

Mapping finite-fault earthquake slip using spatial correlation between seismicity and point-source Coulomb failure stress change

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Author contributions: *Conceptualization:* A. Lomax. *Methodology:* A. Lomax. *Software:* A. Lomax. *Formal Analysis:* A. Lomax. *Writing - Original draft:* A. Lomax. *Writing - Review & Editing:* A. Lomax.

Abstract Most earthquake energy release arises during fault slip many kilometers below the Earth's surface. Understanding earthquakes and their hazard requires mapping the geometry and distribution of this slip. Such *finite-fault* maps are typically derived from surface phenomena, such as seismic and geodetic ground motions. Here we introduce an imaging procedure for mapping finite-fault slip directly from seismicity and aftershocks—phenomena occurring at depth around an earthquake rupture. For specified source and receiver faults, we map source-fault slip in 3D by correlation of point-source Coulomb failure stress change (Δ CFS) kernels across the distribution of seismicity. These *seismicity-stress* maps show relative, static fault slip compatible with the surrounding seismicity given the physics of Δ CFS; they can aid other slip inversions, aftershock forecasting, and study of early instrumental earthquakes and volcanic intrusions. We verify this procedure recovers synthetic fault slip which matches independent estimates of slip for the 2004 Mw 6.0 Parkfield and 2021 Mw 6.0 Antelope Valley California earthquakes. For the 2018 Mw 7.1 Anchorage Alaska intra-slab earthquake, seismicity-stress maps, combined with multi-scale precise hypocenter relocation, resolve the enigma of which mainshock faulting plane ruptured (the gently east-dipping plane), and clarify slab structures activated in the energetic aftershock sequence.

Non-technical summary The energy and shaking from earthquakes radiates from slip across faults many kilometers deep in the Earth. Mapping the geometry and strength of this fault slip for large earthquakes is fundamental for understanding earthquakes and their hazard. Usually such mapping is done with observations at the Earth's surface. Here we introduce a slip mapping procedure using aftershocks, which form virtual observations surrounding the earthquake at depth. A large earthquake produces extensive deformation and aftershocks in the surrounding rock. The distribution of aftershocks is irregular but can be estimated from the orientation of the large earthquake fault and assumed orientations for the aftershock faults. By comparing these estimated aftershock distributions to the distribution of observed aftershocks we obtain maps of the large earthquake slip. We verify that this mapping procedure correctly recovers simulated fault slip, and matches independent estimates of slip for two California earthquakes. For the 2018 Mw 7.1 Anchorage Alaska earthquake we show that these slip maps, combined with precise aftershock relocations, allow determination of which fault the earthquake ruptured, and which secondary faults and geologic structures were activated in its energetic aftershock sequence.

1 Introduction

Earthquakes are well described by elastic rebound, the sudden release of elastic strain energy stored in a rock mass (Reid and Lawson, 1908; Scholz, 2018). This energy release occurs mainly at seismogenic depth, a brittle zone many kilometers below the Earth's surface (Sibson, 1982; Marone and Scholz, 1988), and is clearly manifested by slip across a fault. Comprehensive understanding of large earthquakes and their hazard requires mapping the orientation, extent, amplitude and time-variation of this fault slip. Such *finite-fault slip maps* are typically obtained through static or kinematic inversion of surface phenomena, measured far from the

fault slip at depth. These surface observations include fault offsets, shaking intensity and time-varying or differential ground motion from seismogram, geodetic and space-based data such as global navigation satellite system (GNSS) and interferometric synthetic aperture radar (InSAR) (Olson and Apsel, 1982; Hartzell and Heaton, 1983; Ide, 2007; Mai et al., 2016). The resulting finite-faulting maps depend on the selection, quality and distribution of observations, resolution of the inversion, and on numerous assumptions and parameters, such as a pre-defined, 2D fault surface geometry, models for earth structure, rupture speed and rise-time, an inversion method, and smoothing constraints or other regularization; consequently, finite-faulting maps for a single earthquake produced by different studies are often clearly different (Harris and Segall, 1987; Beresnev,

Production Editor:
Andrea Llenos
Handling Editor:
Ake Fagereng
Copy & Layout Editor:
Ethan Williams

Received:
15 July 2024
Accepted:
13 January 2025
Published:
3 March 2025

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2003; Bos and Spakman, 2003; Hartzell et al., 2007; Ide, 2007; Minson et al., 2013; Mai et al., 2016; Wang et al., 2020; Wong et al., 2024).

The release of strain energy during earthquake slip also produces aftershocks and seismicity—phenomena occurring at seismogenic depth around an earthquake rupture. Effectively, aftershocks and seismicity form in-situ and nearby observations that can provide information on fault slip. Earthquake slip releases strain energy and perturbs stress in the surrounding rock mass through stress transfer (Chinnery, 1963). Under the Coulomb failure stress hypothesis, slip on a fault occurs when shear stress is sufficiently high and normal stress sufficiently low to overcome frictional resistance to sliding. In accordance with these concepts, stress changes due to an earthquake rupture will favor triggering of surrounding aftershocks within the positive lobes of the 3D spatial distribution of change in Coulomb failure stress (Δ CFS) (Stein and Lisowski, 1983; Oppenheimer et al., 1988; King et al., 1994). Numerous studies show that, given a reasonable finite-faulting model for a mainshock, forward calculation of Δ CFS shows positive and negative lobes that often correlate well with observed distributions of occurrence or lack of aftershocks, respectively (Oppenheimer et al., 1988; King et al., 1994; Harris and Simpson, 1996; Harris, 1998; Toda et al., 1998; Stein, 1999; Toda et al., 2011; Sato et al., 2012).

A few previous studies have used aftershock seismicity and Δ CFS to constrain parameters of mainshock slip. For the 1992 M 7.3 Landers, California earthquake, Seeber and Armbruster (2000) use aftershock seismicity and Δ CFS in an iterative procedure to determine slip distributions given a fixed, 2D rupture geometry of 26 segments with uniform slip and a fixed total moment. Seeber and Armbruster (2000) explore numerous method variants, assumptions and parameters to obtain a set of slip distributions which compare favorably with those from seismic waveforms and geodetic data, and with surface slip observations. Budiman et al. (2019), for the 2006 Yogyakarta Earthquake, Indonesia, and Gahalaut et al. (2003) for the 1993 Killari, India earthquake, use the correspondence of calculated Δ CFS with the aftershock distribution to explore and select between different causative fault-planes, and, in Gahalaut et al. (2003), distribution of slip with depth on these planes. All of these studies use somewhat ad-hoc, trial-and-error procedures and require specification of 2D surfaces for mainshock faulting.

Here we introduce and illustrate a straightforward imaging procedure for mapping finite-fault slip (or tensile opening) directly from seismicity and aftershocks, and without need to define a 2D, source fault geometry. For specified source and receiver fault orientations, we construct 3D maps of extended source slip through correlation of point-source Δ CFS kernels across the spatial distribution of seismicity around a source earthquake. This *seismicity-stress* procedure finds, in accordance with the physics of the Coulomb failure stress criteria, a 3D distribution of relative, finite-fault slip compatible with the surrounding distribution of seismicity. Application of the procedure is mainly dependent on

the availability of multi-scale precise relocations of seismicity surrounding a source zone of earthquake rupture, aseismic slip, fault creep, dyke intrusion or other strain source, and, for Δ CFS kernel calculation, on the specification of a shear dislocation or tensional mechanism for the source, a shear slip mechanism and faulting plane for receiver faults, and distance parameters for masking of the singularity at the Δ CFS point-source. Seismicity-stress slip maps might be used as prior constraint on fault geometry and slip for other slip inversion methods and monitoring of volcanic intrusions, for quasi data-driven aftershock forecasting, to aid in rapid shaking characterization and tsunami early-warning, to obtain finite-faulting information for large earthquakes occurring before the advent of modern seismic and geodetic networks, and to identify possible regions of aseismic slip driving foreshock sequences.

We confirm that seismicity-stress imaging correctly maps a synthetic slip distribution on two rectangular fault patches given random seismicity predicted by Δ CFS for the synthetic slip. We next use background seismicity and aftershocks along the Parkfield California segment of the San Andreas fault to produce seismicity-stress slip maps that match well other estimates of co- and post-seismic slip for the 2004 Mw 6.0 Parkfield earthquake, and of adjacent, long-term slip related to fast creep. We further confirm agreement between seismicity-stress slip maps and finite-faulting results from other methods for the 2021 Mw 6.0 Antelope Valley California normal-faulting earthquake. These tests and applications to well studied sequences verify the functioning of the imaging procedure and show its general agreement with previous finite-faulting results; as this is an introductory study, we leave more detailed and extensive examination of the behavior of the methodology for future work. Finally, we show for the 2018 Mw 7.1 Anchorage Alaska, normal-faulting intraslab earthquake how the seismicity-stress procedure, combined with multi-scale precise hypocenter relocation, resolves the enigma of which mainshock faulting plane ruptured (the gently east-dipping plane), and identifies seismotectonic structures in the slab activated in the energetic aftershock sequence.

2 The seismicity-stress procedure for mapping relative finite-fault slip in 3D

If an earthquake fault is considered a simple frictional surface with cohesion, then, under the Coulomb failure criteria (Coulomb, 1773; Weiss et al., 2016), fault slip occurs when shear stress in the direction of slip becomes sufficiently high to overcome cohesion, and normal stress sufficiently low to remove frictional resistance to sliding. Though absolute stresses on a fault cannot be calculated, the change in stresses due to a source earthquake rupture with specified, finite-fault slip can be determined throughout an elastic volume (Chinnery, 1963; Okada, 1992). For specified receiver fault orientations, these stress changes allow determination of the

change in Coulomb failure stress, ΔCFS ,

$$\Delta CFS = \Delta\tau + \mu' \Delta\sigma_n, \quad (1)$$

where $\Delta\tau$ is change in shear stress on the fault, positive in the direction of slip, $\Delta\sigma_n$ is change in normal stress on the fault, positive for unclamping, and μ' an effective coefficient of friction which includes the effects of pore pressure changes (Stein and Lisowski, 1983; Oppenheimer et al., 1988; King et al., 1994; Stein, 1999). Stress changes due to the source slip will favor triggering of aftershocks or larger earthquakes on receiver faults within the positive lobes of the 3D spatial distribution of ΔCFS , and inhibit the occurrence of aftershocks or larger earthquakes in the negative lobes, often called stress shadows.

The 3D, scalar ΔCFS field due to finite-fault slip can be obtained through integration over the fault of the point-source ΔCFS field of stresses due to a point shear dislocation or tensile opening in an elastic half-space (Haskell, 1964; Kikuchi and Kanamori, 1982; Okada, 1992; King et al., 1994; Materna and Wong, 2023). This procedure is a forward calculation, with the resulting ΔCFS field often assessed by how well it agrees with the surrounding distribution of aftershocks and larger earthquakes. The corresponding inverse problem is to infer a distribution of finite-fault slip given the 3D distribution of aftershocks or other seismicity and the physics of the Coulomb failure stress criteria. Here we infer finite-fault slip for a target source event through an imaging methodology: for specified source and receiver fault orientations, we obtain 3D maps of source slip through correlation of point-source ΔCFS kernels across the spatial distribution of post-event seismicity around the target event. This seismicity-stress procedure finds a 3D distribution of relative finite-fault slip which can explain, in accordance with the physics of the Coulomb failure stress criteria, the surrounding distribution of post-event seismicity.

2.1 Correlation of ΔCFS kernels over post-event seismicity

The similarity of two signals or functions can be measured with cross-correlation. Here we are interested in the similarity of a point-source ΔCFS kernel field as it moves across an unknown, seismicity rate change field ΔS represented by a distribution of observed seismicity $\hat{S}$ within a specified time window after the target event. To map this similarity in 3D we cross-correlate the two fields: as the point-source kernel origin is shifted across the seismicity in all three spatial dimensions, the dot product between the shifted kernel field and the seismicity field provides a measure of similarity which is assigned to the current point-source origin. Given a point-source ΔCFS kernel field and a seismicity distribution $\hat{S}$ in some time window, both sampled on a 3D grid with indices (i, j, k) , a seismicity-stress, finite-fault slip field F on the grid is obtained as,

$$F(i, j, k) = \sum_l \sum_m \sum_n \Delta CFS(l+i, m+j, n+k) \hat{S}(l, m, n), \quad (2)$$

where the range of each summation index, l, m, n , is limited by the bounds of the ΔCFS and $\hat{S}$ field grids.

The use of a cross-correlation between a Green's function (the point-source ΔCFS kernel) and observations (the seismicity) to infer a model (fault slip) is well established as adjoint operator back-projection or imaging, which forms an approximate, but practical and robust alternative to formal inversion (Kawakatsu and Montagner, 2008; Claerbout, 2010; Fukahata et al., 2013). The approximation arises in part because imaging ignores physics-based, weighting or scaling functions between the model and data spaces, e.g. the normal matrix in least-squares matrix inversion is implicitly replaced by the identity matrix (Claerbout, 1992). In consequence, the seismicity-stress field F is an imaging or brightness function but not a formally defined physical quantity, and so here we refer to the field F as a measure of *relative* fault dislocation or fracture opening.

A point-source ΔCFS kernel field is fully sampled across the study volume and typically forms a pattern of four positive and four negative lobes for a double-couple, shear dislocation source (Figure 1). In contrast, observed seismicity, $\hat{S}$, may or may not occur in areas of positive ΔCFS , and may occur in areas of negative ΔCFS at a much lower rate than in the positive ΔCFS regions (Hardebeck and Harris, 2022). Seismicity will be absent in parts of the surrounding rock mass that do not support brittle fracture or lack a sufficient density of existing receiver faults and fractures (Liu et al., 2003; Barchi et al., 2021), while the density of seismicity can vary greatly due to spatial variations in the distribution and limits of previous slip, aftershocks of aftershocks, background stress state, and areas of high background seismicity and swarm activity. Thus, the post-source seismicity, $\hat{S}$, is an estimate of the ΔS field and can only provide a sparse, noisy, and irregular sampling of the positive regions of the unknown, true ΔCFS field for the target source event. In order to avoid bias due to this incomplete sampling of true ΔS , we accumulate the spatial distribution of observed seismicity, $\hat{S}$, on a 3D grid using a value of +1 for cells containing seismicity and a value of 0 for other cells. Ideally, we would assign -1 to cells without seismicity and in the negative parts of the true ΔCFS , but this information is unavailable. We should also assign -1 to cells with inhibited or reduced seismicity rates if these can eventually be determined, for example through comparison of pre- and post-event seismicity.

Thus, under the physics of ΔCFS , ΔS values of +1 flag areas where seismicity may be favored by the point-source dislocation or opening, while 0 values flag areas where the seismicity may be either favored or is suppressed. In cross-correlation between the gridded ΔCFS kernel and ΔS seismicity fields, the aftershock grid +1 values can contribute to the correlation sum, while the 0 values will provide no information to and not change the sum. The resulting correlation sum should be somewhat insensitive to missing aftershocks in areas of true positive ΔCFS for the target source event and to the impossibility of identifying lack of aftershocks in areas of true negative ΔCFS , as long as the distribution of available aftershocks in 3D samples sufficiently well the pos-

itive target event Δ CFS field.

2.2 Precision of relocated seismicity

Meaningful application of the seismicity-stress procedure requires multi-scale precise relocations of seismicity surrounding a zone of earthquake slip or fault creep. That is, the distribution and relative positions of relocated hypocenters should represent well the true distribution of seismicity over the range of scales from the size of the study area down to the desired resolution of finite-faulting maps. Otherwise, errors in the hypocenter locations will map directly through correlation with the Δ CFS kernels into distortions or reduced resolution in the finite-faulting maps. Thus the relocated seismicity might use, at least, carefully determined station static corrections, or, preferably, explicitly multi-scale corrections such as source-specific, station traveltime corrections (SSST; Richards-Dinger and Shearer, 2000).

Here we obtain multi-scale high-precision earthquake relocations with the NLL-SSST-coherence procedure (NLL-SC; Lomax and Savvaidis, 2021; Lomax and Henry, 2023). NLL-SC achieves multi-scale precise hypocenter location through the combination of SSST corrections and stacking of probabilistic locations for nearby event based on inter-event waveform coherence. NLL-SC is based on the NonLinLoc location algorithm (NLL hereafter; Lomax et al., 2000, 2014), which performs efficient, global sampling to obtain an estimate of the posterior probability density function in 3D space for hypocenter location. This probability density function provides a comprehensive description of likely hypocentral locations and their uncertainty, and enables application of the waveform coherence relocation. Within NLL, we use the equal differential-timing likelihood function (Zhou, 1994; Font et al., 2004; Lomax, 2005, 2008; Lomax et al., 2014), which is highly robust in the presence of outlier data caused by large error in phase identification, measured arrival-times or predicted travel-times, and thus helps stabilize the generation of SSST corrections which depend strongly on arrival residuals.

2.3 The seismicity-stress imaging procedure

Given a target source event such as an earthquake, aseismic slip, fault creep or dyke opening, application of the seismicity-stress procedure to map finite-fault slip requires multi-scale precise relocations of post-event seismicity in a study volume surrounding a source event, and specification of a fault mechanism for the source and fault orientation and slip parameters for the receiver seismicity. The procedure includes the following steps:

- Specify a source mechanism (e.g. double-couple shear dislocation or tensile fracture opening); the causative source-slip plane does not need to be specified since for a point double-couple the produced stress field is independent of which of the two planes slips.
- Specify a receiver fault plane orientation and rake; the receiver slip plane must be specified so that the

shear and normal stresses produced by the source can be resolved onto the receiving fault plane.

