor in Anchorage occurred at depth of 25–30 km on top of the Yakuta terrane, and low velocity zone (LVZ) may extend to greater depths (Wech, 2016). As lacking of certain plate configuration beneath this region, two possible explanations have been proposed to address the geodynamic mechanism of these complicated tectonic activities in Anchorage: one is the strength and frictional properties change for the medium in the plate interface, and the alternative is a “slab tear” exists in part of the subducting Pacific plate (Gomberg & Prejean, 2013; Peterson & Christensen, 2009). From both the USGS and GCMT earthquake catalogs, numerous earthquakes that occurred in this subduction zone have been recorded, and the anchorage is famous as one of the most seismically active regions in the world (Li & Ghosh, 2017) (Figure 1). Along the direction of the incoming oceanic plate, major and great earthquakes occurred intraslab, including all three different types, that is, oceanic lithosphere strike-slip earthquakes near the subduction zone (e.g., 2018 Mw 7.9 Gulf of Alaska earthquake) (e.g., Krabbenhoeft et al., 2018), megathrust earthquakes in the shallow subduction (e.g., 1964 Mw 9.2 Alaska earthquake) (e.g., Plafker et al., 1994), and this 2018 Mw 7.1 normal earthquakes in intermediate depth. Although the first two types of earthquakes are larger in magnitude and are more frequent than the last one, they occur out at sea with shallow depth and contribute little to understanding the subduction structure in deep depth. Therefore, a reliable fault geometry and slip distribution of this large intermediate-depth normal earthquake is important for us to completely understand the structure and tectonic evolution in this subduction zone.

3. Data Analysis

3.1. InSAR Data

The recent InSAR measurement, advanced in high spatial resolution and large-scale coverage, which overcome the drawback of time- and labor-intensive in situ ground-based observation, has become a routine response in earthquake investigations (He et al., 2019). Present InSAR can provide high-quality coseismic displacement caused not only by large and moderate earthquakes but also by small earthquakes, for example, the 2011 Mw 5.3 Trinidad earthquake (Barnhart et al., 2014), the 2010 Mw 5.0 Ecuadorian Andes earthquake (Champenois et al., 2017), and the 2016 Mw 6.0 Australia earthquake (Polcari et al., 2018), suggesting it should be available for this Mw 7.1 earthquake, even though the earthquake is in intermediate depth. We collect SAR data from the Sentinel-1 radar satellite constellation, which is operated by the Europe Space Agency (ESA) and launched on 3 April 2014. As the third-generation C-band (i.e., wavelength of 5.6 cm) SAR satellite from ESA, its mature constellation guarantees a shorter temporal and spatial baseline for each InSAR pair to keep high coherence, and its new imaging mode improves the ground coverage size of each image and retains a comparative pixel resolution as images from the last two generations of SAR satellites (i.e., ERS-1/2 and Envisat). And most of all, these data are completely open and sharable. As listed in Table 1, two pairs of InSAR data have been processed in this study.

The InSAR pairs in Table 1 with short temporal and spatial baseline were processed with the mature two-pass Differential InSAR (DiInSAR) method by using the commercial GAMMA software. The phase contribution of topography was removed along with the global 30-m digital elevation model (DEM) from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER), which covers land surfaces between 83° south latitude and 83° north latitude, a larger coverage than that of the Shuttle Radar Topography Mission (SRTM) (Tachikawa et al., 2011). The common processing flow for interferogram includes filter, unwrap, and geocode. In addition, a polynomial model was used to consider all possible long wavelength errors (e.g., orbit error or some atmospheric disturbance), and the final interferograms present each pixel displacement from its ground position toward the satellite (Line-Of-Sight, LOS) as shown in Figures 2a and 2b. Due to the southwestern Cook Inlet and northeastern rolling hills, some atmospheric...
disturbances can be found near the mountain region to the northeast and reduce the coherence in Track 065. Fortunately, the major displacement concentrated northwest of Anchorage is clearly visible with a pattern of disk in both interferograms and a maximum coseismic displacement of 4–5 cm, corresponding to a normal faulting earthquake with deeper depth. To analyze the signal-to-noise ratio (SNR) of our interferograms, a 1-D covariance function was used to estimate the uncertainty characteristics of each interferograms by calculating the autocorrelation between every pair of pixels at a given distance in the nondeforming areas (Parsons et al., 2005). The standard deviations were 11 and 14 mm for the T0131 and T065D, respectively, indicating a medium random noise level, which may be attributed to the atmospheric disturbance. Finally, we implement a subsample for the interferograms by using the Quadtree algorithm (Jónsson et al., 2002) to reduce the computational burden of inversion of these interferograms.