- Specify the elastic parameters shear modulus and Lamé's first parameter.
- Establish a point-source Δ CFS field for the specified source and receiver faults within a 3D grid larger than the study volume. Here, this Δ CFS kernel grid is assembled from the point-source Δ CFS fields for a stack of gridded, horizontal layers created using the Elastic_stresses tool (Materna and Wong, 2023) for calculating elastic displacements, strains, stresses and Δ CFS. Appropriate elastic parameters are chosen for the average depth of seismicity and the point source is placed at sufficient depth to minimize effects of the free surface. Figure 1 shows the double-couple point-source Δ CFS field for a vertical, right-lateral strike slip sources and receivers corresponding to mechanisms of events along the Parkfield section of the San Andreas fault. For specification of vertical faults, dip 89.9° deg is used instead of exactly 90.0° which produces a singularity in the Elastic_stresses calculation.
- Mask the Δ CFS kernel grid in a region around the point-source origin coordinates, $\mathbf{X}_{ps}$, to avoid locally very high values of Δ CFS due to singularities in the stress calculations (Okada, 1992). This mask $M(\mathbf{X})$ is defined by:

$$M(\mathbf{X}) = \begin{cases} 0, & |\mathbf{X} - \mathbf{X}_{ps}| \leq R_m \\ 1 - e^{-\frac{(|\mathbf{X} - \mathbf{X}_{ps}| - R_m)^2}{L_m^2}}, & |\mathbf{X} - \mathbf{X}_{ps}| > R_m, \end{cases}$$

where $\mathbf{X}$ is a grid location, R_m is a radius around $\mathbf{X}_{ps}$ within which the mask is set to 0.0, and L_m is a decay length controlling an inverse Gaussian rise of the mask value to 1.0 far from the point-source. Full masking of the singularity around the point-source is controlled by R_m , which may be typically set to 1-2 times the grid spacing. The decay length, L_m , sets the smoothness of the termination of this masking and controls the distance from the point-source to the larger positive and negative amplitudes in the kernel field. L_m thus allows targeting the seismicity-stress imaging to the length scale of source rupture and typical distance between rupture and post-event seismicity. L_m might typically be set to roughly the extent of denser, post-event seismicity around the target source event and around 10 times R_m . For a large, multi-rupture earthquake, L_m might be set to reflect the length scale of the main rupture segments. Supplementary Figure S1 shows the masking functions used in this study and illustrates different settings for L_m and R_m and their effects on the masking.

- Accumulate the post-event seismicity into a similar size 3D grid by assigning +1 to a grid cell if there are one or more hypocenters in the cell and assigning 0 otherwise.

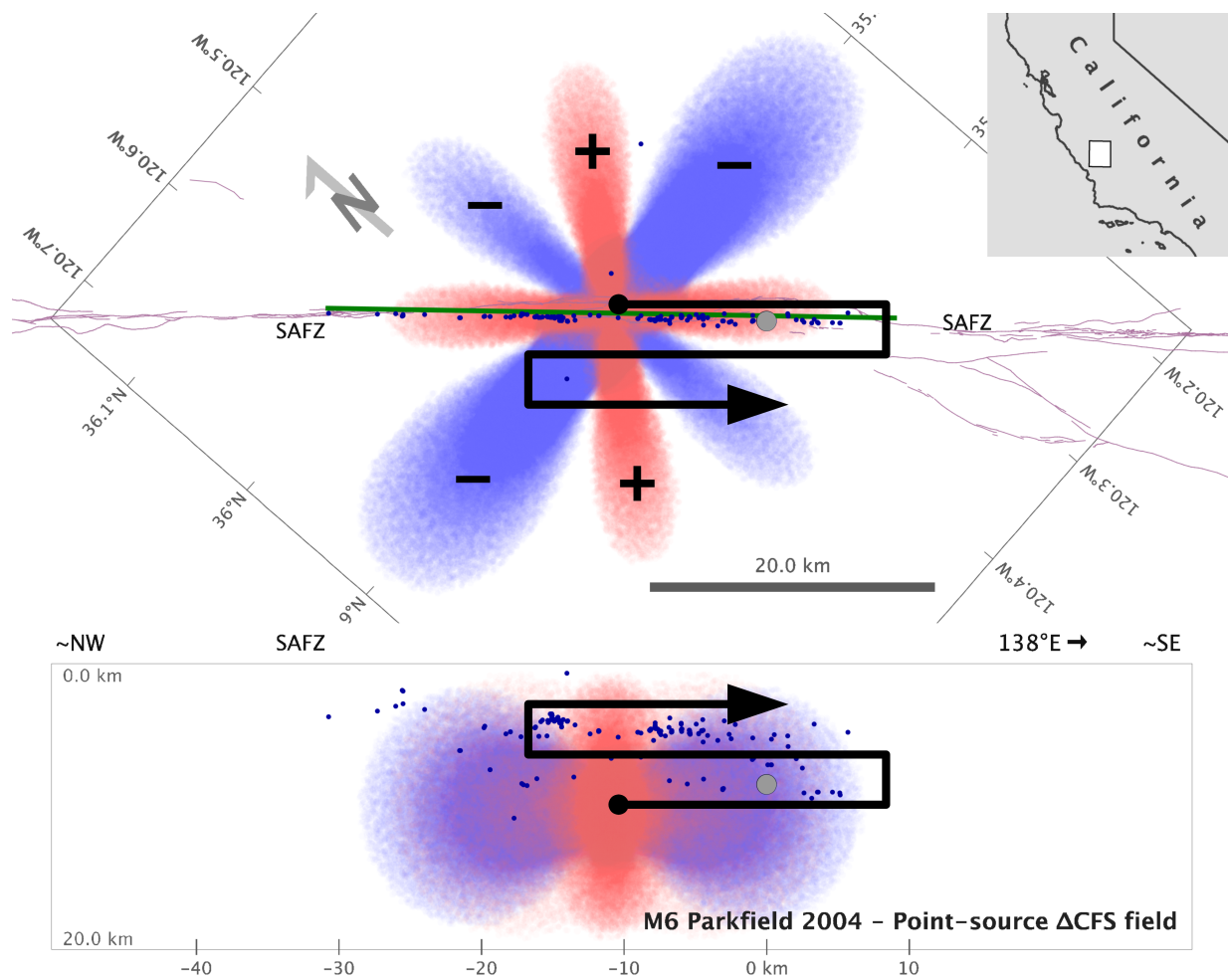


Figure 1 Point-source ΔCFS kernel field for vertical, right-lateral strike slip source and receiver faults corresponding to the geometry of the 2004 Parkfield sequence. Blue dots show 4 hours of aftershocks after the 2004 Parkfield mainshock (large gray dot). The ΔCFS field is plotted as a 3D density cloud (red positive; blue negative) and is masked to avoid singularities in the stress calculations (zero mask radius $R_m = 0.5$ km, decay length, $L_m = 10$ km, see Supplementary Figure S1). The position and depth of the plotted kernel field in the study area is arbitrary. The arrow paths represent 3D shifting and correlation of the point-source kernel across the seismicity with assignment of the correlation value to the corresponding center of the kernel (black dot) to give seismicity-stress imaging. The green line shows the source fault strike, chosen to align with aftershocks and the overall strike of the San Andreas fault zone (SAFZ). Purple lines show faults from the USGS Quaternary fault and fold database for the United States.

- Correlate the masked, ΔCFS kernel grid across the seismicity grid to produce a 3D grid of potential, relative, static finite-fault slip.
- Normalize the finite-fault grid to its maximum value. Here, to aid in visualization and analysis, we create a second, high-potential grid where values ≤ 0.5 are removed and remaining values rescaled so that values $[0.5, 1.0]$ map linearly to $[0.0, 1.0]$ this grid shows the extent of faulting potential that is greater than half the peak value.
- Optionally, clip the finite-fault grids to the convex hull of post-event seismicity. This clipping is useful for visualization and interpretation since, in general, potential slip in volumes not surrounded by seismicity will not be well constrained by the procedure. Similar masking is used in many inversion and imaging methods, for example, in seismic tomography for areas with little or no ray coverage.

There may be cases where seismicity-stress clipping is undesirable, for example in attempting to image strain due to tremor or ductile slip below an active seismogenic zone.

- Output the finite-fault slip grids for visualization and further analysis.

The resulting finite-fault grids map a 3D distribution of static slip (or tensile opening) which explains, in accordance with the physics of ΔCFS , the surrounding distribution of post-event seismicity. Essentially, the procedure resolves complex, 3D distributions of finite-fault slip through reverse mapping of aftershock locations through the non-isotropic lobes of the ΔCFS field.

The seismicity-stress maps show *relative* slip since no formal theory relating source faulting to the distribution or rate of seismicity is used and, moreover, the seismicity is simplified to counts of 0 or +1 on a grid. The maps show *potential* slip since they show the envelope of

an ensemble of allowable slip solutions compatible with the seismicity, and not a single slip model that additionally satisfies physical constraints such as energy conservation and localization of slip to 2D surfaces. The timing of the slip is only constrained to have been before or during the time range of the seismicity; some of the nominally post-event seismicity may be a response to seismic or a-seismic slip before the target source event.

The seismicity-stress procedure is simple and rapid. On an 8-core desktop workstation, the calculation of a Δ CFS kernel grid takes around 1 minute. Subsequent cross-correlation with an aftershock grid and output of a 3D finite-fault grid takes around 10 seconds, mostly for input and output. Thus, after a large earthquake, given relocations of aftershocks and pre-calculated Δ CFS kernel grids for different source and receiver faulting parameters, sets of 3D finite-fault grids reflecting the evolving seismicity and different faulting scenarios can be generated in near real-time to aid in rapid hazard analysis and aftershock forecasting.

3 Validation for synthetic fault slip on rectangular fault patches

To validate the seismicity-stress procedure in a controlled and simplified case we consider synthetic slip on a vertical right-lateral, strike-slip fault with strike $\sim$ N140°E, following the geometry of the Parkfield segment of the central San Andreas fault (SAF; Figure 1). Slip is imposed in two different sized patches on the fault, each of which has constant slip and the same total moment release (Figure 2). We generate a representative set of synthetic aftershocks that might be produced by the synthetic slip distribution by: 1) calculating the stress changes and Δ CFS field (without masking) for the synthetic slip using the Elastic_stresses tool (Materna and Wong, 2023), 2) sampling in proportion to the amplitudes in the positive lobes of this field to obtain a set of 474 synthetic aftershocks.

We specify the same strike-slip faulting as the synthetic slip for source and receiver faults to construct a double-couple, point-source Δ CFS kernel field on a grid of 80 km $\times$ 80 km horizontal $\times$ 20 km in depth grid with cell size 0.5 km (Figure 1). We use a zero mask radius, R_m , of 0.5 km and decay length, L_m , of 10 km (Supplementary Figure S1).

Correlation of the masked, Δ CFS kernel grid across the gridded, synthetic seismicity produces the 3D grid of potential, relative, static finite-fault slip represented in Figure 2. The position and lateral extent of the two patches of synthetic fault slip are well recovered within the high-potential areas of the finite-faulting field (red). However, these high-potential areas spread beyond the imposed slip in depth and perpendicular to the fault horizontally, likely due to the fairly dense seismicity within the positive lobes of the Δ CFS field for the synthetic slip in these directions. Apparently, the vertically extended, sheet-like lobes of the point-source Δ CFS kernel (Figure 1) correlate well with the seismicity when the kernel origin is displaced from the synthetic slip within some distance, likely related to the decay length,

L_m , of the kernel masking.

4 Tests on two California earthquake sequences

We test the seismicity-stress imaging procedure through application to two recent earthquake sequences in California, the 2004, Mw 6.0 Parkfield sequence, a geometrically simple case of strike-slip faulting and most aftershocks on the same vertical plane, and the 2021, Mw 6.0 Antelope Valley sequence, with a normal faulting mainshock and a more complex, 3D distribution of aftershocks. In both cases we find generally good agreement with other finite-faulting results based on seismic waveform and geodetic data, but also a discrepancy in the location of highest slip relative to that found by inversion of seismic waveforms only.

4.1 The 2004, Mw 6.0 Parkfield sequence

We first consider the 2004 Mw 6.0 Parkfield, California sequence and background seismicity along the well instrumented, Parkfield segment of the central San Andreas fault (SAF; Figure 3). This segment is at the southeastern end of a 150 km long, near-linear stretch of the SAF which exhibits surface creep, repeating earthquakes and aseismic slip (Steinbrugge et al., 1960; Savage and Burford, 1973; Nadeau and McEvilly, 2004; Titus et al., 2006; Jolivet et al., 2015), and is just northwest of a long, locked section of the fault which last ruptured in the 1857 M 7.9 Fort Tejon earthquake (Bakun et al., 2005; Langbein et al., 2005). The 1966 M $\sim$ 6 and 2004 Mw 6.0 Parkfield earthquakes both ruptured nearly the same fault patch on the Parkfield segment but initiated at opposite ends of this patch and propagated in opposing directions (Bakun et al., 2005). At seismogenic depth the Parkfield segment is considered to be a single, near-planar surface along the Southwest Fracture zone, which aligns linearly with the SAF to the northwest and southeast of Parkfield, but does not follow the curved, main San Andreas surface trace near Parkfield (Simpson et al., 2006; Thurber et al., 2006; Lomax and Henry, 2023).

For seismicity-stress finite-faulting analysis, we consider here multi-scale precise, NLL-SC relocations for the Parkfield area (Lomax and Savvaidis, 2021; Lomax and Henry, 2023) using the earthquake catalog and phase arrival data provided by the Northern California Earthquake Data Center (NCEDC). These relocations places most background seismicity and 2004 aftershocks on a smooth, near-planar surface striking $\sim$ N140°E under and parallel to the Southwest Fracture zone. As the 2004 mainshock and most aftershocks and background seismicity have right-lateral strike-faults focal mechanisms (Thurber et al., 2006) with one fault plane in close agreement to the strike of the planar surface, we use vertical, strike-slip faulting with strike N140°E as the faulting mechanism for source and receiver faults. We construct a double-couple, point-source Δ CFS kernel field on a grid of 80 km $\times$ 80 km horizontal $\times$ 20 km in depth grid with cell size 0.5 km

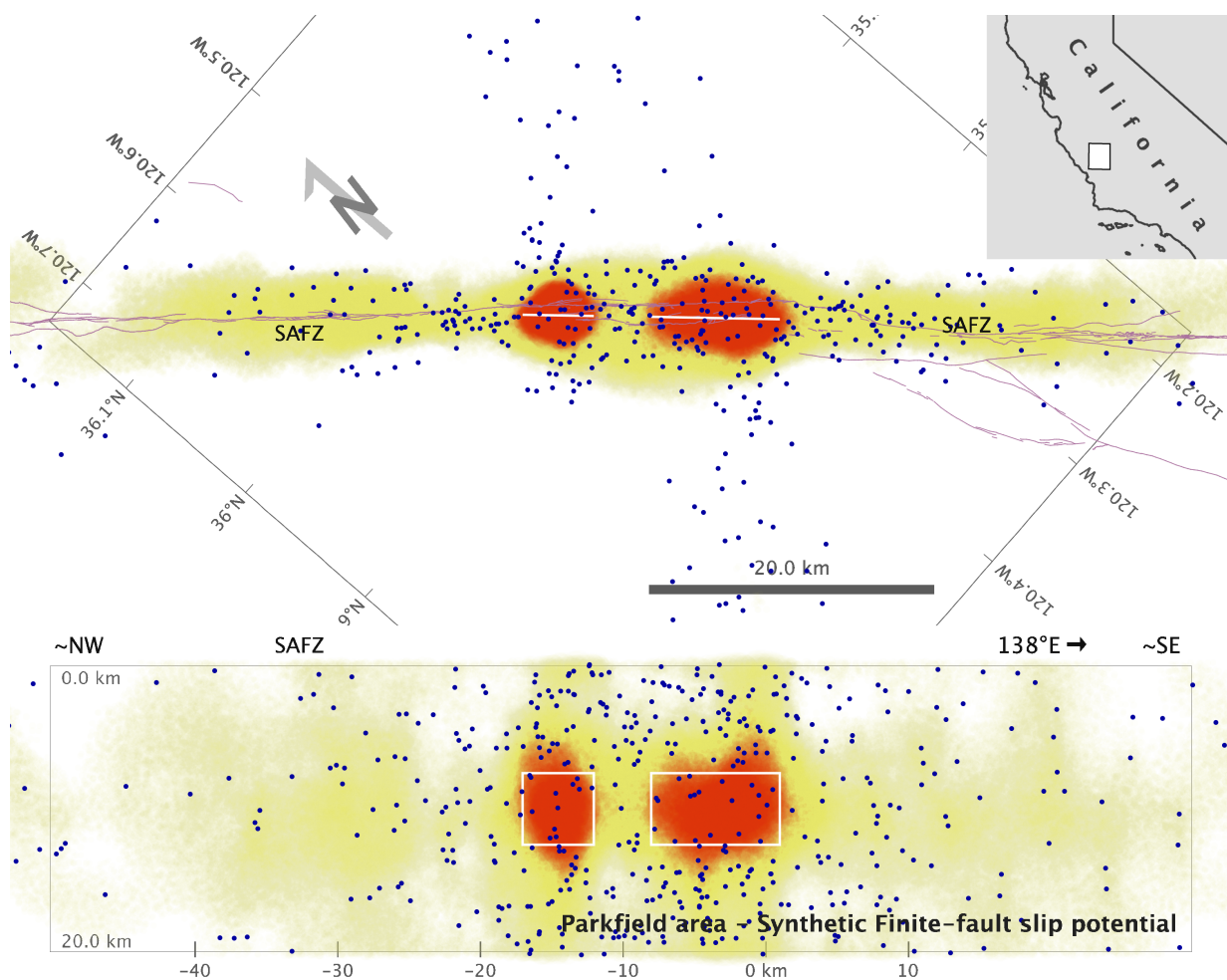


Figure 2 Seismicity-stress, 3D finite-faulting potential inferred from seismicity (blue dots) sampled from the ΔCFS for synthetic rupture on a vertical fault (white rectangles). Synthetic slip for vertical, right-lateral strike-slip faulting corresponding to the geometry of the 2004 Parkfield sequence (Figure 1) is distributed uniformly within two patches: 5×5 km, to the northwest, and 5×10 km, to the southeast (white rectangles). Correlation is performed with a ΔCFS kernel for a point-source and receiver faults with the same faulting as the synthetic slip (Figure 1). The relative amplitude of the seismicity-stress finite-faulting field is shown as a 3D density cloud in tones of yellow for portions of the field with normalized correlation < 0.5 and in red for the high-potential portions of the field with normalized correlation ≥ 0.5 . Field values are plotted in 3D for each grid cell as transparent disks with radius increasing with correlation value; higher color saturation indicates higher correlation and/or deeper volumes of high correlation. The fields are not clipped to the convex hull of the seismicity. SAFZ indicates the San Andreas fault zone; purple lines show faults from the USGS Quaternary fault and fold database for the United States.

(Figure 1). We use a zero mask radius, R_m , of 0.5 km and decay length, L_m , of 10 km (Supplementary Figure S1).

Correlation of the masked, ΔCFS kernel grid across the gridded NLL-SC seismicity for the first 4 hours after the 2004 Parkfield mainshock origin (at 2004-09-28 17:15:24 UTC) produces the 3D grid of potential, relative, static finite-fault slip represented in Figure 3. This finite-faulting field falls along and tightly around the vertical plane of seismicity and likely San Andreas fault plane at seismogenic depth, a correspondence that is constrained by the distribution of seismicity mainly along this plane and by the positive lobes of the ΔCFS kernel field aligned parallel to this plane (Figure 1).

The high-potential region of the finite-fault field (red) forms a horizontal band at $\sim 3 - 7$ km depths extending from just above to about 15 – 20 km northwest of the Mw 6.0 hypocenter, with the highest slip potential about 11 km northwest of the hypocenter in an ~ 4 km diame-

ter, circular zone. The depth and along-fault ranges of this high potential slip region corresponds well to independent estimates of the geometry of co-seismic slip and energy radiation, especially given the diversity and uncertainty of these estimates (e.g. Hartzell et al., 2007). A main difference is that most slip inversions based on seismic and GNSS data find highest slip at around 15 – 20 km northwest of the Mw 6.0 hypocenter (Johanson et al., 2006; Langbein et al., 2006; Liu et al., 2006; Wang et al., 2012; Twardzik et al., 2012), though some inversions (e.g. Kim and Dreger, 2008; Bruhat et al., 2011) find highest slip closer to the 11 km northwest of the hypocenter found by seismicity-stress imaging. The position of the seismicity-stress peak agrees better with network and array based analyses of high-frequency seismic radiation from the 2004 mainshock which find the largest sources of high-frequency radiation $\sim 7 - 13$ km northwest of the hypocenter (Fletcher et al., 2006; Allmann and Shearer, 2007; Fountoulakis and Evangelidis, 2024).

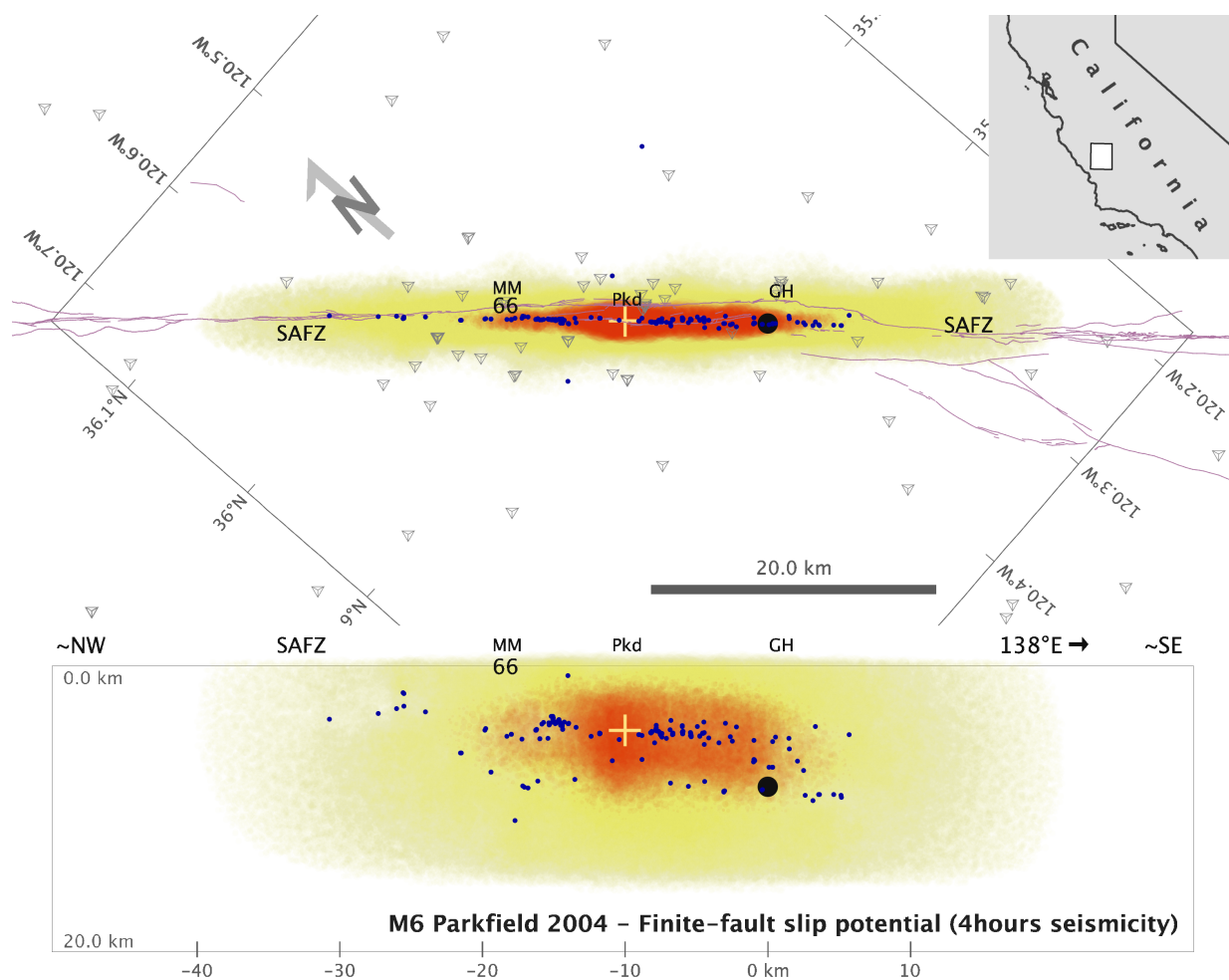


Figure 3 Seismicity-stress, 3D finite-faulting potential inferred from the first 4 hours of aftershocks (blue dots) after the 2004 Parkfield mainshock (large black dot). Correlation is performed with a Δ CFS kernel for vertical, right-lateral strike-slip point-source and receiver faults (Figure 1). The relative amplitude of the seismicity-stress finite-faulting field is shown as a 3D density cloud in tones of yellow for portions of the field with normalized correlation < 0.5 and in red for the high-potential portions of the field with normalized correlation ≥ 0.5 , the yellow cross shows the peak value in this field. Field values are plotted in 3D for each grid cell as transparent disks with radius increasing with correlation value; higher color saturation indicates higher correlation and/or deeper volumes of high correlation. The fields are not clipped to the convex hull of the seismicity, but fading of the fields towards the limits of the plots may be an artifact of lack of available seismicity outside the plot and not absence of potential slip. The 1966 Parkfield mainshock epicenter (McEvelly et al., 1967) is indicated by “66”, locations of seismographs shown as inverted triangles, other abbreviations are: SAFZ – San Andreas fault zone, MM – Middle Mountain, Pkd – Parkfield, GH – Gold Hill. Purple lines show faults from the USGS Quaternary fault and fold database for the United States.

The seismicity-stress procedure can also map temporal changes in areas of potential slip by application to seismicity in different time windows. Figure 4 shows 3D potential finite-fault slip maps along the SAF for seismicity in time windows before and after the 2004 Parkfield mainshock origin. The potential slip map for the first 4 hours after the mainshock (second row) shows the same results as Figure 3, with highest slip along the likely 2004 rupture zone and representing mainly co-seismic rupture. The potential slip maps for the 10 year periods 1984-1994 (Supplementary Figure S2) and 1994 to 2004 (top row in Figure 4) show highest slip potential along the southeastern end of the 150 km section of the central SAF where there is known background, fast creep, repeating earthquakes and aseismic slip, and the southeastern termination of this highest slip potential falls at the northwestern limit of the Parkfield

locked asperity and 1966 and 2004 Parkfield rupture zone around Middle Mountain (Nadeau and McEvelly, 2004; Titus et al., 2006; Maurer and Johnson, 2014; Jolivet et al., 2015).

The potential slip map for the period from 1 week to 1 month after the 2004 mainshock shows highest slip potential along the 2004 co-seismic rupture zone (as defined by the potential slip map for the first 4 hours), but also further northwest into the fast-creeping section, which may represent capture of background seismicity and creep in this section, not slip triggered by the mainshock. This slip map also shows a broadening, deepening and shallowing of highest slip potential around the 2004 co-seismic rupture, which may represent after-slip of this rupture, in general agreement with previous studies (e.g. Johanson et al., 2006; Langbein et al., 2006; Freed, 2007).

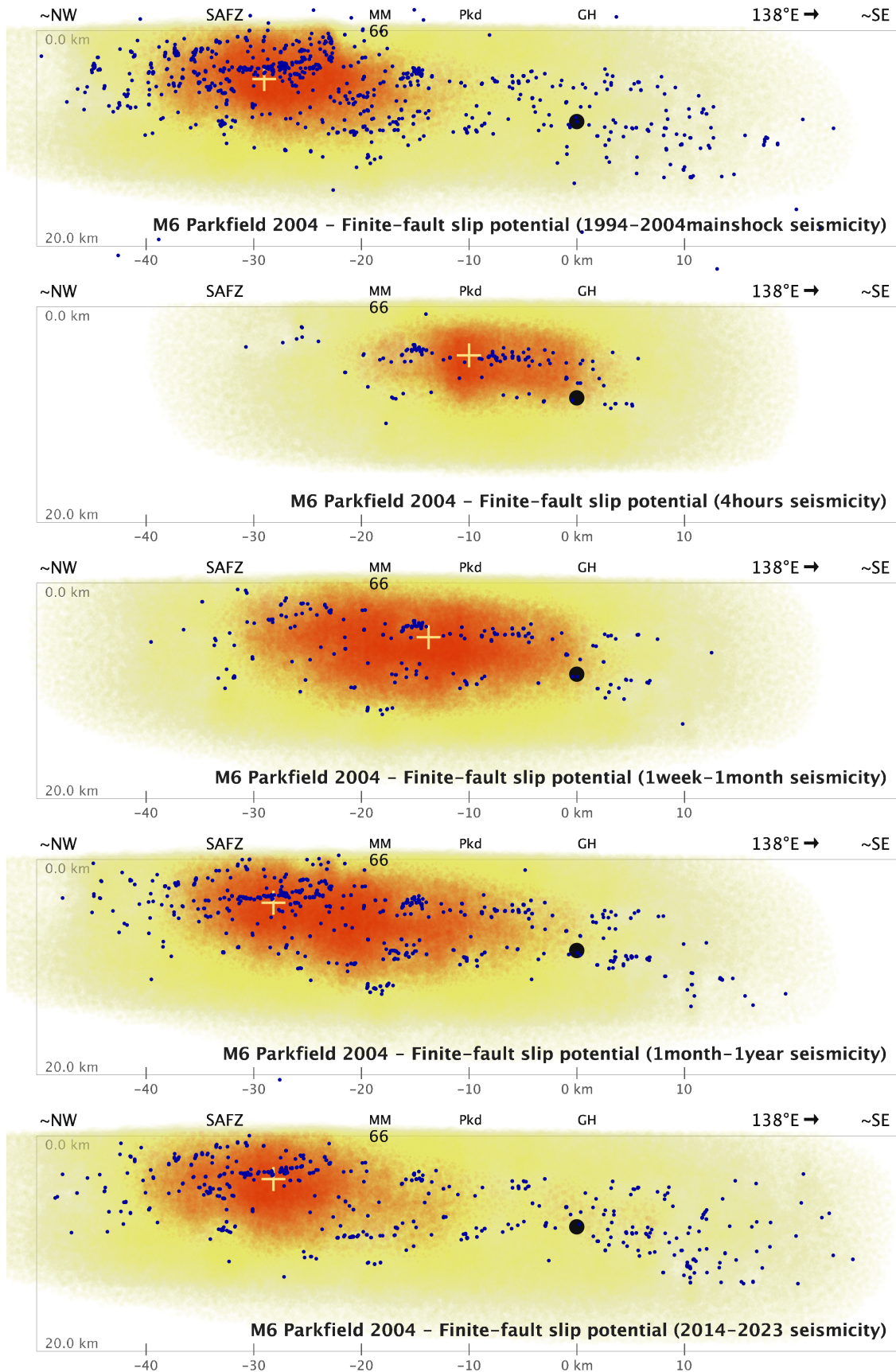


Figure 4 Potential slip maps along the SAF for seismicity in a selection of time windows around the 2004 Parkfield mainshock. Maps labelled with date pairs show potential slip for aftershocks between those dates, those labelled with time spans show potential slip for that span after the 2004 mainshock origin. Map elements as in Figure 3. These and additional slip maps forming a contiguous set of time windows between 1984 and 2023 are presented in Supplementary Figure S2.

The potential slip map for the period from 1 month to 1 year after the 2004 mainshock shows a continuation of increasing response to background seismicity and creep to the northwest, and decreasing imaging of after-slip around the mainshock rupture. The peak of potential slip (yellow crosses in Figure 4) is shifted to the northwest into the fast-creeping section at about the same location as for the 10 year periods before and after the 2004 sequence. The slip map for the 9+ year period 2014 to 2023 (bottom row in Figure 4) is similar to those for 1984–1994 (Supplementary Figure S2) and 1994 to 2004 (top row in Figure 4), indicating that the majority of slip at depth and seismicity by 10 years after the 2004 mainshock is not distinguishable from the pre-mainshock background seismicity. Notably, the south-east limit of slip potential for the creeping section of the SAF as imaged by background seismicity from 1984–2004 and from 2014 to 2023 corresponds closely to the northwest limit of higher, 2004 co-seismic slip potential around 15–20 km northwest of the 2004 hypocenter (Figure 3).

The nature of the seismicity-stress methodology entails that these space-time variations in potential slip reflect local changes in density and extent of clusters of seismicity along the SAF in the Parkfield area. The overall distribution of clusters of seismicity, however, appears similar in the ~10 year time windows at the beginning (1994–2004) and end (2014–2023) of the study period, despite the occurrence of the Mw 6.0 2004 rupture, which supports the interpretation that seismically active and silent patches of the fault reflect differences in rheological and geometrical properties of the fault, and not transient stress variations (Thurber et al., 2006).

4.2 The 2021, Mw 6.0 Antelope Valley sequence

The 2021, Mw 6.0 Antelope Valley, California earthquake occurred in an area of normal-fault bounded basins within the complex Walker Lane zone of trans-tensional deformation, between the Sierra Nevada batholith to the west and the Basin and Range extensional province to the east (Wesnousky, 2005; Wesnousky et al., 2012; Pierce et al., 2021). Estimates for the 2021 mainshock show hypocentral and centroid depths of ~8–10 km and an approximately north-south striking, normal-faulting mechanism with causative fault plane dipping ~45° to the east based on the overall geometry of aftershocks (U.S.G.S., 2017; Pollitz et al., 2022; Wang et al., 2023).

We perform multi-scale precise, NLL-SC relocations (Lomax and Savvaidis, 2021; Lomax and Henry, 2023) for the 2021 Antelope Valley sequence using the earthquakes catalog and phase arrival data for the period 2021-01-01 to 2023-07-14 provided by NCEDC and the Nevada Seismological Laboratory (NSL). For seismicity-stress finite-faulting analysis of the 2021 Antelope Valley sequence we use the first 4 days of relocated aftershocks after the Mw 6.0 mainshock origin (at 2021-07-08 22:48:59 UTC). We use the NSL regional moment tensor (rCMT; Ichinose et al., 2003, available from U.S.G.S., 2017) (nodal planes 1 / 2: strike = 194° / 356°, rake = 42° /

49°, dip = -76° / -102°) for source faulting and the east dipping plane of this mechanism for receiver faults to generate a point-source Δ CFS kernel (Supplementary Figure S3). The kernel has a zero mask radius, R_m , of 0.5 km and decay length, L_m , of 10 km (Supplementary Figure S1). We note here that first-motion mechanisms we obtain for our 2021 Antelope Valley relocations show a mix of normal faulting aftershocks compatible with the mainshock moment tensors and strike-slip aftershocks, where all focal mechanisms share an east-west orientation for the tensional axis. Thus future seismicity-stress analysis of this event should examine the case of using a point-source Δ CFS kernel with strike-slip receiver faults, and also application separately to normal and strike-slip faulting aftershock sets. In Section 5 of this study we examine two cases of receiver fault plane orientation in our analysis of the 2018 Mw 7.1 Anchorage Alaska sequence.