### 3.2. GPS Data

GPS, an alternative important space geodetic technique, has been commonly used to observe surface deformation with millimeter precision in three dimensions (3-D), which plays a comparable role in

| Track (A/D) | Master YYYYMMDD | Slave YYYYMMDD | Perp. B m | Inc. Angle ° | Azi. Angle ° | σ mm | α km |
|-------------|-----------------|----------------|-----------|--------------|-------------|------|-----|
| 131 (D)     | 20181122        | 20181204       | 11        | 30–44        | −12         | 11   | 10.3|
| 065 (A)     | 20181031        | 20181206       | 144       | 30–44        | −168        | 14   | 8.2 |

Note. Data download from https://vertex.daac.asf.alaska.edu/. σ is the standard deviation of the line-of-sight (LOS) measurements in the nondeforming regions of the interferograms and α is the e-folding correlation length scale of the 1-D covariance functions of the interferograms.

![Figure 2](https://example.com/figure2.png)

**Figure 2.** InSAR coseismic displacements from Sentinel-1 images for (a) ascending track T131 and (b) ascending track T065. The corresponding synthetic interferograms on (c) and (d) of the preferred uniform model and their residual interferograms in (e) and (f), respectively. In panel (a–b), the blue dashed rectangular frame indicates the main deformation region, and the outside region can be seen as nondeforming region which are used to estimate the quantity of the interferograms. In panel (c–d), the black dashed rectangular denotes the fault geometry determined by preferred uniform model, and the red line indicates the seismological fault of this Anchorage earthquake on the surface. The fault geometry is also used hereinafter in Figures 3 and 8.
Earth and Space Science 10.1029/2019EA000924

earthquake investigations. As mentioned above, InSAR is advanced in high spatial resolution, but only in one dimension (look direction) and with low sensitivity to the horizontal component, particularly in N-S direction. In addition, a small magnitude of surface displacement compared to noise in InSAR data is prone to lead an unstable inversion problem, suggesting significant trade-off between fault plane geometry and distributed slip (e.g., Barnhart et al., 2014; Scott et al., 2014). The complimentary nature of InSAR and GPS data allows us to view a complete and reliable coseismic displacement field caused by the mainshock and then tightly constrain for the fault geometry. Thanks to the Plate Boundary Observatory (PBO) network, a dense continuous GPS sites can be available in this region, and its coseismic displacement measurements are published on the NGL website.

For the GPS coseismic displacement from NGL, as shown in Figures 3a and 3b, the maximum horizontal and vertical displacements are about 2 and 3 cm near the Anchorage, respectively, corresponding to the precision of 1 and 3 mm. As shown in Figure 3a, the displacement in the east-west component exhibits a further extension in the east than in the west; that ~1-cm displacement has been found in sites AC14 and AC11 with a distance of ~200 km from the Anchorage. In the north-south component, it shows a contraction that sites north of the epicenter move south and south of the epicenter move north. Most sites near the epicenter in vertical component (Figure 3b) show a subsidence displacement, except the two sites AC20 and AC15. Compared with the InSAR coseismic deformation (Figures 2a and 2b), a similar deformation pattern is revealed by GPS data. In addition, the discrepancies between GPS sites and the corresponding InSAR pixel data are of ~1-cm accuracy near the epicenter, suggesting a good agreement between the two data sets and no systematic error in the interferogram. It is worth noting that the good agreement between GPS and InSAR data attests that the InSAR data are mainly in response to coseismic slip rupture, rather than any large-scale atmospheric disturbance. In the

Figure 3. Observed coseismic GPS displacements with green arrows in horizontal (a) and vertical component (b) (http://geodesy.unr.edu/index.php). The blue arrows in (a) and (b) indicate the synthetic coseismic displacement of the preferred uniform model, and their residuals are with yellow arrows in (c) and (d), respectively.
following, both GPS and InSAR data are used to constrain the fault geometry and slip distribution for the earthquake.