The Antelope Valley seismicity-stress analysis produces the 3D grid of potential, relative, static finite-fault slip shown in Figure 5. The high-potential area of the finite-faulting field (red) forms an ~8 × 3 km, tabular patch at about 6–9 km depth with north-south strike and dip parallel to the east-dipping fault plane of the rCMT and other centroid estimates (U.S.G.S., 2017). The geometry of this high potential slip is constrained primarily by the distribution of seismicity around and mainly above the tabular patch along the east-dipping plane, and secondarily by the shallower seismicity to the east; most of this seismicity falls within positive lobes of the Δ CFS kernel fields (Supplementary Figure S3) for point sources centered within the tabular patch.

The dipping, 3D tabular patch of high-potential, seismicity-stress slip is compatible with rupture on a 2D surface parallel to the east-dipping plane of the rCMT and other centroid estimates for mainshock. However, the vertical position of the patch suggest a rupture plane shifted about 1 km shallower than a similarly oriented plane passing through the relocated hypocenter. This shift could be real, suggesting the hypocenter is not on the main slip plane, or a result of bias in the mainshock hypocenter depth relative to aftershocks depths due difficulty in picking arrivals, especially S, on larger amplitude and longer period large earthquake waveforms (Lomax, 2020). Recall that the seismicity-stress procedure does not require prior specification of the mainshock hypocenter location or a mainshock fault plane depth or orientation, though the specifications of source and receiver faults put constraints on the recovered orientation of volumes of higher mainshock slip.

The high-potential, tabular patch of seismicity-stress finite-faulting matches well the position, extent and orientation of the USGS finite-fault model for the 2021 Antelope Valley mainshock based on teleseismic and regional seismic waveforms (U.S.G.S., 2017), but only shows a general match in overall position to a more complex slip model found by Pollitz et al. (2022) using seismic and geodetic data. The seismicity-stress high-potential patch also matches well the extent and general orientation of higher slip area for the finite-source model of Wang et al. (2023) from joint, kinematic inversion of seismic waveforms, InSAR, and GNSS data

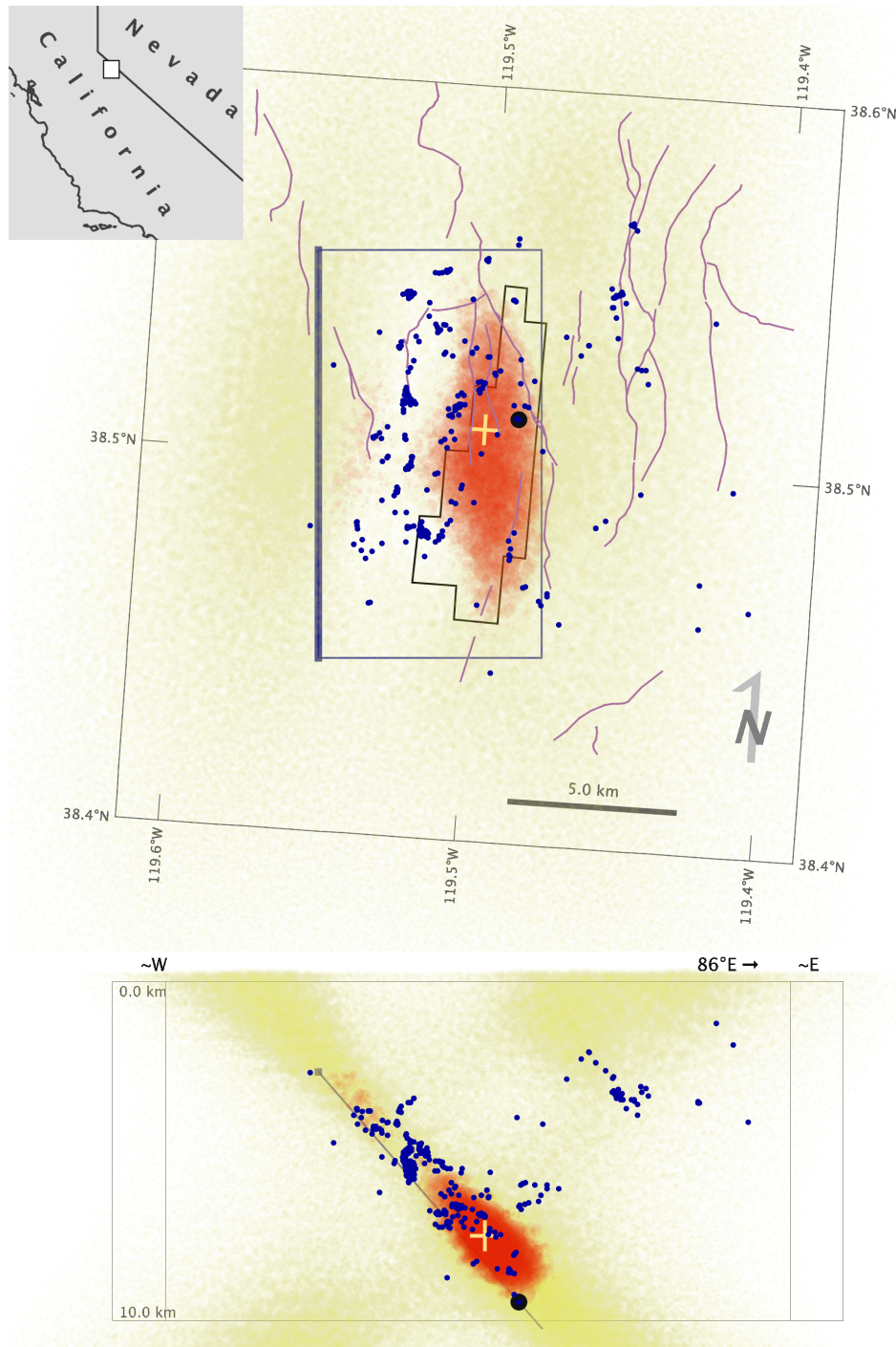


Figure 5 Seismicity-stress, 3D finite-faulting potential slip maps inferred from the first 4 days of aftershocks (blue dots) after the 2021 Antelope Valley mainshock (large black dot). Map (upper) and profile (lower) views are rotated to the strike (4°W) of the NSL rCMT mainshock mechanism (gray rectangle with thick line on upper edge, positioned to intersect the relocated mainshock hypocenter and with arbitrary depth limits). Correlation is performed with a ΔCFS kernel (Supplementary Figure S3) for a normal-faulting point-source specified by the NSL rCMT and receiver faults corresponding to the east-dipping solution of the NSL rCMT. The relative amplitude of the seismicity-stress finite-faulting field is shown as a 3D density cloud in tones of yellow for portions of the field with normalized correlation < 0.5 . The high-potential portions of the field with normalized correlation ≥ 0.5 is clipped to the convex hull of the seismicity and shown as a red density cloud, the yellow cross shows the peak value in this field. Field values are plotted in 3D for each grid cell as transparent disks with radius increasing with correlation value; higher color saturation indicates higher correlation and/or deeper volumes of high correlation. The black polygon outlines higher-slip areas (> 20 cm) of the preferred, kinematic slip model of Wang et al. (2023) based on geodetic and seismic data, shifted 1 km ENE to align their mainshock epicenter with that found here (large black dot). Purple lines show faults from the USGS Quaternary fault and fold database for the United States.

(black polygon in Figure 5). One notable difference between these results is that the peak potential slip for the seismicity-stress model (yellow cross in Figure 5) is about 2 km north of the peak of slip in the Wang et al. (2023) model, the latter being further from the hypocenter, similar to our results for peak slip for the 2004 Parkfield mainshock. However, for two models based only on geodetic data shown by Wang et al. (2023, their figure 9) the area of peak slip overlaps with that of the seismicity-stress maps.

5 Application to the 2018, Mw 7.1 Anchorage, Alaska intra-slab earthquake sequence

The 2018, Mw 7.1 Anchorage, Alaska earthquake (Ruppert and Witter, 2019) occurred within the north-northwest subducting Pacific plate under southern Alaska at an intermediate depth of about 45 km (Figure 6). The earthquake was felt throughout the greater Anchorage area, where it generated the strongest shaking since the great, 1964, M 9.2 Alaska megathrust earthquake, and caused moderate but widespread damage (West et al., 2019). Several point-source moment tensors for the 2018 mainshock (U.S.G.S., 2017) show an approximately north-south striking, normal-faulting mechanism with one fault plane gently dipping $\sim 30^\circ$ to the east and the other steeply dipping $\sim 60^\circ$ to the west, consistent with down-dip extension in the subducting slab. The earthquake produced a highly productive aftershock sequence with the majority of moment tensors for larger aftershocks showing mechanisms similar to the mainshock (Ruppert et al., 2019; West et al., 2019; Drolet et al., 2022).

Aftershocks and finite-faulting models of the 2018 mainshock give ambiguous constraint on which of the fault planes defined by point-source moment tensors hosted mainshock rupture. Some finite-fault models show a weak (U.S.G.S., 2017; Liu et al., 2019) or strong (He et al., 2020) preference for mainshock slip on the steeply west-dipping plane, while it is also suggested that the rupture was simultaneous and conjugate on both planes (Guo et al., 2020). Thus, identification of the causative rupture plane for this earthquake remains an enigma, which we show can be resolved with the combination of multi-scale precise earthquake relocation and seismicity-stress finite-faulting analysis.

We perform multi-scale precise, NLL-SC relocations (Lomax and Savvaidis, 2021; Lomax and Henry, 2023) for seismicity in the area of the 2018 Anchorage sequence for the period 2014-01-01 to 2023-11-25 using catalog events and phase arrival data provided by the Alaska Earthquake Center. The hypocenters for aftershocks of the 2018, Mw 7.1 event (Figure 7) form two, main, northern and southern sub-clusters (Ruppert et al., 2019), as well as surrounding lineations and diffuse patches. The relocated seismicity since 2014 in the greater region around the 2018 sequence (Figure 6, Supplementary Movie S1) shows diffuse but organized seismicity mainly in an $\sim 10 - 20$ km thick, downward-bending tabular zone within the subducting Pacific

plate around and below the general depth of the 2018 sequence. In profile view this seismicity shows a double seismic zone (Hasegawa et al., 1978; Brudzinski et al., 2007) in the descending Pacific slab (Ratchkovsky et al., 1997) under the area of the main 2018 sequence. In map view there is a larger scale organization and moderate clustering of seismicity around north to northeast trending lineations, though these do not show a simple relation to the direction of Pacific – North American plate convergence or to the strike of the dipping slab (see 40 km depth contour of Slab2 in Figure 7). However, along a northwest-southeast trend passing near the 2018 sequence, these lineations exhibit possible changes in orientations and character that could be related to the proximity of the southwest boundary in the subducting slab of the Yakutat microplate (Haeussler et al., 2008), and to a change in strike of the dipping slab apparent in the 40 km depth contour of Slab2. Multi-scale precise relocation of seismicity over a larger area is necessary to verify and better interpret these trends and features in the seismicity.

For seismicity-stress finite-faulting analysis of the 2018 sequence, we use the first day of relocated aftershocks after the Mw 7.1 mainshock origin (at 2018-11-30 17:29:29 UTC). We construct point-source Δ CFS kernels using, for source faulting, the focal-mechanism of the USGS W-phase mainshock moment-tensor (USGS-WCMT; U.S.G.S., 2017), (nodal planes 1 / 2: strike = 6° / 189° , rake = 28° / 189° , dip = -93° / -88°), which is representative of other point-source moment tensor solutions for the mainshock. The kernel has a zero mask radius, R_m , of 1.0 km and decay length, L_m , of 10 km (Supplementary Figure S1).

For the receiver faults we considered two cases: the gently east- and steeply west-dipping fault planes of the USGS-WCMT mechanism. Only the case of west-dipping receiver faults (Δ CFS kernel shown in Supplementary Figure S4) produces a seismicity-stress map (Figure 7, Supplementary Movie S2) with a main patch of highest potential slip (red) that is a good candidate for mainshock rupture: this main patch 1) is tabular, and sub-parallel to and contains one of the USGS-WCMT fault planes through the hypocenter (the east-dipping plane), 2) abuts the mainshock hypocenter, and 3) contains few aftershock hypocenters but is bounded by aftershocks, especially far from the hypocenter and in the hanging wall, extensional quadrant of rupture.

In contrast, the seismicity-stress map for east-dipping receiver faults (Supplementary Figure S5) has a main patch of highest potential slip in a steeply west-dipping, tabular region that is sub-parallel to the west-dipping USGS-WCMT fault plane, but offset about 5 km to the west, is further than 10 km from the mainshock hypocenter, and contains numerous aftershocks, thus not a good candidate for mainshock rupture. This seismicity-stress map does contain a secondary, tabular patch of potential slip that is parallel to the east-dipping USGS-WCMT plane, but this patch is weaker than and floats much higher above the east-dipping fault plane than the main patch obtained with west-dipping receiver faults. For neither of the receiver fault orientations do the seismicity-stress maps show larger poten-

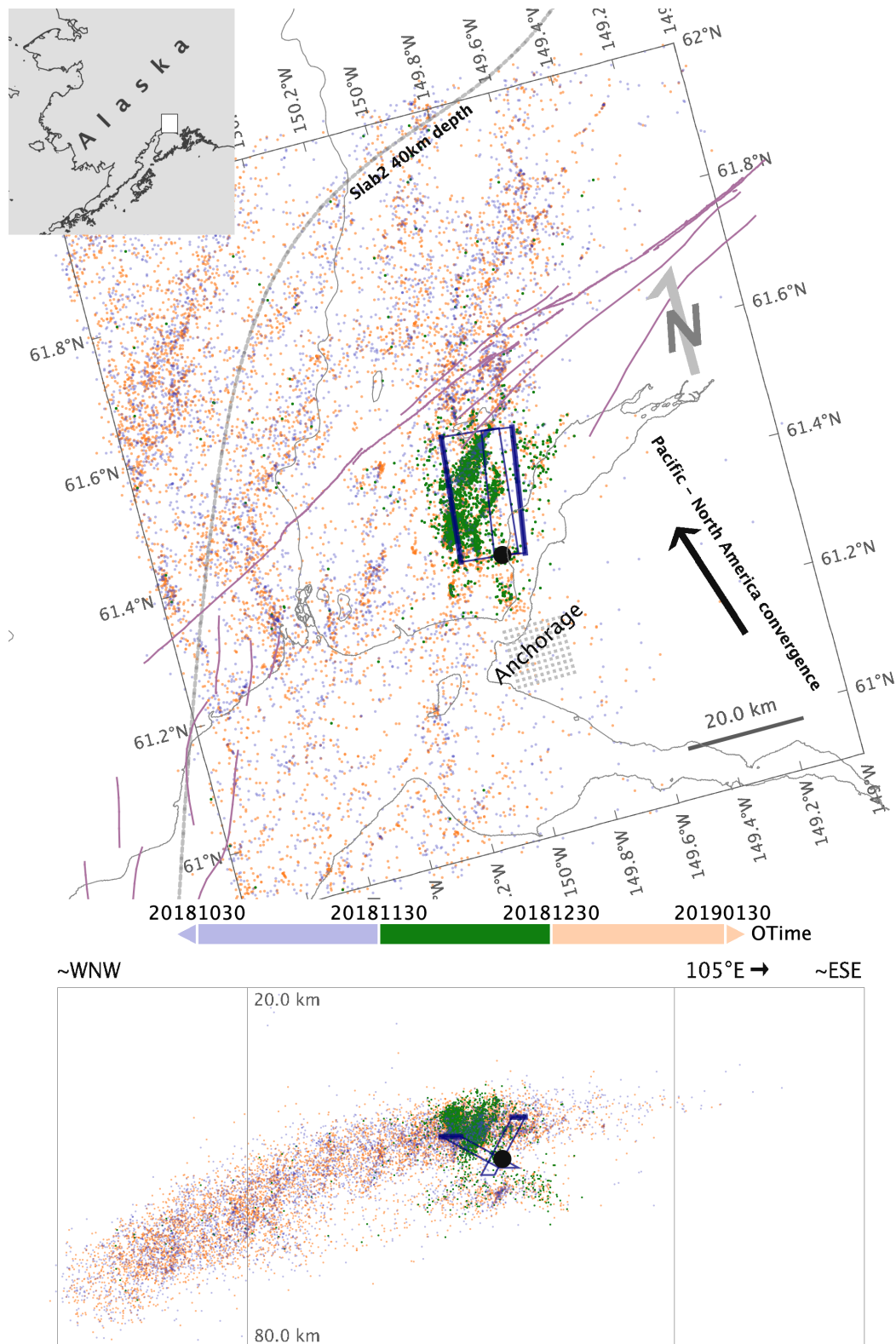


Figure 6 Overview map of the 2018, Mw 7.1 Anchorage, Alaska sequence and background seismicity 2014-2023. Map (upper) and profile (lower) views are rotated to 15°E to roughly match the strike of the plane of dipping slab seismicity and corresponding strike of the Slab2 (Hayes, 2018) 40km depth contour (thick gray line) to the west of the 2018 sequence. Note that the top of the profile view is at 20 km depth. NLL-SC relocated seismicity shown for: 2014 – 2018 mainshock (light blue), 2018 mainshock – 1 month after mainshock (green), 1 month after mainshock through 2023 (light orange); large black dot indicates the Mw 7.1 mainshock hypocenter. The USGS-WCMT mainshock mechanism is shown as blue rectangles with thick line on upper edges, positioned to intersect the relocated mainshock hypocenter and with arbitrary length and upper and lower depth limits. Arrow shows Pacific – North America relative plate convergence direction (Haeussler et al., 2008); purple lines show surface faults from the USGS Quaternary fault and fold database for the United States.

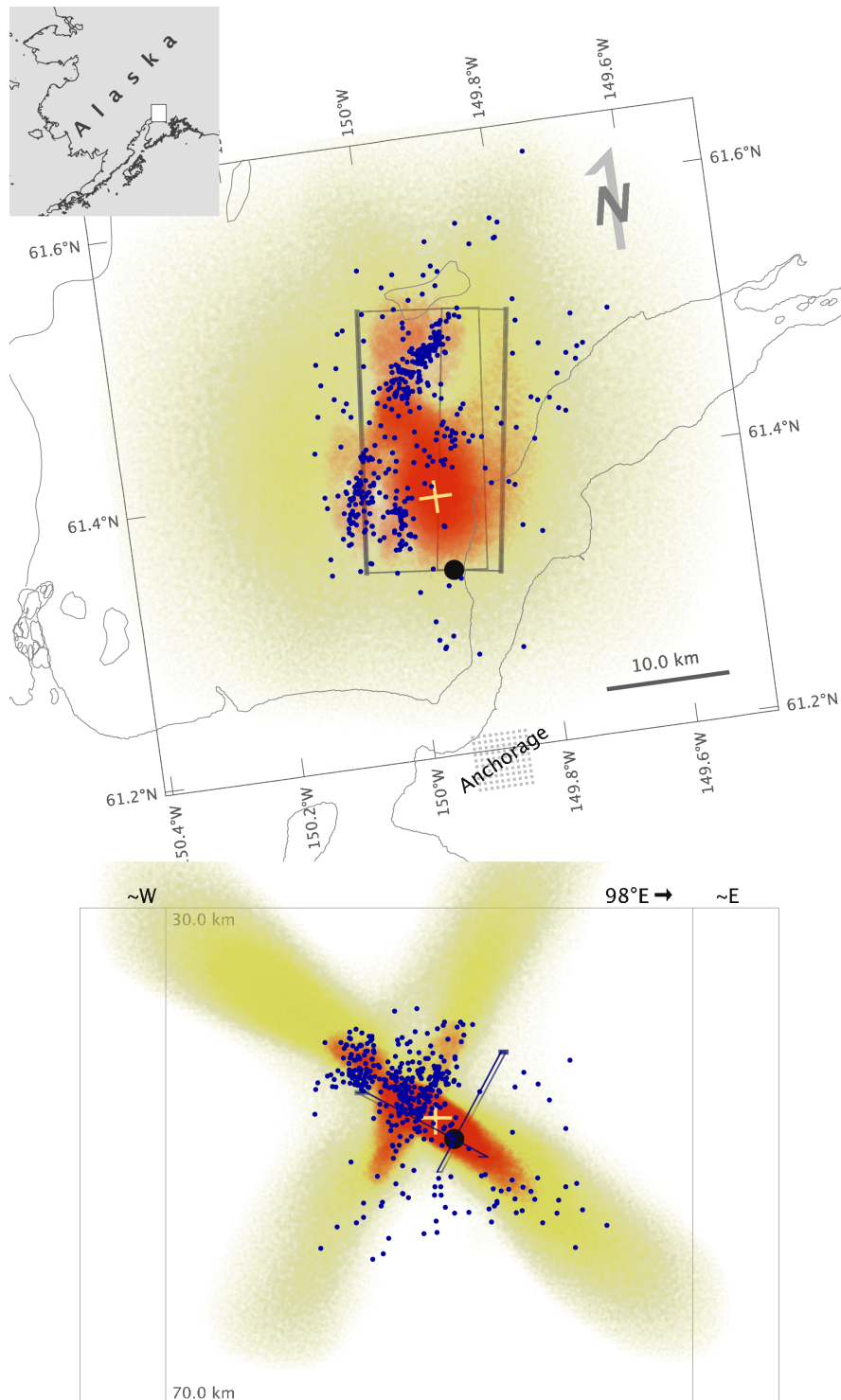


Figure 7 Seismicity-stress, 3D finite-faulting potential slip maps for west-dipping receiver faults inferred from the first 1 day of aftershocks (blue dots) after the 2018 Mw 7.1 Anchorage, Alaska mainshock (large black dot). Map (upper) and profile (lower) views are rotated to the strike of the neutral axis (8°E) of the USGS-WCMT mainshock mechanism (gray rectangles with thick line on upper edges, positioned to intersect the relocated mainshock hypocenter and with arbitrary length and upper and lower depth limits). Correlation is performed with a Δ CFS kernel for the normal-faulting, USGS-WCMT mechanism point-source and for receiver faults corresponding to the west-dipping fault of the USGS-WCMT. The relative amplitude of the seismicity-stress finite-faulting field is shown as a 3D density cloud in tones of yellow for portions of the field with normalized correlation < 0.5 . The high-potential portions of the field with normalized correlation ≥ 0.5 is clipped to the convex hull of the seismicity and shown as a red density cloud, the yellow cross shows the peak value in this field. Field values are plotted in 3D for each grid cell as transparent disks with radius increasing with correlation value; higher color saturation indicates higher correlation and/or deeper volumes of high correlation.

tial slip in a patch overlapping and parallel to the west-dipping USGS-WCMT plane, which is weakly to strongly preferred for mainshock rupture in most previous studies (U.S.G.S., 2017; Liu et al., 2019; Guo et al., 2020; He et al., 2020).

Thus the seismicity-stress analysis strongly suggests a mainshock faulting geometry with slip on the gently east-dipping fault plane, and supports rupture towards the north with peak-slip about 5 km north of the hypocenter, and a main rupture area (slip potential greater than half the peak value) of about 10×10 km (Figure 7). This rupture geometry matches well the non-preferred, joint seismic-geodetic finite-fault inversion of Liu et al. (2019, their supporting information figure S10) for the east-dipping fault model, except that the extent of their main rupture patch (slip greater than half the peak value) is larger at around 20×20 km and their main slip is about 5 km further from the hypocenter than obtained with the seismicity-stress analysis.

6 Discussion

We have introduced a seismicity-stress imaging procedure for 3D mapping of earthquake finite-fault slip (or tensile opening) directly from seismicity and aftershocks. For specified source and receiver faulting, the procedure generate seismicity-stress maps through correlation of point-source Coulomb failure stress change (Δ CFS) kernels with the seismicity around an area of earthquake or other seismic slip. These maps show potential, relative, static fault slip as the envelope of an ensemble of slip solutions compatible with the surrounding seismicity given the physics of the Coulomb failure stress criteria. The reverse mapping of aftershock locations through the non-isotropic lobes of the Δ CFS field allows the method to resolve complex, 3D distributions of finite-fault slip, and makes it different from basic, geometrical interpretation of faulting from aftershock distributions (e.g. Mendoza and Hartzell, 1988).

Application of the seismicity-stress procedure requires multi-scale precise relocations of seismicity surrounding a source zone of earthquake rupture, aseismic slip, fault creep, dyke intrusion or other strain source. Calculation of the point-source, Δ CFS kernels requires specification of a shear dislocation or tensional mechanism for the source, a shear slip mechanism and faulting plane for receiver faults, and distance parameters for masking of the singularity at the Δ CFS point-source. Typical seismicity-stress computational time is small, less than a few minutes on current, standard, desktop workstations.

The time delay for rapid seismicity-stress analysis after a large earthquake depends on the occurrence of and location of sufficient aftershock seismicity. The aftershock time window after the main event should be as short as possible, otherwise the seismicity-stress slip imaging may capture after-slip, as illustrated in Figure 4. However, the imaging quality should increase as the number of aftershocks increases. The Parkfield application in this study uses 4 hours of aftershock seismicity, long after a moment tensor would typically be available, so seismicity-stress imaging can be obtained

in less than a few hours after a mainshock with good station coverage and a productive aftershock sequence. In the near future, high-performance, automated processing using machine-learning seismic phase picking, which can detect several times more events than current automatic picking algorithms (Liu et al., 2020; Cianetti et al., 2021), should allow effective application of the seismicity-stress procedure in even less time after a mainshock.

The seismicity-stress method is based on information from seismicity at the depth of and surrounding the area of target seismic slip, making it qualitatively different from finite-fault inversion procedures which use surface observations. With the dense seismic sensor coverage available in many areas of significant seismicity and earthquake hazard, and precise earthquake location procedures such as SSST (Richards-Dinger and Shearer, 2000) and NLL-SC (Lomax and Savvaidis, 2021; Lomax and Henry, 2023), the relative positions of seismic events across multiple scales can be determined with much smaller error (e.g. 100's of meters) than the typical scale length of interest and resolution in finite-fault mapping (e.g. multi kilometer). Thus, with aftershock or background seismicity that fills an adequate volume around the target slip area, the seismicity-stress analysis can give strong, 3D constraint on potential slip.

Besides using information at depth, the seismicity-stress procedure has additional important differences with other methods for determining finite faulting. The seismicity-stress procedure does not require definition of the position and orientation of a fixed surface as locus for fault slip or fracture opening, nor does it require or make use of a hypocentral location or other initiation point for target slip; instead the procedure maps potential slip throughout the 3D volume covered by the seismicity. This unrestricted 3D mapping provides more freedom for identifying the location and orientation of main slip, as we saw with the case of resolving ambiguous rupture orientation for the 2018 Anchorage earthquake. The procedure also does not require definition of parameters like rupture speed and rise-time, and, as an imaging method, does not involve a formal inversion with its dependence on smoothing constraints or other regularization. However, the seismicity-stress procedure does not directly quantify directions or magnitude of slip and moment release, nor does it produce temporal information on rupture evolution. Given information such as a point-source, centroid moment tensor and elastic properties for a source rupture, it is possible that mean and perhaps variations in slip, stress-drop and other quantities can be derived from the area of high seismicity-stress slip potential. Seismic moment estimates for large earthquakes are now available within minutes after the event occurs (Lomax and Michelini, 2012; Goldberg et al., 2023) and, given a rupture area, moment can be used to determine mean values for slip amplitude and an approximate stress drop (Aki, 1972). Thus slip and stress drop estimates can be obtained from seismicity-stress imaging since it provides information on rupture area, e.g. within the limits of high-potential faulting. Furthermore, the spatial variation of relative seismicity-stress slip may be ef-

fective for determining heterogeneous slip maps and, in consequence, heterogeneous stress drop (Madariaga, 1979). The seismicity-stress procedure can also be used to map migration of areas of slip, after-slip and creep over longer time periods, as we illustrated for the Parkfield stretch of the SAF (Fig. 4).

Though we have only explored shear dislocation sources in this study, the seismicity-stress methodology should be valid for mapping extended sources of tensional faulting or fracture opening, such as volcanic dyke intrusions, as long as these sources produce seismicity on surrounding faults or fractures. The theory for the change in stresses due to a tensile source is well developed (Okada, 1992), and existing algorithms for calculating strain and Δ CFS support these sources (e.g. Toda et al., 2011; Materna and Wong, 2023).

6.1 Application to larger earthquakes and multi-segment rupture

Seismicity-stress imaging can be applied to larger earthquakes and events with complex, multi-segment fault rupture. The main complications are that multiple source fault mechanisms must be tested when the mainshock rupture involves a curved fault or segments with different orientations or slip direction, and multiple receiver faulting mechanisms may be required to represent diverse aftershock mechanisms. A simplified, preliminary investigation of such application for the 2023 Mw 7.8 and Mw 7.5 Kahramanmaraş, Turkey (Türkiye) earthquake sequence (e.g. Goldberg et al., 2023; Karabulut et al., 2023) shows good agreement with previous slip models for the distribution of slip and location of areas of largest slip (Supplementary Text S1). A complete examination of this topic and development of relevant algorithms and tools is left to future work.

6.2 Further applications and uses of seismicity-stress slip maps

There are a number of other potentially significant applications of seismicity-stress analysis and uses for the obtained slip maps. We outline a few of these applications and uses here, without attempting to explore further their implementation or performance.

Seismicity-stress slip maps can provide prior constraint on the number and geometry of faults and orientation of slip for finite-fault inversions that use seismic, geodetic and other surface measures. This procedure would combine the strong, 3D spatial information on slip derived by seismicity-stress from the distribution of seismicity at depth around a rupture with the temporal and more explicitly quantitative information derived by formal inversion methods that use surface data.

Another important use for seismicity-stress maps could be estimation of the spatial likelihood of future seismicity as a component of aftershock forecasting (Steacy et al., 2004). This would be a quasi data-driven and physics based procedure: Aftershocks up to the current time are used to develop a seismicity-stress map, which is an extended distribution of weighted, Δ CFS point-sources. The weighted sum over the Δ CFS from

these point-sources then provides an estimate of the full Δ CFS field for the extended mainshock rupture. This field represents the relative spatial likelihood of future aftershock occurrence, including for areas where aftershocks have not yet occurred.

Seismicity-stress slip maps may also aid in rapid shaking characterization and perhaps tsunami early-warning for large earthquakes when data transmission and basic event processing (e.g. determination of a mainshock mechanism or moment-tensor and relocation of early aftershocks) are available in near-realtime. In this case, the seismicity-stress slip information could be integrated into algorithms that estimate shaking intensity or model tsunami generation and propagation from finite faulting models. Depending on instrumental coverage and the aftershock productivity of a seismic sequence, seismicity-stress slip maps might be available within a few hours of the main event, as illustrated in the Parkfield application in this study.

A particularly consequential application of seismicity-stress analysis would be to obtain finite-faulting information for large earthquakes that occurred before the advent of modern seismic, geodetic and space-based networks. Any event in the early-instrumental period (e.g. before the 1960's (Villasenor and Engdahl, 2007)) for which a faulting mechanism can be established and for which there are reliably located aftershocks in and around the source region (e.g. Tape and Lomax, 2022) is a candidate for such analysis. For the largest events, aftershocks located with epicentral uncertainty much less than the extent of the expected rupture zone, and with minimal depth constraint or depth constrained by independent information (e.g. known megathrust depth profiles, recent seismicity, or regional tectonics) might qualify as reliably located.

Lastly, seismicity-stress analysis could be applied to foreshock sequences to identify possible regions of aseismic slip or creep that trigger the foreshocks through Coulomb stress transfer, and thus aid in distinguishing between cascading and pre-slip models of earthquake nucleation (Ellsworth and Beroza, 1995; Mignan, 2014). Seismicity-stress mapping using foreshock seismicity might resolve concentrated areas of aseismic slip active before a mainshock in a manner analogous to the seismicity-stress mapping of creep related slip and after-slip over longer time periods for the Parkfield application in this study. A lack of identification of concentrated pre-slip might provide supporting evidence for a cascading process for a foreshock sequence.

6.3 Validation of the seismicity-stress procedure for mapping 3D finite-fault slip

We verified in Section 3 that the seismicity-stress procedure correctly recovers the location and form of two patches of synthetic, double-couple slip on a rectangular fault, given a random sample of aftershocks distributed according to the theoretical Δ CFS field for the extended slip source (Figure 2). For clarity, this test uses a somewhat ideal distribution of seismicity that statisti-

cally follows the complete ΔCFS field. In practice, a heterogeneous distribution of ambient stresses, of existing faults and fractures, and of geologic units with varying mineralogy, heat flow, fabric, rheology and other properties will also affect strongly the distribution of seismicity (Collettini et al., 2009; Hardebeck, 2022; Hardebeck and Harris, 2022).

For a more realistic test of the seismicity-stress procedure we computed slip potential maps for the 2004 Mw 6.0 Parkfield CA and 2021 Mw 6.0 Antelope Valley CA earthquakes and compared them with published finite-fault inversions and other independent information on these events. For Parkfield the seismicity-stress analysis using vertical, strike-slip source and receiver faults and 4 hours of aftershocks recovers highest slip potential concentrated along a vertical plane, as expected for this stretch of the SAF (Figure 3). For Antelope Valley the analysis using 4 days of aftershocks and normal faulting source and receiver faults recovers concentrated slip potential around a dipping plane that matches the causative rupture plane expected from aftershocks and mainshock centroid moment-tensor solutions (Figure 5). Both of these analyses produce maps of highest slip potential that match well the distribution of larger slip obtained from finite-fault inversions, though the seismicity-stress results place peak slip ~5 km closer to the mainshock hypocenter for Parkfield and ~2 km closer to the hypocenter for Antelope Valley. We discuss this discrepancy in more detail below.

For Parkfield we also examined seismicity-stress maps for a number of time windows of up to two decades before and after the 2004 mainshock (Figure 5 and Supplementary Figure S2) These maps show possible migration of after-slip around the co-seismic rupture area in the weeks and months after the mainshock. They also show highest slip potential in the decades before the mainshock falls as expected in the well documented, fast-creeping stretch of the SAF just northwest of the 2004 and 1966 Parkfield rupture zone, and a slow recovery towards a similar slip distribution over the two decades after the 2004 mainshock. This temporal analysis illustrates use of the seismicity-stress procedure for longer term mapping of source areas of main co-seismic, post-seismic, creep related, and perhaps other types of slip.