3.3. Telesismic Data

Telesismic waveform data provide good constraints on the temporal process of slip heterogeneity and the total released seismic moment and can avoid the oversmoothing of the slip model estimated by geodetic inversion (Pritchard et al., 2006); as a result, it has become an important complementary data together with static geodetic data to estimate finite-fault slip distribution in present studies (e.g., Ji et al., 2002; Yue, 2014; Liu et al., 2019). Therefore, we collect telesismic waveform data recorded by the global seismological network (GSN) stations from the Incorporated Research Institutions for Seismology (IRIS) Data Management Centre. We analyze 52 broadband seismic records involving 32 P and 20 SH waveforms from the GSN at epicentral distances between 30° and 90°, and their location exhibits uniform azimuth distribution shown in Figure 4. We use records extending 60 s after the arrival of each phase. In the telesismic data processing, we obtain displacement records by deconvolving the station response of each waveform data and integrating them. All data are then resampled by band-pass filter with frequency between 0.0033 and 1.0 Hz. The processed telesismic waveforms with high SNR shown in Figure 4 will be used for the slip distribution inversion in Section 4.

4. Source Model

A finite-fault dislocation model, which describes the ground deformation in response to subsurface slip on a single rectangular plane with nine parameters (i.e., rupture plane location (longitude and latitude), length,
width, depth, strike, dip, rake, and slip), has been commonly used to construct the focal mechanism associated with earthquakes in elastic half-space (Okada, 1992). We invert the source model with two steps in this study. In the first step, we use a uniform slip model to invert the focal mechanism with geodetic data, in order to determine robust fault geometry. To resolve the nonlinear inversion in step one, we adopt a simplex search algorithm (Clarke et al., 1997) to find a global minimum solution, corresponding to each initial model. We use 100 initial models randomly sampled based on Gaussian normal distribution to generate 100 solutions and then chose the final solution with a statistic analysis, and detailed descriptions are as in previous studies (e.g., He et al., 2017; Parsons et al., 2005). As data sets are of the same type, the relative weight between GPS and InSAR data is determined by their errors. In the second step, we only invert the slip distribution by fixing the fault geometry derived from step one. Integrating both teleseismic and geodetic data, we solve the slip following a kinematic inversion finite-fault strategy (Ji et al., 2002) with simulated annealing method (Sen & Stoffa, 1991), and we weight them equally (Liu et al., 2019).

### 4.1. Fault Geometry Determination

As shown in Table 2, the preferred model is a fault plane with a strike of 186.0°, a dip of 64.0° to the north-east, a rake of −88°, and other parameters, implying this earthquake broke presumably a highly west-dipping normal fault, with nearly north-south striking. Our model rules out the other alternative nodal plane solution from USGS or GCMT. In addition, the fault plane located between 42.8 and 67.4 km is much deeper than that in the preliminary result from neither USGS nor GCMT, corresponding to an intermediate-depth event. The uncertainties of these parameters are evaluated by a Monte Carlo bootstrap simulation technique, illustrated in Figure 5. In comparison with the uncertainty of other parameters, the fault length and slip magnitude in uniform model exhibit a large variation range, suggesting significant trade-off for their uncertainty range. Therefore, the uniform model with geodetic data suggests a tight constraint for the fault geometry, though a relatively loose constraint for the slip parameter. The total geodetic moment of uniform model is 3.2 × 10^{19} N·m, which is slightly smaller than 4.8 × 10^{19} N·m estimated by the GCMT catalog. Our best-fitting solution for a uniform planar fault is illustrated in Figures 2c and 2d and Figures 3a and 3b, and their residual is shown in Figures 2e and 2f and Figures 3c and 3d, respectively. Both the model pattern and their residual displacements show our preferred fault geometry can describe the geodetic data well, which features normal faulting for this event.

### 4.2. Slip Rupture

With the determined fault geometry, we invert the coseismic slip distribution and seismic moment on the planar fault constrained by both geodetic and teleseismic waveform data. The fault plane is extended with a size of 57 km × 48 km to avoid edge effects and then divided into small patches with size of 3 km × 3 km. In each patch, the dislocation amplitude can vary from 0–300 cm, the rake angle can vary from 250–290, the average rupture velocity allows to vary from 2.0 to 3.5 km/s, and rise time from 0.6 to 6.6 s at 0.6-s intervals, respectively. To trade off both the geodetic and teleseismic data, the preferred epicenter depth of 60 km is used. With the same configurations, we perform three inversions from geodetic data, teleseismic data, and their joint, respectively, in order to determine slip distribution for the Anchorage event.

### Table 2

| Model     | Lon.  | Lat.  | Strike | Dip  | Rake | Depth | MinDepth | MaxDepth | Length | Slip | Moment |
|-----------|-------|-------|--------|------|------|-------|----------|----------|--------|------|--------|
| USGS      | −149.955 | 61.346 | 189    | 62   | −88  | 46.7  | —        | —        | 4.703  | 7.05 |
| GCMT      | −150.02 | 61.49  | 189    | 64   | −89  | 48.2  | —        | —        | 4.8   | 7.1 |
| Uniform   | −149.48 | 61.40  | 186.0  | 64.0 | −88.0| 42.8  | 67.4      | 29.5     | 1.2    | 3.2  | 6.94   |
| Distributed | −149.768 | 61.3401 | 186   | 64   | −60  | —     | 57        | —        | 5.3    | 7.1 |

HE ET AL. 8 of 15
As shown in Figures 6a and 6b, the slip distributions from only geodetic or teleseismic data are different in slip pattern, magnitude, depth, and roughness, except their main slip rupture locations are close, suggesting a different detection ability between the two types of data, in particular of this intermediate-depth event. To consider each data set's contribution, our final joint inversion is shown in Figure 6c. The rupture zone over 1 m is mainly concentrated at depth between 55 and 70 km, with a peak value of 2.8 m. The rupture length is up to 50 km, and most slips occurred on upper depth to subsurface. The total seismic moment of the joint inversion is $5.32 \times 10^{19}$ N·m, equivalent to a magnitude $M_w$ 7.1. The model predictions show good agreement between the observations and synthetic data (the synthetic seismic waveforms is illustrated in Figure 4, and the synthetic geodetic displacement can be seen in Figures S1 and S2 in the Supporting Information). To illustrate the degree to which these geodetic data are able to resolve the slip model for such a deep-epicenter event, we do the checkerboard test (Figure S3). The tests suggest our data can constrain the slip pattern in different depths well, while the resolution of peak slip decreases with the epicentral depth. With the constraint of teleseismic data, the kinematic slip rupture process as well as static slip distribution has been revealed. The total released moment terminates at about 18 s with two peaks as shown in Figure 6d. In the first 6 s, the rupture produces the pulse in moment rate function, and its accumulated seismic moment

Figure 5. The preferred parameters and their uncertainty estimation with uniform model based on the Monte Carlo method.
is $2.5 \times 10^{19}$ N·m, which is half of the total released moment (i.e., $5.32 \times 10^{19}$ N·m) and equivalent to an Mw 6.87 event. In the later 12 s, the second rupture pulse releases the other half moment. As shown in the snapshots of the joint rupture model in Figure 7, the energetic rupture propagates in downdip direction concentrated in the period of 2.0–5.0 s and then propagates in the northwestward direction.

### 5. Discussion

#### 5.1. Surface Deformation Associated With Intermediate-Depth Earthquake

Surface deformation associated with earthquakes observed by advanced geodetic data, either GPS or InSAR, provides significant information to understand the relevant focal mechanism, seismic cycle, and physical process of lithospheric evolution (e.g., Grandin et al., 2016; Li & Freymueller, 2018; Diao et al., 2019). Note that most of these investigations have been implemented for large or shallow deformation but rarely for small or deep-depth events; as a result, surface displacement in response to subsurface activities sharply attenuates with depth. However, the seismicity in subduction zones, involving thrust earthquake from plate converge at shallow depth and normal faulting from extensional stresses at deep depth, can extend to the mantle margin along the downdip slab (Hayes et al., 2018). As yet, the slab earthquakes with normal faulting have been rarely recorded by geodetic data, because the local deformation associated with deep-depth events are too small to be captured by sparse GPS sites or InSAR measurement with centimeter precision. With the development of highly dense GPS network and mature-design SAR satellite constellation, the capacity of geodetic data has improved and has attempted to constrain some deep-depth events, for example, the 2013 Okhotsk Sea earthquake at depth of 70 km with 1.5- to 2-cm onshore coseismic offsets from GPS (Xu et al., 2017) and the Mw 6.5 Botswana earthquake at depth of 29 km with 4-cm LOS displacement from InSAR (Gardonio et al., 2018). In this study, both ascending and descending InSAR interferograms from Sentinel-1 SAR data reveal 4- to 5-cm LOS displacements caused by this Anchorage earthquake with high coherence (Figure 2). In addition, GPS shows horizontal and vertical displacements are about 2 and 3 cm near the Anchorage from NGL (Figure 3) and supports our InSAR data with high-quality 1-cm precision level. To some extent, this case is the first intermediate-depth slab event observed by complete geodetic