6.4 Discrepancy in location of highest-slip patch

For the 2004, Mw 6.0 Parkfield sequence the seismicity-stress procedure maps a co-seismic finite-faulting field (Figure 3) that closely follows the vertical SAF, with areas of high potential slip that match well the distribution of main slip obtained by other methods, given their uncertainties. The patch of highest seismicity-stress slip potential locates about 11 km northwest of the 2004 hypocenter, which agrees with the location of principal sources from inversions of high-frequency seismic radiation, but not with most seismic and GNSS based slip inversions which find highest slip 5 km or more further to the northwest.

This discrepancy in position of maximum slip could

be due to shortcomings in the seismicity-stress procedure or available aftershocks locations. However, seismicity-stress slip maps for background seismicity in the period 1984-2004 and from 2014 to 2023 (Figure 4 and Supplementary Figure S2) show a southeast limit zone of slip potential for the creeping section of the SAF around 15 – 20 km northwest of the 2004 hypocenter which is coincident with the northwest limit of the patch of highest, 2004 co-seismic slip potential. This abutting of slip regions suggests little or no slip deficit would be available for significant 2004 rupture as far as 20 km northwest of the hypocenter. In this case, the discrepancy in location of highest-slip patch could be due to the lack of GNSS sites to the southeast of the rupture zone (Houlié et al., 2014), or a bias in peak slip location in seismic waveform inversions.

For the 2021, Mw 6.0 Antelope Valley mainshock we found a similar, though smaller, discrepancy in peak slip location, with the seismicity-stress peak located ~2 km closer to the hypocenter than found through joint inversion of seismic waveforms and geodetic data (Wang et al., 2023). However, in this case the location of the seismicity-stress peak corresponds well to that of inversion by Wang et al. (2023) with geodetic data only.

A shift in peak slip away from the hypocenter with broadband seismic waveform inversion, that is not present with high-frequency seismic inversion or geodetic inversion (nor with seismicity-stress mapping), suggests a bias unique to broadband seismic waveform inversion. Perhaps such slip inversion is biased towards imaging sources of high-amplitude, broadband stopping phases related to rupture arrest and rupture approaching the free surface (Savage, 1965; Madariaga, 1977; Madariaga and Ruiz, 2016), and not towards imaging areas of strongest total moment release, which, if smooth, may radiate mainly low amplitude, long period waves (Madariaga, 1977) that might not be observable. Similarly, if the true rupture is crack-like with a long rise time and strong, late slip near the middle of the fault (Madariaga, 1977), or involves reversal of the direction of the slip front (e.g. Hicks et al., 2020), then inversions for which these cases are not allowed in the modeled rupture evolution might produce erroneous mapping of slip too far from the hypocenter. There could also be shifts in peak slip location due to directivity or other near-source waveform effects (Archuleta and Hartzell, 1981) if they, for example, produce augmented, high-amplitude signal late on the waveforms for seismic sensors located above the fault or in the direction of rupture which are incorrectly mapped to high-amplitude slip near these sensors.

6.5 Sensitivity of seismicity-stress imaging to faulting parameters and stress components

To provide some insight into the sensitivity and variation of seismicity-stress imaging with respect to some of the numerous parameters controlling the procedure, we first show changes in imaging for the 2004, Mw 6.0 Parkfield sequence with respect to changes in the strike, dip and rake of the source and receiver faults (Sup-

plementary Figures S7-9). The Parkfield sequence slip maps change notably in intensity but not in overall form with a $\pm 10^\circ$ change in strike, with an asymmetry in the variation due to slight asymmetry of the positive lobes of the ΔCFS kernel (Figure 1). The slip maps change notably in intensity and in form with a $\pm 20\text{--}40^\circ$ change in rake, and change slightly in intensity but little in form with up to a 40° reduction in dip.

The sensitivity of seismicity-stress imaging for the Parkfield sequence with respect to changes in the effective coefficient of friction, μ' , (Equation 2) is shown in Supplementary Figures S10. The Parkfield sequence slip maps change very slightly in form and the position of the slip peak with change in μ' .

We also examine the contribution to the imaging of the different stress components of the Coulomb failure stress, ΔCFS , (Equation 2) used in generating the point-source kernel. We consider three stress types: *coulomb* (Equation 2) used for the analyses in this study, *shear* (the first term on the right of Equation 2), and *normal* (the second term on the right of Equation 2). For the 2004, Mw 6.0 Parkfield sequence (Supplementary Figure S11) there are only small differences in the stress type *shear* slip map relative to *coulomb* since for the Parkfield strike-slip geometry (Figure 1) almost all aftershocks fall along a line parallel to the maximum of the positive shear lobes, while there is effectively no slip recovered for stress type *normal* since the positive *normal* lobes are perpendicular to the trend of aftershocks (c.f. King et al., 1994, their fig. 2a). The contributions of cross-correlation of the kernel moving along aftershocks aligned along a positive kernel lobe will locally produce a strong, constructive sum to total slip, while the contributions of aftershocks aligned perpendicular to positive lobes will tend to be cancelled out by adjacent, negative lobes.

For the 2021, Mw 6.0 Antelope Valley sequence (Supplementary Figure S12) there are only small differences in the stress type *shear* slip map relative to *coulomb*, while the stress type *normal* map shows almost no high-potential slip, especially around the likely fault plane and aftershock seismicity. However, the *normal* kernel must still contribute enough to perturbing the *coulomb* kernel lobes to account for the difference in *coulomb* and *shear* maps.

These preliminary results on the contribution of different stress components of ΔCFS suggest that the *shear* component in Equation 2 greatly dominates over the *normal* component for imaging slip and thus for explaining the aftershock distribution.

In addition to the issue of sensitivity to faulting and other parameters, the seismicity-stress methodology as presented here assumes aftershocks or other seismicity is primarily due to quasi-static stress change due to dislocation or crack opening in an elastic medium. The methodology does not explicitly consider stress changes and possible resulting seismicity due to dynamic, poroelastic, and viscoelastic effects, although application of the methodology over consecutive time windows (e.g. Figure 4 and Supplementary Figure S2) can give some information on these sources of stress changes. There is much analysis, discussion and con-

troversy on all these issues, as well as on the numerous assumptions and simplifications leading to the formulation of Coulomb stress change in Equation 2 (see, e.g. Wang et al., 2014, and references therein). Future work on the seismicity-stress imaging should address some of these issues.

6.6 The 2018, Mw 7.1 Anchorage, Alaska intra-slab earthquake sequence

We generated seismicity-stress maps for the 2018, Mw 7.1 Anchorage sequence using the representative, USGS-WCMT mainshock mechanism to define the Coulomb point-source faulting and each of the fault planes from this mechanism to define two cases of receiver fault orientation and slip. The resulting pair of potential slip maps and relocated seismicity (Figure 7 and Supplementary Figure S5) strongly support the case of mainshock rupture on a gently east-dipping plane with most aftershocks involving normal slip on planes parallel to the steeply west-dipping, USGS-WCMT plane. The seismicity-stress maps and relocated aftershock seismicity do not support mainshock rupture on a steeply west-dipping plane.

The potential slip maps are mainly constrained by concentrations of aftershock activity along the four, positive lobes of the point-source ΔCFS kernel (Figure 7 and Supplementary Figure S5, Supplementary Movies S1 and S2) as it is displaced along the resolved, high slip regions. This constraint includes numerous aftershocks in the hanging wall and extensional quadrant of rupture, and also two lobes of aftershock seismicity under sequence extending into the lower part of the double seismic zone (Figures 6 and 7). We can thus attribute the activation of this deeper aftershock seismicity to the two down-going positive lobes of the Coulomb stress field from extended mainshock rupture.

Our relocated seismicity shows many aftershocks in northern and southern sub-clusters on or near an $\sim 20\text{km}$ long, $\sim 8\text{ km}$ vertical, steeply northwest-dipping plane which strikes about 40°E (Figure 8, Supplementary Movies S1 and S2), markedly clockwise to the $\sim 8^\circ\text{E}$ strike of the USGS-WCMT planes. Seismicity along this plane is also apparent within the lineations in the pre- and post- sequence background seismicity (Figures 6 and 8). Other nearby aftershocks fall on shorter, oblique or conjugate trends on the east side of this 40°E striking plane, the ensemble suggesting a larger scale, perhaps rhomboidal fracture system. Drolet et al. (2022, their figure 6) identify some events near the 40°E striking plane with a moment-tensor fault plane sub-parallel to the plane, but abundant seismicity with slip on this plane seems incompatible with the seismicity-stress result that most aftershock slip should occur on normal faults sub-parallel to the west-dipping USGS-WCMT plane. This discrepancy may be related to the nature of brittle fracture in slabs at intermediate depth, perhaps involving volume processes such as dehydration embrittlement or thermal shear instability (Raleigh and Paterson, 1965; Hobbs and Ord, 1988; Kirby et al., 1996; Hasegawa and Nakajima, 2017). For example, background and aftershock activity may occur on smaller

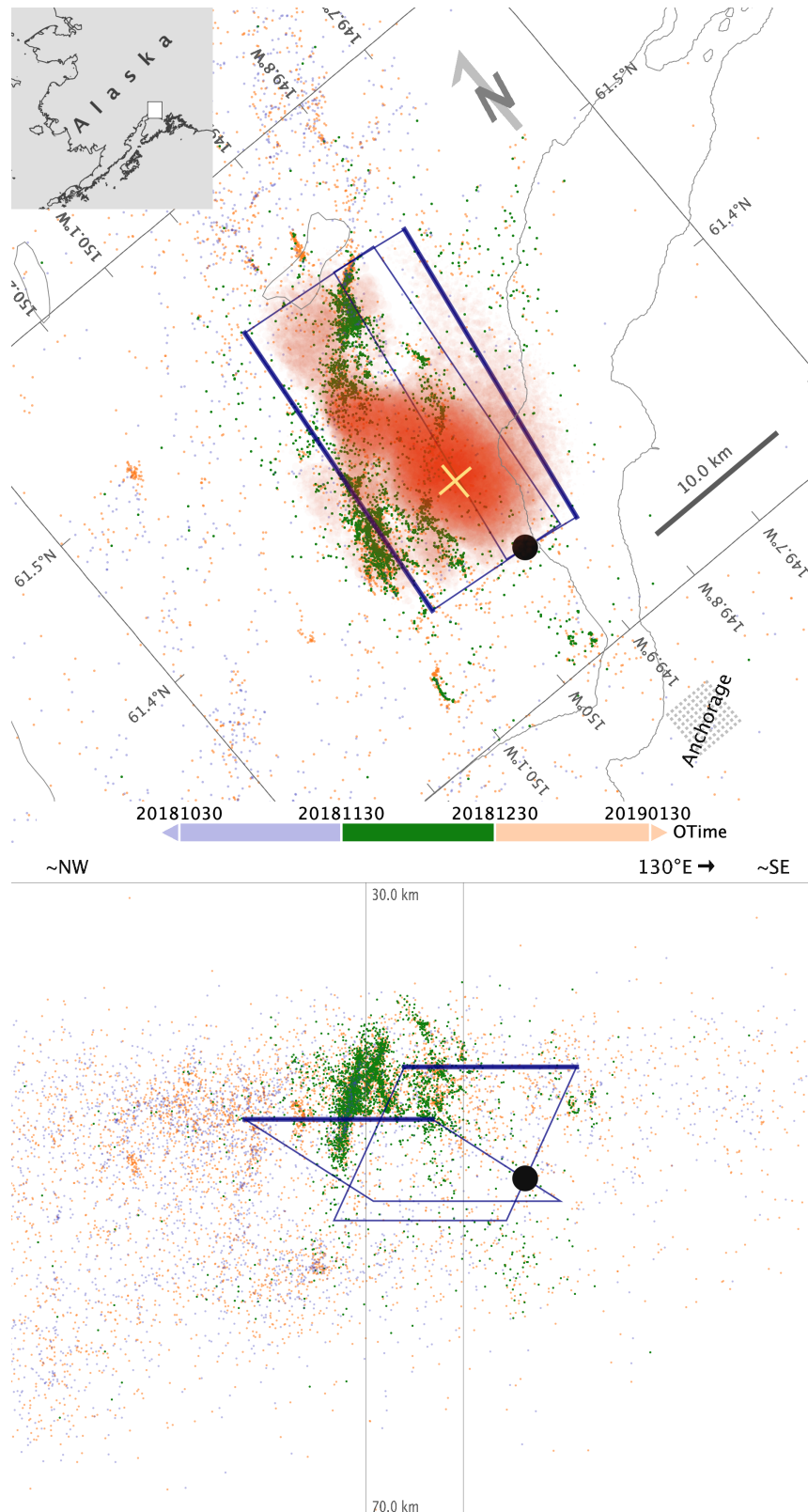


Figure 8 Detail map of the 2018 Mw 7.1 Anchorage, Alaska sequence and background seismicity 2014-2023. Map (upper) and profile (lower) views are rotated to align along the steeply northwest dipping plane of aftershocks striking 40°E. Note that the top of the profile view is at 30 km depth. NLL-SC relocated seismicity shown for: 2014 to 2018 mainshock (light blue), 2018 mainshock to 1 month after mainshock (green), 1 month after mainshock through 2023 (light orange); large black dot indicates the Mw 7.1 mainshock hypocenter. The USGS-WCMT mainshock mechanism is shown as blue rectangles with thick line on upper edges, positioned to intersect the relocated mainshock hypocenter and with arbitrary length and upper and lower depth limits. The high-potential portions of the preferred seismicity-stress finite-faulting field from Figure 7 is shown in red in the map view, the yellow cross shows the peak value in this field.

scales as a complex volume-filling process in which individual slip planes are oblique to larger scale trends of seismicity, in analogy to crustal processes such as wrench tectonics (Anderson, 1905; Sibson et al., 2012) or fault-fracture meshes (Sibson, 1996). In this case the 40°E trend of aftershocks might not define a surface of active faulting, but instead a western limit to a zone of brittle failure, perhaps composed of fractures, faults and anomalous rheology formed during earlier plate bending at the outer rise or plate formation at the spreading ridge.

The seismicity along and east of the 40°E striking plane appears rooted on the east-dipping, mainshock rupture plane, up-dip from the main slip and extending up to ~7 km above this plane (Figures 7 and 8). Thus much of the aftershock activity occurs off the main rupture plane in complex faulting in the hanging wall, and mainshock rupture (red field in Figure 8) occurs below a mainly aseismic gap between zones of aftershock seismicity, as previously proposed by Ruppert et al. (2019).

7 Conclusions

We have introduced a seismicity-stress imaging procedure for 3D mapping of finite-fault slip or tensile opening directly from seismicity and aftershocks following the physics of the Coulomb failure stress criteria. As this observed seismicity occurs at depth around an earthquake rupture or other seismic source, it provides strong constraint on the 3D distribution of potential, static fault slip that could cause the seismicity. The seismicity-stress procedure is fairly simple and requires few assumptions; it does not need specification of the location and orientation of a 2D surface as locus for source slip, only requiring a faulting mechanism for the source. Seismicity-stress maps may be useful as prior constraint for other slip inversion procedures, for quasi data-driven, physics based aftershock forecasting, and for rapid shaking characterization and perhaps tsunami early-warning. The seismicity-stress procedure should be applicable to map finite-faulting for early instrumental earthquakes, to search for aseismic slip or creep during foreshock sequences, and to other problems.

We verify that seismicity-stress maps correctly recover two patches of synthetic fault slip on a vertical fault, and match well the results from other inversion for finite-fault slip for the 2004 Mw 6.0 Parkfield CA and 2021 Mw 6.0 Antelope Valley CA earthquakes. We find indication of a discrepancy between locations of peak slip found with the seismicity-stress procedure and seismic and GNSS based slip inversions. It is not clear if this discrepancy reflects a limitation of the seismicity-stress procedure or of other slip inversion procedures.

Our seismicity-stress analysis of the 2018, Mw 7.1 Anchorage sequence shows normal-faulting mainshock slip on a gently south-east dipping plane, resolving the ambiguity in causative fault plane given by other finite-faulting analyses. The seismicity-stress map and multi-scale precise hypocenter relocations for the 2018 sequence show abundant aftershock activity in the hanging wall and extensional quadrant of mainshock rupture. Most aftershocks concentrate in a volume along

and east of a 40°E striking plane, suggesting a rhomboidal fabric perhaps controlled by earlier plate bending and faulting. The analysis indicates two diffuse lobes of aftershocks descending into the lower part of a double, slab seismic zone are likely due to Coulomb stress triggering from mainshock rupture.

Acknowledgments

I gratefully thank Pierre Henry and Kathryn Materna for discussion and comments that greatly improved this work, and to Jean Paul Ampuero for pointing out the relation between cross-correlation and adjoint operator imaging. The comments and suggestions of two anonymous reviewers and the editor Ake Fagereng allowed much improvement and clarification of the manuscript. A special thanks to all Seismica volunteers for their help and contributions and to the field crews, analysts, technicians and scientists who enabled the high-quality NCEDC, NSL and Alaska Earthquake Center arrival-time and event catalogs which made the precision of this work possible. LibreOffice (<https://www.libreoffice.org>) is used for word processing, spreadsheet calculations and drawings, and Zotero (<https://www.zotero.org>) for citation management. This work was supported by the author's personal funds.

8 Data and code availability

The supplementary information for this article includes Figure S1-12 and Text S1. Also, in a repository (<https://zenodo.org/doi/10.5281/zenodo.10939981>), Movies S1 and S2 showing animated views of relocated seismicity and seismicity-stress results for the 2018 Mw 7.1 Anchorage sequence, and CSV and NLL-Hypocenter format catalogs of NLL-SSST-coherence relocations used in this study.