![Figure 6. Slip distribution and seismic rupture history. (a) The coseismic slip distribution from the geodetic data (a), teleseismic data (b), and their joint (c). (d) The moment rate release from finite source inversion by teleseismic-only (black) and joint inversion (red).](image-url)
data, rising a significant value that geodetic data can supplement the teleseismic data to constrain the intermediate-depth structure in subduction zones.

5.2. Slip Distribution Models from the Geodetic, Teleseismic and Joint Data

Fault geometry determined by geodetic data suggests this event ruptures a normal fault with approximately N-S strike and high dip angle over a 30-km length, which is similar to the first nodal plane determined by teleseismic data and rules out the second one (Table 1). Our uniform slip model indicates the rupture ranges a wide depth trade-off varying from between 42.8 and 67.4 km, and its average depth of 55 km is larger than the focal mechanism from USGS and GCMT. However, the present focal mechanisms agree that this event occurred at intermediate depth. According to the tectonic setting, the rupture area should be in the Yakutat terrane and dominated by the tensile stresses oriented in the dip direction, opposite to the 1964 Mw 9.2 Alaska thrust earthquake. Note that the slip distribution in Figure 6a with geodetic data was centered at 68 km, which is larger than that in uniform model, even if we shallow the initial rupture point with 55-km depth. We speculate the main reason is that the uniform model with a uniform slip magnitude will reduce its sensitivity to slip depth.

The teleseismic and geodetic data indicate different slip patterns with their single inversion as shown in Figure 6, suggesting different roles in constraining fault slip (Pritchard et al., 2006). In comparison to geodetic-only inversion, the slip from teleseismic-only inversion is rough, extends a wider zone with two asperities, and locates at shallower depth. Our results show that the difference between slip models from geodetic-only or teleseismic-only inversion is large, which may be amplified by the deep depth of this intermediate-depth earthquake (Pritchard et al., 2006). The slip distribution from joint inversion shown in Figure 6c is a trade-off between the two types of data such that the depth of its main slip is close to geodetic-only inversion and slip pattern is close to teleseismic-only inversion. As shown in Figure 4, the synthetic teleseismic data from both teleseismic-only inversion and joint inversion can fit the observed teleseismic waveform well, suggesting the teleseismic data are not sensitive to the parameter of slip depth. Therefore, we consider the joint inversion as our final model. In a previous study, the slip model for this Anchorage earthquake discussed by Liu et al. (2019), who used teleseismic, strong-motion, and GPS data, indicates high consistency among each data set, and a shallower slip depth at 55 km was used as our teleseismic-only inversion. In their inversion strategy, they inverted the slip model with the fault geometry derived from the USGS fault nodal plane solution, and no InSAR data constraint. Note that the teleseismic,

![Figure 7. The 3-s snapshots of joint inversion model, and the dashed pink circle denotes the location of rupture front at end of time window if the rupture speed is 2.5 km/s. The blue star indicates the hypocenter selected by our finite-fault geometry (lon:-149.7680; lat: 61.3401; depth: 60 km).](image-url)
strong-motion, and GPS data are in situ measurements at each site, and their observations show low spatial relation between stations. In contrast, InSAR provides surface displacement with nearly continuous distribution, which might improve the sensitivity of focal mechanism.

5.3. Subduction Structure and Seismic Behaviors

A clear subduction structure can help us to understand the tectonic activities beneath the Anchorage. With the autocorrelation of local earthquake coda, an enhanced resolution of the subducting plate interface in the Central Alaska was proposed in Figure 8 (after Kim et al., 2018). Beneath the Anchorage region, the intracrustal thickness is ~30 km from both Slab2 (Hayes et al., 2018) and seismic image (Kim et al., 2018), in which the slow slip events and tectonic tremor have been observed by GPS and seismic data (e.g., Fu et al., 2015; Wech, 2016). Then an LVZ layer beneath the intracrustal slab at depth of 30–40 km is formed by the shallow subduction leads fluids at the slab interface in central Alaska akin to warm subduction zone and has become an accepted explanation for the origin of the above tremor (Chuang et al., 2017). In our preferred slip model, the main slip is concentrated at depth of 55–70 km beneath the LVZ, where dense local earthquakes have been found (Kim et al., 2018). Though the depth of the main slip is partly below the depth of Moho, the thermal structure model in southern Alaska subducting slab shows our main slip is located with an average temperature contour of 500 °C (Figure 8b) (Ponko & Peacock, 1995), obeying to the largest seismogenic depth with temperature of 600 °C (Jackson et al., 2008). Note that the seismicity of local earthquake experiences a remarkable increase imitated from the seismogenic fault of this event. In addition, the slab dip also steeply changes at the loci of this event. Prior investigation with earthquake hypocenters supports that the slab dip steeply change at ~70 km depth (Page et al., 1989). The slab dip change can easily cause buoyancy differences within the slab to generate a normal-slip event, which has been attested by this Anchorage earthquake. However, the average dip angle of our slip is 64°, which is too steep relative to the slab dip, and our slip cut the interface of Yakutat Moho. Therefore, we speculate this event was generated by a slab tear due to the breaking of the slab, but not slab bending, as the 2017 Mw 8.2 Chiapas earthquake (Ye et al., 2017).

Seismic behavior in subduction zone indicates a strong relationship related to subduction structure. As proposed by previous studies, the large-scale dip change of the slab has played an important role in controlling the local stress state, resulting in depth-varying rupture properties in subduction zones (Lay et al., 2012; Ye et al., 2017). That means we can delineate or verify the subduction structure change if we determine the focal mechanism of large historical earthquakes in different depths. During the westward underthrusting of the thickened Yakutat terrane adjacent to normal oceanic plate, the Yakutat slab dip transitions from a
shallow-dip near the coast to a near-vertical dip in the Denali volcanic field (Jiang et al., 2018). Correspondingly, the large historical earthquakes along the Alaska subduction interface yield to a general pattern that strike-slip earthquake (i.e., 2018 Mw 7.9 Gulf of Alaska earthquake) occurred outlier ocean plate, large shallow thrust-slip (purple region of 1964 Mw 9.2 Alaska) occurred in boundary between upper intracrustal and ocean plate, and normal slip in the deep ocean plate, as shown in Figure 9. The Anchorage earthquake sustains the results of seismic image from previous results (Eberhart-Phillips et al., 2006; Kim et al., 2018). On the other hand, earthquakes in subduction zones can be divided into intracrustal thrust event and intraoceanic strike-slip/normal event (Krabbenhoeft et al., 2018). This Anchorage earthquake determined by our slip model indicates it occurred in the Yakutat oceanic crust with normal faulting, consistent with the knowledge of stress accumulation in subduction zone.

Acknowledgments
We thank the editor Prof. Andrea Donnellan and two anonymous reviewers whose insightful comments improved the manuscript. We thank Dr. Doyeon Kim for providing seismic image data for Figures 8 and 9. We thank Dr. Chengli Liu for the discussion of seismic data inversion. The Sentinel-1A SAR data were downloaded from the Sentinel-1 Scientific Data Hub (https://scihub.copernicus.eu/dhus). The GPS displacement data are download from the Nevada Geodetic Laboratory (NGL) website (http://geodesy.unr.edu/index.php). The teleseismic waveform records are obtained from the Incorporated Research Institutions for Seismology’s (IRIS) website (https://ds.iris.edu). The figures were plotted using the open Generic Mapping Tools (GMT) written by Paul Wessel and Walter H. F. Smith. This work is supported by the National Natural Science Foundation of China (41774011, 41704005, 41861134009, and 41974004), the National Universities (CUGL180410) and the National Key Basic Research Development Program (973 Program) (No. 2013CB733304).