Earthquake catalogs and corresponding phase arrival times, waveforms and metadata were accessed through the Northern California Earthquake Data Center (NCEDC), <http://doi.org/10.7932/NCEDC>; information produced by the Alaska Earthquake Center was retrieved from the ANSS Comprehensive Catalog operated by the U.S. Geological Survey (U.S.G.S., 2017). Fault trace data are from <https://www.usgs.gov/programs/earthquake-hazards/faults>.

Software for running seismicity-stress imaging and examples from the three sequences examined in this study are available at: https://github.com/alomax/seismicity_stress

Earthquake relocations were performed with NonLinLoc (Lomax et al., 2001, 2014, <http://www.alomax.net/nllloc>, <https://github.com/ut-beg-texnet/NonLinLoc>) following the workflow presented in (Lomax and Savvaidis, 2021). Stress calculations were performed with Elastic_stresses_py (Materna and Wong, 2023, https://github.com/kmaterna/Elastic_stresses_py). SeismicityViewer (<http://www.alomax.net/software>) was used for 3D seismicity analysis and plotting, ObsPy (Beyreuther et al., 2010; Krischer et al., 2015, <http://obs.py.org>) for reading and processing seismicity catalogs and for coherence calculations.

9 Competing interests

The author declares having no competing interests.

References

- Aki, K. Earthquake mechanism. *Tectonophysics*, 13(1–4):423–446, Apr. 1972. doi: 10.1016/0040-1951(72)90032-7.
- Allmann, B. P. and Shearer, P. M. A High-Frequency Secondary Event During the 2004 Parkfield Earthquake. *Science*, 318(5854): 1279–1283, Nov. 2007. doi: 10.1126/science.1146537.
- Anderson, E. The dynamics of faulting. *Transactions of the Edinburgh Geological Society*, 8(3):387–402, Jan. 1905. doi: 10.1144/transed.8.3.387.
- Archuleta, R. J. and Hartzell, S. H. Effects of fault finiteness on near-source ground motion. *Bulletin of the Seismological Society of America*, 71(4):939–957, Aug. 1981. doi: 10.1785/bssa0710040939.
- Bakun, W. H., Aagaard, B., Dost, B., Ellsworth, W. L., Hardebeck, J. L., Harris, R. A., Ji, C., Johnston, M. J. S., Langbein, J., Lienkaemper, J. J., Michael, A. J., Murray, J. R., Nadeau, R. M., Reasenber, P. A., Reichle, M. S., Roeloffs, E. A., Shakal, A., Simpson, R. W., and Waldhauser, F. Implications for prediction and hazard assessment from the 2004 Parkfield earthquake. *Nature*, 437(7061):969–974, Oct. 2005. doi: 10.1038/nature04067.
- Barchi, M., Carboni, F., Michele, M., Ercoli, M., Giorgetti, C., Porreca, M., Azzaro, S., and Chiaraluce, L. The influence of subsurface geology on the distribution of earthquakes during the 2016–2017 Central Italy seismic sequence. *Tectonophysics*, 807: 228797, May 2021. doi: 10.1016/j.tecto.2021.228797.
- Beresnev, I. A. Uncertainties in Finite-Fault Slip Inversions: To What Extent to Believe? (A Critical Review). *Bulletin of the Seismological Society of America*, 93(6):2445–2458, Dec. 2003. doi: 10.1785/0120020225.
- Beyreuther, M., Barsch, R., Krischer, L., Megies, T., Behr, Y., and Wassermann, J. ObsPy: A Python Toolbox for Seismology. *Seismological Research Letters*, 81(3):530–533, May 2010. doi: 10.1785/gssrl.81.3.530.
- Bos, A. G. and Spakman, W. The resolving power of coseismic surface displacement data for fault slip distribution at depth. *Geophysical Research Letters*, 30(21), Nov. 2003. doi: 10.1029/2003gl017946.
- Brudzinski, M. R., Thurber, C. H., Hacker, B. R., and Engdahl, E. R. Global Prevalence of Double Benioff Zones. *Science*, 316(5830): 1472–1474, June 2007. doi: 10.1126/science.1139204.
- Bruhat, L., Barbot, S., and Avouac, J.-P. Evidence for postseismic deformation of the lower crust following the 2004 Mw6.0 Parkfield earthquake. *Journal of Geophysical Research*, 116(B8), Aug. 2011. doi: 10.1029/2010jb008073.
- Budiman, R., Sahara, D. P., and Nugraha, A. D. Determining Source Model and Aftershocks of 2006 Yogyakarta Earthquake, Indonesia using Coulomb Stress Change. *IOP Conference Series: Earth and Environmental Science*, 318(1):012026, Aug. 2019. doi: 10.1088/1755-1315/318/1/012026.
- Chinnery, M. A. The stress changes that accompany strike-slip faulting. *Bulletin of the Seismological Society of America*, 53(5): 921–932, Oct. 1963. doi: 10.1785/bssa0530050921.
- Cianetti, S., Bruni, R., Gaviano, S., Keir, D., Piccinini, D., Saccorotti, G., and Giunchi, C. Comparison of Deep Learning Techniques for the Investigation of a Seismic Sequence: An Application to the 2019, Mw 4.5 Mugello (Italy) Earthquake. *Journal of Geophysical Research: Solid Earth*, 126(12), Dec. 2021. doi: 10.1029/2021jb023405.
- Claerbout, J. *Earth soundings analysis : processing versus inversion*. Blackwell Scientific Publications, 1992.
- Claerbout, J. *Basic Earth Imaging*. 2010.
- Collettini, C., Niemeijer, A., Viti, C., and Marone, C. Fault zone fabric and fault weakness. *Nature*, 462(7275):907–910, Dec. 2009. doi: 10.1038/nature08585.
- Coulomb, C. Essai sur une application des regles de maximis et minimis a quelques problemes de statique relatifs a l'architecture. *Mem. Div. Sav. Acad*, 1773.
- Drolet, D., Bostock, M. G., Plourde, A. P., and Sammis, C. G. Aftershock distributions, moment tensors and stress evolution of the 2016 Iniskin and 2018 Anchorage Mw 7.1 Alaskan intraslab earthquakes. *Geophysical Journal International*, 231(1): 199–214, Apr. 2022. doi: 10.1093/gji/ggac165.
- Ellsworth, W. L. and Beroza, G. C. Seismic Evidence for an Earthquake Nucleation Phase. *Science*, 268(5212):851–855, May 1995. doi: 10.1126/science.268.5212.851.
- Fletcher, J., Spudich, P., and Baker, L. Rupture Propagation of the 2004 Parkfield, California, Earthquake from Observations at the UPSAR. *Bulletin of the Seismological Society of America*, 96(4B): S129–S142, Sept. 2006. doi: 10.1785/0120050812.
- Font, Y., Kao, H., Lallemand, S., Liu, C.-S., and Chiao, L.-Y. Hypocentre determination offshore of eastern Taiwan using the Maximum Intersection method. *Geophysical Journal International*, 158(2):655–675, Aug. 2004. doi: 10.1111/j.1365-246x.2004.02317.x.
- Fountoulakis, I. and Evangelidis, C. P. SSA2py: A High-Performance Python Implementation of the Source-Scanning Algorithm for Spatiotemporal Seismic Source Imaging. *Seismological Research Letters*, 95(4):2506–2518, Mar. 2024. doi: 10.1785/0220230335.
- Freed, A. M. Afterslip (and only afterslip) following the 2004 Parkfield, California, earthquake. *Geophysical Research Letters*, 34(6), Mar. 2007. doi: 10.1029/2006gl029155.
- Fukahata, Y., Yagi, Y., and Rivera, L. Theoretical relationship between back-projection imaging and classical linear inverse solutions. *Geophysical Journal International*, 196(1):552–559, Oct. 2013. doi: 10.1093/gji/ggt392.
- Gahalaut, V., Kalpna, and Raju, P. Rupture mechanism of the 1993 Killari earthquake, India: constraints from aftershocks and static stress change. *Tectonophysics*, 369(1–2):71–78, July 2003. doi: 10.1016/s0040-1951(03)00135-5.
- Goldberg, D. E., Taymaz, T., Reitman, N. G., Hatem, A. E., Yolsal-Çevikbilen, S., Barnhart, W. D., Irmak, T. S., Wald, D. J., Öcalan, T., Yeck, W. L., Özkan, B., Thompson Jobe, J. A., Shelly, D. R., Thompson, E. M., DuRoss, C. B., Earle, P. S., Briggs, R. W., Benz, H., Erman, C., Doğan, A. H., and Altuntaş, C. Rapid Characterization of the February 2023 Kahramanmaraş, Türkiye, Earthquake Sequence. *The Seismic Record*, 3(2):156–167, Apr. 2023. doi: 10.1785/0320230009.
- Guo, Y., Miyakoshi, K., and Tsurugi, M. Simultaneous rupture on conjugate faults during the 2018 Anchorage, Alaska, intraslab earthquake (MW 7.1) inverted from strong-motion waveforms. *Earth, Planets and Space*, 72(1), Nov. 2020. doi: 10.1186/s40623-020-01315-x.
- Haeussler, P. J., Freymueller, J. T., Wesson, R. L., and Ekström, G. An Overview of the Neotectonics of Interior Alaska: Far-Field Deformation from the Yakutat Microplate Collision, page 83–108. American Geophysical Union, Mar. 2008. doi: 10.1029/179gm05.
- Hardebeck, J. L. Physical Properties of the Crust Influence Aftershock Locations. *Journal of Geophysical Research: Solid Earth*, 127(10), Oct. 2022. doi: 10.1029/2022jb024727.
- Hardebeck, J. L. and Harris, R. A. Earthquakes in the Shadows: Why

- Aftershocks Occur at Surprising Locations. *The Seismic Record*, 2(3):207–216, July 2022. doi: 10.1785/0320220023.
- Harris, R. A. Introduction to Special Section: Stress Triggers, Stress Shadows, and Implications for Seismic Hazard. *Journal of Geophysical Research: Solid Earth*, 103(B10):24347–24358, Oct. 1998. doi: 10.1029/98jb01576.
- Harris, R. A. and Segall, P. Detection of a locked zone at depth on the Parkfield, California, segment of the San Andreas Fault. *Journal of Geophysical Research: Solid Earth*, 92(B8):7945–7962, July 1987. doi: 10.1029/jb092ib08p07945.
- Harris, R. A. and Simpson, R. W. In the shadow of 1857—the effect of the Great Ft. Tejon Earthquake on subsequent earthquakes in southern California. *Geophysical Research Letters*, 23(3):229–232, Feb. 1996. doi: 10.1029/96gl00015.
- Hartzell, S., Liu, P., Mendoza, C., Ji, C., and Larson, K. M. Stability and Uncertainty of Finite-Fault Slip Inversions: Application to the 2004 Parkfield, California, Earthquake. *Bulletin of the Seismological Society of America*, 97(6):1911–1934, Dec. 2007. doi: 10.1785/0120070080.
- Hartzell, S. H. and Heaton, T. H. Inversion of strong ground motion and teleseismic waveform data for the fault rupture history of the 1979 Imperial Valley, California, earthquake. *Bulletin of the Seismological Society of America*, 73(6A):1553–1583, Dec. 1983. doi: 10.1785/bssa07306a1553.
- Hasegawa, A. and Nakajima, J. Seismic imaging of slab metamorphism and genesis of intermediate-depth intraslab earthquakes. *Progress in Earth and Planetary Science*, 4(1), Apr. 2017. doi: 10.1186/s40645-017-0126-9.
- Hasegawa, A., Umino, N., and Takagi, A. Double-planed structure of the deep seismic zone in the northeastern Japan arc. *Tectonophysics*, 47(1–2):43–58, May 1978. doi: 10.1016/0040-1951(78)90150-6.
- Haskell, N. A. Total energy and energy spectral density of elastic wave radiation from propagating faults. *Bulletin of the Seismological Society of America*, 54(6A):1811–1841, Dec. 1964. doi: 10.1785/bssa05406a1811.
- Hayes, G. Slab2 - A Comprehensive Subduction Zone Geometry Model, 2018.
- He, P., Wen, Y., Chen, Y., Xu, C., and Ding, K. 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. *Earth and Space Science*, 7(1), Jan. 2020. doi: 10.1029/2019ea000924.
- Hicks, S. P., Okuwaki, R., Steinberg, A., Rychert, C. A., Harmon, N., Abercrombie, R. E., Bogiatzis, P., Schlaphorst, D., Zahradnik, J., Kendall, J.-M., Yagi, Y., Shimizu, K., and Sudhaus, H. Back-propagating supershear rupture in the 2016 Mw 7.1 Romanche transform fault earthquake. *Nature Geoscience*, 13(9):647–653, Aug. 2020. doi: 10.1038/s41561-020-0619-9.
- Hobbs, B. E. and Ord, A. Plastic instabilities: Implications for the origin of intermediate and deep focus earthquakes. *Journal of Geophysical Research: Solid Earth*, 93(B9):10521–10540, Sept. 1988. doi: 10.1029/jb093ib09p10521.
- Houlié, N., Dreger, D., and Kim, A. GPS source solution of the 2004 Parkfield earthquake. *Scientific Reports*, 4(1), Jan. 2014. doi: 10.1038/srep03646.
- Ichinose, G., Anderson, J., Smith, K., and Zeng, Y. Source Parameters of Eastern California and Western Nevada Earthquakes from Regional Moment Tensor Inversion. *Bulletin of the Seismological Society of America*, 93(1):61–84, Feb. 2003. doi: 10.1785/0120020063.
- Ide, S. 4.07 - Slip Inversion, page 193–223. Elsevier, 2007. doi: 10.1016/b978-044452748-6.00068-7.
- Johanson, I., Fielding, E., Rolandone, F., and Burgmann, R. Coseismic and Postseismic Slip of the 2004 Parkfield Earthquake from Space-Geodetic Data. *Bulletin of the Seismological Society of America*, 96(4B):S269–S282, Sept. 2006. doi: 10.1785/0120050818.
- Jolivet, R., Simons, M., Agram, P. S., Duputel, Z., and Shen, Z. Aseismic slip and seismogenic coupling along the central San Andreas Fault. *Geophysical Research Letters*, 42(2):297–306, Jan. 2015. doi: 10.1002/2014gl062222.
- Karabulut, H., Guvercin, S. E., Hollingsworth, J., and Konca, A. O. Long silence on the East Anatolian Fault Zone (Southern Turkey) ends with devastating double earthquakes (6 February 2023) over a seismic gap: implications for the seismic potential in the Eastern Mediterranean region. *Journal of the Geological Society*, 180(3), May 2023. doi: 10.1144/jgs2023-021.
- Kawakatsu, H. and Montagner, J.-P. Time-reversal seismic-source imaging and moment-tensor inversion. *Geophysical Journal International*, 175(2):686–688, Nov. 2008. doi: 10.1111/j.1365-246x.2008.03926.x.
- Kikuchi, M. and Kanamori, H. Inversion of complex body waves. *Bulletin of the Seismological Society of America*, 72(2):491–506, Dec. 1982. doi: https://doi.org/10.1785/BSSA0720020491.
- Kim, A. and Dreger, D. S. Rupture process of the 2004 Parkfield earthquake from near-fault seismic waveform and geodetic records. *Journal of Geophysical Research: Solid Earth*, 113(B7), July 2008. doi: 10.1029/2007jb005115.
- King, G., Stein, R., and Lin, J. Static stress changes and the triggering of earthquakes. *Bulletin of the Seismological Society of America*, 84:935–953, 1994. doi: 10.1785/BSSA0840030935.
- Kirby, S., Engdahl, R. E., and Denlinger, R. *Intermediate-Depth Intraslab Earthquakes and Arc Volcanism as Physical Expressions of Crustal and Uppermost Mantle Metamorphism in Subducting Slabs*, page 195–214. American Geophysical Union, Mar. 1996. doi: 10.1029/gm096p0195.
- Krischer, L., Megies, T., Barsch, R., Beyreuther, M., Lecocq, T., Caudron, C., and Wassermann, J. ObsPy: a bridge for seismology into the scientific Python ecosystem. *Computational Science & Discovery*, 8(1):014003, May 2015. doi: 10.1088/1749-4699/8/1/014003.
- Langbein, J., Borchardt, R., Dreger, D., Fletcher, J., Hardebeck, J. L., Hellweg, M., Ji, C., Johnston, M., Murray, J. R., Nadeau, R., Rymer, M. J., and Treiman, J. A. Preliminary Report on the 28 September 2004, M 6.0 Parkfield, California Earthquake. *Seismological Research Letters*, 76(1):10–26, Jan. 2005. doi: 10.1785/gssrl.76.1.10.
- Langbein, J., Murray, J., and Snyder, H. Coseismic and Initial Postseismic Deformation from the 2004 Parkfield, California, Earthquake, Observed by Global Positioning System, Electronic Distance Meter, Creepmeters, and Borehole Strainmeters. *Bulletin of the Seismological Society of America*, 96(4B):S304–S320, Sept. 2006. doi: 10.1785/0120050823.
- Liu, C., Lay, T., Xie, Z., and Xiong, X. Intraslab Deformation in the 30 November 2018 Anchorage, Alaska, MW 7.1 Earthquake. *Geophysical Research Letters*, 46(5):2449–2457, Mar. 2019. doi: 10.1029/2019gl082041.
- Liu, J., Sieh, K., and Hauksson, E. A Structural Interpretation of the Aftershock “Cloud” of the 1992 Mw 7.3 Landers Earthquake. *Bulletin of the Seismological Society of America*, 93(3):1333–1344, June 2003. doi: 10.1785/0120020060.
- Liu, M., Zhang, M., Zhu, W., Ellsworth, W. L., and Li, H. Rapid Characterization of the July 2019 Ridgecrest, California, Earthquake Sequence From Raw Seismic Data Using Machine-Learning Phase Picker. *Geophysical Research Letters*, 47(4), Feb. 2020.