References
Barnhart, W. D., Benz, H. M., Hayes, G. P., Rubinstein, J. L., & Bergman, E. (2014). Seismological and geodetic constraints on the 2011 Mw 5.3 Trinidad, Colorado earthquake and induced deformation in the Raton Basin. Journal of Geophysical Research: Solid Earth, 119, 7923–7933. https://doi.org/10.1002/2014JB011227
Bird, P. (2003). An updated digital model of plate boundaries. Geochemistry, Geophysics, Geosystems, 4(3), 1–52. https://doi.org/10.1029/2001GC000252
Champenois, J., Baize, S., Vallée, M., Jomard, H., Alvarado, A., Espin, P., et al. (2017). Evidences of surface rupture associated with a low-magnitude (Mw 5.0) shallow earthquake in the Ecuadorian Andes. Journal of Geophysical Research: Solid Earth, 122, 8446–8458. https://doi.org/10.1002/2017JB013928
Christensen, D. H., & Ruff, L. J. (1985). Analysis of the trade-off between hypocentral depth and source time function. Bulletin of the Seismological Society of America, 75(6), 1637–1656.
Chuang, L., Bostock, M., Wech, A., & Plourde, A. (2017). Plateau subduction, intraslab seismicity, and the Denali (Alaska) volcanic gap. Geology, 45(7), 647–650. https://doi.org/10.1130/G38867.1
Clarke, P., Paradisios, D., Briole, P., England, P., Parsons, B., Billiris, H., & Ruegg, J. (1997). Geodetic investigation of the 13 May 1995 Kozani-Grevena (Greece) earthquake. Geophysical Research Letters, 24(6), 707–710. https://doi.org/10.1029/97GL00430

Figure 9. Schematic map of subduction structure and its seismic behavior beneath the south-central Alaska (after Krabbenhoeft et al., 2018; Kim et al., 2018).
Martin-Short, R., Allen, R. M., & Bastow, I. D. (2016). Subduction geometry beneath south central Alaska and its relationship to volcanism. *Geophysical Research Letters*, 43, 9509–9517. https://doi.org/10.1002/2016GL070580

Ohta, Y., Freymueller, J. T., HeeßelDöttler, S., & Suito, H. (2006). A large slow slip event and the depth of the seismogenic zone in the south central Alaska subduction zone. *Earth and Planetary Science Letters*, 247(1-2), 108–116. https://doi.org/10.1016/j.epsl.2006.05.013

Okada, Y. (1992). Internal deformation due to shear and tensile faults in a half-space. *Bulletin of the Seismological Society of America*, 82(2), 1018–1040.

Okuwaki, R., & Yagi, Y. (2017). Rupture process during the Mw 8.1 2017 Chiapas Mexico earthquake: Shallow intraplate normal faulting by slab bending. *Geophysical Research Letters*, 44. https://doi.org/10.1002/2017GL075956

Page, R. A., Stephens, C. D., & Lahr, J. C. (1989). Seismicity of the Wrangell and Aleutian Wadati-Benioff zones and the North American plate along the Trans-Alaska crustal transect, Chugach Mountains and Copper River basin, southern Alaska. *Journal of Geophysical Research*, 94(B11), 16,059–16,082. https://doi.org/10.1029/JB094iB11p16059

Parsons, B., Wright, T., Rowe, P., Andrews, J., Jackson, J., Walker, R., & Engdahl, E. R. (2005). The 1994 Seâlar earthquake. *Physics of the Earth and Planetary Interiors*, 164(1), 202–217. https://doi.org/10.1016/j.epsl.2005.02665.x

Petersen, C. L., & Christensen, D. H. (2009). Possible relationship between nonvolcanic tremor and the 1998–2001 slow slip event, south central Alaska. *Journal of Geophysical Research*, 114, B06302. https://doi.org/10.1029/2008JB006096

Plafker, G., Moore, J. C., & Winkler, G. R. (1994). Geology of the southern Alaska margin. In G. Plafker & H. C. Berg (Eds.), *The geology of Alaska (Geology of North America, Vol. G)* (pp. 389–449). Boulder, CO: Geology Society American. https://doi.org/10.1130/DNAG-GNA-GL389

Polcari, M., Albino, M., Atzori, S., Bignami, C., & Stramondo, S. (2018). The intraplate 2016 Mw 6.0 Australia earthquake studied by InSAR data. In *IGARSS 2016-2018 IEEE Geoscience and Remote Sensing Symposium*, 7263-7285. https://doi.org/10.1109/IGARSS.2018.8517542

Ponko, S. C., & Peacock, S. M. (1995). Thermal modeling of the southern Alaska subduction zone: Insight into the petrology of the subducting slab and overlying mantle wedge. *Journal of Geophysical Research: Solid Earth, 100*(B11), 22,117–22,128. https://doi.org/10.1029/95JB02506

Pritchard, M. E., Ji, C., & Simons, M. (2006). Distribution of slip from 11 Mw > 6 earthquakes in the northern Chile subduction zone. *Journal of Geophysical Research: Solid Earth, 111*, B0302. https://doi.org/10.1029/2005JB004013

Pritchard, M. E., Norabuena, E. O., Ji, C., Boroschek, R., Comte, D., Simons, M., et al. (2007). Geodetic, teleseismic, and strong motion constraints on slip from recent southern Peru subduction zone earthquakes. *Journal of Geophysical Research: Solid Earth, 112*, B03307. https://doi.org/10.1029/2006JB004294

Polcari, M., Albino, M., Atzori, S., Bignami, C., & Stramondo, S. (2018). The intraplate 2016 Mw 6.0 Australia earthquake studied by InSAR data. In *IGARSS 2016-2018 IEEE Geoscience and Remote Sensing Symposium*, 7263-7285. https://doi.org/10.1109/IGARSS.2018.8517542

Sen, M. K., & Stoffa, P. L. (1991). Nonlinear one-dimensional seismic waveform inversion using simulated annealing. *Geophysics*, 56(10), 1624–1638. https://doi.org/10.1190/1.1442973

Shillington, D. J., Beccé, A., Nedimović, M. R., Kuehn, H., Webb, S. C., Abers, G. A., et al. (2015). Link between plate fabric, hydration and subduction zone seismicity in Alaska. *Nature Geoscience*, 8(12), 961. https://doi.org/10.1038/ngeo2586

Silwal, V., & Tape, C. (2016). Seismic moment tensors and estimated uncertainties in southern Alaska. *Journal of Geophysical Research: Solid Earth, 121*, 2772–2797. https://doi.org/10.1002/2015JB012588

Tachikawa, T., Hato, M., Kaku, M., & Iwasaki, A. (2011). Characteristics of ASTER GDEM version 2. In *2011 IEEE International Geoscience and Remote Sensing Symposium*, 3657-3660. https://doi.org/10.1109/IGARSS.2011.6030017

Tape, C., Holtkamp, S., Silwal, V., Hawthorne, J., Kaneko, Y., Ampuero, J. P., et al. (2018). Earthquake nucleation and fault slip complexity in the lower crust of central Alaska. *Geophysical Journal International*, 170(7), 536–541. https://doi.org/10.1093/gji/ggy442

Wech, A. G. (2016). Extending Alaska's plate boundary: Tectonic tremor generated by Yakutat subduction. *Geology, 44*(7), 587–590. https://doi.org/10.1130/0191.1

Wei, M., McGuire, J. J., & Richardson, E. (2012). A slow slip event in the south central Alaska subduction zone and related seismicity anomaly. *Geophysical Research Letters*, 39, L15309. https://doi.org/10.1029/2012GL052351

Wetzel, N., Lay, T., Brodsky, E. E., & Kanamori, H. (2017). Rupture-depth-varying seismicity patterns for major and great (Mw > 7.0) megathrust earthquakes. *Geophysical Research Letters*, 44, 9663–9671. https://doi.org/10.1002/2017GL074573

Xu, C., Xu, L., Liu, T., & Sun, W. (2017). Geodetic observations of the co-and post-seismic deformation of the 2013 Okhotsk Sea deep-focus earthquake. *Geophysical Journal International*, 209(3), 1924–1933. https://doi.org/10.1093/gji/ggx123

Ye, L., Lay, T., Bai, Y., Cheung, K. F., & Kanamori, H. (2017). The 2017 Mw 8.2 Chiapas, Mexico, earthquake: Energetic slab detachment. *Geophysical Research Letters*, 44(23), 11,824–11,832. https://doi.org/10.1002/2017GL076685

Ye, L., Lay, T., Kanamori, H., & Rivera, L. (2016). Rupture characteristics of major and great (Mw > 7.0) megathrust earthquakes from 1990 to 2015: I. Source parameter scaling relationships. *Journal of Geophysical Research: Solid Earth, 121*, 826–844. https://doi.org/10.1002/2015JB012426

Yue, H. (2014). Toward resolving stable high-resolution kinematic rupture models of large earthquakes by joint inversion of seismic, geodetic and tsunami observations. (Doctoral dissertation, UC Santa Cruz).