- doi: 10.1029/2019gl086189.
- Liu, P., Custódio, S., and Archuleta, R. Kinematic Inversion of the 2004 M 6.0 Parkfield Earthquake Including an Approximation to Site Effects. *Bulletin of the Seismological Society of America*, 96 (4B):S143–S158, Sept. 2006. doi: 10.1785/0120050826.
- Lomax, A. A Reanalysis of the Hypocentral Location and Related Observations for the Great 1906 California Earthquake. *Bulletin of the Seismological Society of America*, 95(3):861–877, June 2005. doi: 10.1785/0120040141.
- Lomax, A. Location of the Focus and Tectonics of the Focal Region of the California Earthquake of 18 April 1906. *Bulletin of the Seismological Society of America*, 98(2):846–860, Apr. 2008. doi: 10.1785/0120060405.
- Lomax, A. Absolute Location of 2019 Ridgecrest Seismicity Reveals a Shallow Mw 7.1 Hypocenter, Migrating and Pulsing Mw 7.1 Foreshocks, and Duplex Mw 6.4 Ruptures. *Bulletin of the Seismological Society of America*, 110(4):1845–1858, May 2020. doi: 10.1785/0120200006.
- Lomax, A. and Henry, P. Major California faults are smooth across multiple scales at seismogenic depth. *Seismica*, 2(1), June 2023. doi: 10.26443/seismica.v2i1.324.
- Lomax, A. and Michelini, A. Tsunami Early Warning Within Five Minutes. *Pure and Applied Geophysics*, 170(9–10):1385–1395, July 2012. doi: 10.1007/s00024-012-0512-6.
- Lomax, A. and Savvaidis, A. High-Precision Earthquake Location Using Source-Specific Station Terms and Inter-Event Waveform Similarity. *Journal of Geophysical Research: Solid Earth*, 127(1), Dec. 2021. doi: 10.1029/2021jb023190.
- Lomax, A., Virieux, J., Volant, P., and Berge-Thierry, C. *Probabilistic Earthquake Location in 3D and Layered Models*, page 101–134. Springer Netherlands, 2000. doi: 10.1007/978-94-015-9536-0_5.
- Lomax, A., Zollo, A., Capuano, P., and Virieux, J. Precise, absolute earthquake location under Somma-Vesuvius volcano using a new three-dimensional velocity model. *Geophysical Journal International*, 146(2):313–331, Aug. 2001. doi: 10.1046/j.0956-540x.2001.01444.x.
- Lomax, A., Michelini, A., and Curtis, A. *Earthquake Location, Direct, Global-Search Methods*, page 1–33. Springer New York, 2014. doi: 10.1007/978-3-642-27737-5_150-2.
- Madariaga, R. High-frequency radiation from crack (stress drop) models of earthquake faulting. *Geophysical Journal International*, 51(3):625–651, Dec. 1977. doi: 10.1111/j.1365-246x.1977.tb04211.x.
- Madariaga, R. On the relation between seismic moment and stress drop in the presence of stress and strength heterogeneity. *Journal of Geophysical Research: Solid Earth*, 84(B5):2243–2250, May 1979. doi: 10.1029/jb084ib05p02243.
- Madariaga, R. and Ruiz, S. Earthquake dynamics on circular faults: a review 1970–2015. *Journal of Seismology*, 20(4):1235–1252, June 2016. doi: 10.1007/s10950-016-9590-8.
- Mai, P. M., Schorlemmer, D., Page, M., Ampuero, J., Asano, K., Causse, M., Custodio, S., Fan, W., Festa, G., Galis, M., Galovic, F., Imperatori, W., Käser, M., Malysky, D., Okuwaki, R., Pollitz, F., Passone, L., Razafindrakoto, H. N. T., Sekiguchi, H., Song, S. G., Somala, S. N., Thingbaijam, K. K. S., Twardzik, C., van Driel, M., Vyas, J. C., Wang, R., Yagi, Y., and Zielke, O. The Earthquake-Source Inversion Validation (SIV) Project. *Seismological Research Letters*, 87(3):690–708, Apr. 2016. doi: 10.1785/0220150231.
- Marone, C. and Scholz, C. H. The depth of seismic faulting and the upper transition from stable to unstable slip regimes. *Geophysical Research Letters*, 15(6):621–624, June 1988. doi: 10.1029/gl015i006p00621.
- Materna, K. and Wong, J. Elastic stresses py: v1.0.0, 2023.
- Maurer, J. and Johnson, K. Fault coupling and potential for earthquakes on the creeping section of the central San Andreas Fault. *Journal of Geophysical Research: Solid Earth*, 119(5):4414–4428, May 2014. doi: 10.1002/2013jb010741.
- McEvilly, T. V., Bakun, W. H., and Casaday, K. B. The Parkfield, California, earthquakes of 1966. *Bulletin of the Seismological Society of America*, 57(6):1221–1244, Dec. 1967. doi: 10.1785/bssa0570061221.
- Mendoza, C. and Hartzell, S. Aftershock patterns and main shock faulting. *Bulletin of the Seismological Society of America*, 78: 1438–1449, 1988.
- Mignan, A. The debate on the prognostic value of earthquake foreshocks: A meta-analysis. *Scientific Reports*, 4(1), Feb. 2014. doi: 10.1038/srep04099.
- Minson, S. E., Simons, M., and Beck, J. L. Bayesian inversion for finite fault earthquake source models I—theory and algorithm. *Geophysical Journal International*, 194(3):1701–1726, June 2013. doi: 10.1093/gji/ggt180.
- Nadeau, R. M. and McEvilly, T. V. Periodic Pulsing of Characteristic Microearthquakes on the San Andreas Fault. *Science*, 303(5655): 220–222, Jan. 2004. doi: 10.1126/science.1090353.
- Okada, Y. Internal deformation due to shear and tensile faults in a half-space. *Bulletin of the Seismological Society of America*, 82 (2):1018–1040, Apr. 1992. doi: 10.1785/bssa0820021018.
- Olson, A. H. and Apsel, R. J. Finite faults and inverse theory with applications to the 1979 Imperial Valley earthquake. *Bulletin of the Seismological Society of America*, 72(6A):1969–2001, Dec. 1982. doi: 10.1785/bssa07206a1969.
- Oppenheimer, D. H., Reasenber, P. A., and Simpson, R. W. Fault plane solutions for the 1984 Morgan Hill, California, Earthquake Sequence: Evidence for the state of stress on the Calaveras Fault. *Journal of Geophysical Research: Solid Earth*, 93(B8): 9007–9026, Aug. 1988. doi: 10.1029/jb093ib08p09007.
- Pierce, I. K. D., Wesnousky, S. G., Owen, L. A., Bormann, J. M., Li, X., and Caffee, M. Accommodation of Plate Motion in an Incipient Strike-Slip System: The Central Walker Lane. *Tectonics*, 40(2), Feb. 2021. doi: 10.1029/2019tc005612.
- Pollitz, F. F., Wicks, C. W., and Hammond, W. C. Kinematic Slip Model of the 2021 M 6.0 Antelope Valley, California, Earthquake. *The Seismic Record*, 2(1):20–28, Jan. 2022. doi: 10.1785/0320210043.
- Raleigh, C. B. and Paterson, M. S. Experimental deformation of serpentinite and its tectonic implications. *Journal of Geophysical Research*, 70(16):3965–3985, Aug. 1965. doi: 10.1029/jz070i016p03965.
- Ratchkovsky, N. A., Pujol, J., and Biswas, N. N. Stress pattern in the double seismic zone beneath Cook Inlet, south-central Alaska. *Tectonophysics*, 281(3–4):163–171, Nov. 1997. doi: 10.1016/s0040-1951(97)00042-5.
- Reid, H. and Lawson, A. The California earthquake of April 18, 1906: report of the State Earthquake Investigation Commission, in two volumes and atlas, 1908. Retrieved from.
- Richards-Dinger, K. B. and Shearer, P. M. Earthquake locations in southern California obtained using source-specific station terms. *Journal of Geophysical Research: Solid Earth*, 105(B5): 10939–10960, May 2000. doi: 10.1029/2000jb900014.
- Ruppert, N. A. and Witter, R. C. Preface to the Focus Section on the 30 November 2018 Mw 7.1 Anchorage, Alaska, Earthquake. *Seismological Research Letters*, 91(1):16–18, Dec. 2019. doi: 10.1785/0220190344.
- Ruppert, N. A., Nayak, A., Thurber, C., and Richards, C. After-shock Analysis of the 2018 Mw 7.1 Anchorage, Alaska, Earth-

- quake: Relocations and Regional Moment Tensors. *Seismological Research Letters*, 91(1):114–125, Nov. 2019. doi: 10.1785/0220190199.
- Sato, T., Hiratsuka, S., and Mori, J. Coulomb stress change for the normal-fault aftershocks triggered near the Japan Trench by the 2011 M w 9.0 Tohoku-Oki earthquake. *Earth, Planets and Space*, 64(12):1239–1243, Dec. 2012. doi: 10.5047/eps.2012.04.003.
- Savage, J. C. The stopping phase on seismograms. *Bulletin of the Seismological Society of America*, 55(1):47–58, Feb. 1965. doi: 10.1785/bssa0550010047.
- Savage, J. C. and Burford, R. O. Geodetic determination of relative plate motion in central California. *Journal of Geophysical Research*, 78(5):832–845, Feb. 1973. doi: 10.1029/jb078i005p00832.
- Scholz, C. H. *The Mechanics of Earthquakes and Faulting*. Cambridge University Press, Dec. 2018. doi: 10.1017/9781316681473.
- Seeber, L. and Armbruster, J. G. Earthquakes as beacons of stress change. *Nature*, 407(6800):69–72, Sept. 2000. doi: 10.1038/35024055.
- Sibson, R. Fault zone models, heat flow, and the depth distribution of earthquakes in the continental crust of the United States. *Bulletin of the Seismological Society of America*, 72:151–163, 1982. doi: 10.1785/BSSA0720010151.
- Sibson, R. H. Structural permeability of fluid-driven fault-fracture meshes. *Journal of Structural Geology*, 18(8):1031–1042, Aug. 1996. doi: 10.1016/0191-8141(96)00032-6.
- Sibson, R. H., Ghisetti, F. C., and Crookbain, R. A. Andersonian wrench faulting in a regional stress field during the 2010–2011 Canterbury, New Zealand, earthquake sequence. *Geological Society, London, Special Publications*, 367(1):7–18, Aug. 2012. doi: 10.1144/sp367.2.
- Simpson, R., Barall, M., Langbein, J., Murray, J., and Rymer, M. San Andreas Fault Geometry in the Parkfield, California, Region. *Bulletin of the Seismological Society of America*, 96(4B):S28–S37, Sept. 2006. doi: 10.1785/0120050824.
- Steacy, S., Marsan, D., Nalbant, S. S., and McCloskey, J. Sensitivity of static stress calculations to the earthquake slip distribution. *Journal of Geophysical Research: Solid Earth*, 109(B4), Apr. 2004. doi: 10.1029/2002jb002365.
- Stein, R. and Lisowski, M. The 1979 Homestead Valley Earthquake Sequence, California: Control of aftershocks and postseismic deformation. *Journal of Geophysical Research: Solid Earth*, 88: 6477–6490, 1983. doi: 10.1029/JB088iB08p06477.
- Stein, R. S. The role of stress transfer in earthquake occurrence. *Nature*, 402(6762):605–609, Dec. 1999. doi: 10.1038/45144.
- Steinbrugge, K. V., Zacher, E. G., Tocher, D., Whitten, C. A., and Claire, C. N. Creep on the San Andreas fault. *Bulletin of the Seismological Society of America*, 50(3):389–415, July 1960. doi: 10.1785/bssa0500030389.
- Tape, C. and Lomax, A. Aftershock Regions of Aleutian-Alaska Megathrust Earthquakes, 1938–2021. *Journal of Geophysical Research: Solid Earth*, 127(7), July 2022. doi: 10.1029/2022jb024336.
- Thurber, C., Zhang, H., Waldhauser, F., Hardebeck, J., Michael, A., and Eberhart-Phillips, D. Three-Dimensional Compressional Wavespeed Model, Earthquake Relocations, and Focal Mechanisms for the Parkfield, California, Region. *Bulletin of the Seismological Society of America*, 96(4B):S38–S49, Sept. 2006. doi: 10.1785/0120050825.
- Titus, S., DeMets, C., and Tikoff, B. Thirty-Five-Year Creep Rates for the Creeping Segment of the San Andreas Fault and the Effects of the 2004 Parkfield Earthquake: Constraints from Alignment Arrays, Continuous Global Positioning System, and Creepmeters. *Bulletin of the Seismological Society of America*, 96(4B): S250–S268, Sept. 2006. doi: 10.1785/0120050811.
- Toda, S., Stein, R. S., Reasenberg, P. A., Dieterich, J. H., and Yoshida, A. Stress transferred by the 1995 Mw = 6.9 Kobe, Japan, shock: Effect on aftershocks and future earthquake probabilities. *Journal of Geophysical Research: Solid Earth*, 103(B10):24543–24565, Oct. 1998. doi: 10.1029/98jb00765.
- Toda, S., Lin, J., and , R. S. Using the 2011 M w 9.0 off the Pacific coast of Tohoku Earthquake to test the Coulomb stress triggering hypothesis and to calculate faults brought closer to failure. *Earth, Planets and Space*, 63(7):725–730, July 2011. doi: 10.5047/eps.2011.05.010.
- Twardzik, C., Madariaga, R., Das, S., and Custódio, S. Robust features of the source process for the 2004 Parkfield, California, earthquake from strong-motion seismograms. *Geophysical Journal International*, pages no–no, Oct. 2012. doi: 10.1111/j.1365-246x.2012.05653.x.
- U.S.G.S. Advanced National Seismic System (ANSS) Comprehensive Catalog, 2017.
- Villasenor, A. and Engdahl, E. R. Systematic Relocation of Early Instrumental Seismicity: Earthquakes in the International Seismological Summary for 1960–1963. *Bulletin of the Seismological Society of America*, 97(6):1820–1832, Dec. 2007. doi: 10.1785/0120060118.
- Wang, J., Xu, C., Freymueller, J. T., Li, Z., and Shen, W. Sensitivity of Coulomb stress change to the parameters of the Coulomb failure model: A case study using the 2008 Mw 7.9 Wenchuan earthquake. *Journal of Geophysical Research: Solid Earth*, 119 (4):3371–3392, Apr. 2014. doi: 10.1002/2012jb009860.
- Wang, K., Dreger, D. S., Tinti, E., Bürgmann, R., and Taira, T. Rupture Process of the 2019 Ridgecrest, California Mw 6.4 Foreshock and Mw 7.1 Earthquake Constrained by Seismic and Geodetic Data. *Bulletin of the Seismological Society of America*, 110(4): 1603–1626, July 2020. doi: 10.1785/0120200108.
- Wang, K., Dreger, D. S., Burgmann, R., and Taira, T. Finite-Source Model of the 8 July 2021 M 6.0 Antelope Valley, California, Earthquake. *Seismological Research Letters*, Jan. 2023. doi: 10.1785/0220220262.
- Wang, L., Hainzl, S., Zöller, G., and Holschneider, M. Stress- and aftershock-constrained joint inversions for coseismic and post-seismic slip applied to the 2004 M6.0 Parkfield earthquake. *Journal of Geophysical Research: Solid Earth*, 117(B7), July 2012. doi: 10.1029/2011jb009017.
- Weiss, J., Pellissier, V., Marsan, D., Arnaud, L., and Renard, F. Cohesion versus friction in controlling the long-term strength of a self-healing experimental fault. *Journal of Geophysical Research: Solid Earth*, 121(12):8523–8547, Dec. 2016. doi: 10.1002/2016jb013110.
- Wesnousky, S. G. Active faulting in the Walker Lane. *Tectonics*, 24 (3), June 2005. doi: 10.1029/2004tc001645.
- Wesnousky, S. G., Bormann, J. M., Kreemer, C., Hammond, W. C., and Brune, J. N. Neotectonics, geodesy, and seismic hazard in the Northern Walker Lane of Western North America: Thirty kilometers of crustal shear and no strike-slip? *Earth and Planetary Science Letters*, 329–330:133–140, May 2012. doi: 10.1016/j.epsl.2012.02.018.
- West, M. E., Bender, A., Gardine, M., Gardine, L., Gately, K., Haeusler, P., Hassan, W., Meyer, F., Richards, C., Ruppert, N., Tape, C., Thornley, J., and Witter, R. The 30 November 2018 Mw 7.1 Anchorage Earthquake. *Seismological Research Letters*, 91(1): 66–84, Oct. 2019. doi: 10.1785/0220190176.
- Wong, J. W. C., Fan, W., and Gabriel, A. A Quantitative Comparison and Validation of Finite-Fault Models: The 2011 Tohoku-Oki

Earthquake. *Journal of Geophysical Research: Solid Earth*, 129 (10), Oct. 2024. doi: 10.1029/2024jb029212.

Zhou, H. Rapid three-dimensional hypocentral determination using a master station method. *Journal of Geophysical Research: Solid Earth*, 99(B8):15439–15455, Aug. 1994. doi: 10.1029/94jb00934.

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