**Supporting Information File for manuscript entitled “Is the eastern Denali Fault still active?”
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Text DR1. Data and Velocity Model

The distribution of seismograph stations used in this study are mapped in Fig. DR-1. These stations provide a nearly complete azimuthal distribution around the epicentral region of the St. Elias earthquake sequence.

Figure DR-1. Map of stations used in this study. Red symbols show earthquakes scaled by magnitude indicate the location of the St. Elias earthquake sequence.

Waveform data used in this study were downloaded from the IRIS Data Center. The crust model used in the calculation of moment tensor inversion and double-difference relocation is CRUST1.0 (Laske et al., 2013). CRUST1.0 provides the P and S wave velocity model in the crust at each grid node as 1-D layers. There are 7 layers in the models, each of which represents water, ice, sediments, upper crust, middle crust, lower crust and mantle from the top to the bottom. Velocity models from four different coordinates in CRUST1.0 were used to provide a pseudo-3D model for the study area. The stations are divided into four groups (Fig. DR-2). The median coordinate of each group was chosen as the input coordinates to CRUST1.0.

Figure DR-2. Four velocity models from CRUST1.0 that were used in this study; Vp: P-wave velocity, Vs: S-wave velocity, ρ : density. Stations (blue triangles) in each annotated region share the same model.

Text DR2. Hierarchical Clustering

Clustering analysis of seismicity distributions enables refinement of potential structures that can be discerned in the data. Clustering is defined as an unsupervised classification of observations into groups (clusters) (Jain et al. 1999, Eaton 2018). Clustering has been widely used in microseismic, induced seismic and aftershock interpretations (e.g. Nadeau & Johnson 1998, Igarashi et al. 2003, Schaff & Richards 2004).

This study employs an hierarchical clustering method based on waveform similarity. The algorithm is implemented as a combination of cross-correlation and hierarchical clustering concepts. The first part of the algorithm is to quantify the degree of similarity between waveforms using the normalized correlation coefficient. After correlation time series are calculated for a user-defined time window centered on the event phase picks, the maximum value is chosen. For N events, this results in a $N \times N$ correlation matrix, $\mathbf{S}$. The matrix is then transformed to a distance matrix that represents the closeness of one event with another in a

range of 0 to 1, which can be simply done by $\mathbf{I}-\mathbf{S}$, where $\mathbf{I}$ is the identity matrix. This distance matrix becomes the criterion for the hierarchical clustering. A distance threshold of 0.7 was selected to sub-divide the events up into clusters. Among the extensive clusters, only those composed of more than 20 events and those having an average correlation value more than 0.6 were chosen.

The clustering method used in this study is a connectivity-based clustering algorithm where agglomerative clustering is used initially, followed by divisive clustering. Python modules SciPy and ObsPy were used for cross-correlation and clustering. The process is represented in a dendrogram where the y-axis marks the distance at which the clusters merge, and the events are placed along the x-axis. Clusters continue merging until the distance becomes 1, the largest distance, at which all events ultimately are bound to one tree. The resulting dendrogram provides an extensive hierarchy of clusters. Divisive clustering is applied by cutting the tree based on a defined distance threshold.

The seismicity catalog (Table DR-1) lists locations, times and magnitudes of the 1,314 events considered in this study, along with clustering statistics including the hierarchical cluster ID. Figure DR-3a shows the distribution of aftershocks, with individual clusters represented by distinct symbols. The clusters are grouped, based on spatial location of events, into those that occur near faults 1 and 2, as well as events that occur in the region between the two faults (Figure 3). Figure DR-3b shows the temporal evolution of seismicity in each of these fault regions. Figure DR-4 shows the cross-correlation matrix, $\mathbf{S}$, as well as a dendrogram showing the hierarchical relationships between cluster.

Figure DR-3. a) Relocated aftershock epicenters of the St. Elias earthquake sequence, with symbols that represent the hierarchical cluster IDs for the largest 20 clusters (based on the number of events). Each cluster contains events with similar waveforms, as indicated by cross-correlation analysis. The clusters are grouped such that those plotted in green are interpreted to be linked to the fault that slipped during the first mainshock, those plotted in red are linked to the fault that slipped during the second mainshock, and those plotted in cyan are located between the two mainshocks as shown in Figure 3b in the main manuscript. Magenta symbols show later activity on fault 2b) Temporal distribution of aftershocks, using the same colors as in part (a).

Figure DR-4. The process of hierarchical clustering based on waveform similarity. Left: Maximum correlation coefficient matrix S for 1,314 events, gathered by waveform similarity. Red squares show representative clusters. Right: Hierarchical clustering results presented in a dendrogram.

Text DR3. Moment-tensor inversion

A moment tensor is a mathematical representation of seismic radiation patterns (Aki and Richards, 2002). In order to understand physical processes at the source, a moment tensor is commonly decomposed into double couple (DC), isotropic (ISO) and a compensated-linear-vector-dipole (CLVD) components, where DC represents a slip on a planar fault, ISO represents a volumetric change in the source region (Knopoff and Randall, 1970). The CLVD is variously interpreted as multiple near-synchronous DC components or complex deformation during faulting or magmatic processes (Stein and Wyssession, 2011). The DC component provides two possible fault planes, where the true fault plane can often be discerned using aftershocks, the occurrence of which is a manifestation of localized stress relaxation of a fault (Eaton, 2018). Moment tensors of the mainshock doublet and 5 aftershocks were calculated by the inversion of full waveforms using algorithms in Computer Programs in Seismology (Herrmann, 2013), which is based on grid search and generalized least-squares methods.

To compute moment tensors for each event, ten-minute long windows of seismic waveform data were obtained from IRIS. Preprocessing, such as instrumental response removal and linear trend removal, was applied to the continuous raw data. For each station, a time window was selected to assure all the stations contain the full waveform and an identical number of sampling points. Accordingly, about six-minute-long waveforms were cut from the data. Next, arrival times of the P wave were manually picked, and horizontal components of the seismograms were rotated into transverse and radial components. Both observed and synthetic data were preconditioned through the same procedure in order to obtain a good signal-to-noise ratio and consistency. High and low-frequency noise was removed by applying a 0.025-0.075Hz bandpass filter. Data windows were then trimmed to 5s before and 6min after the P arrival to include the meaningful data in that frequency range. Finally, each data window was interpolated and re-sampled with a unified sampling rate.

In the second step, synthetic waveforms were matched with the observed data. The synthetic waveforms were generated as preconditioned Green's functions using the waveform integration method (Herrmann, 2013). The moment tensor parameters that minimize the sum of squared residuals between the synthetic and the observed data are selected after searching a range of focal depths and moment tensor coefficients. Since the Green's functions were generated from flat-layered models, they do not account for the complicated propagation phenomena associated with lateral heterogeneity in the Earth's crust. To mitigate, this, the Cut-and-Paste (CAP) technique was used, which applies time shifts to align the synthetics to the observed data (Zhao and Helmberger, 1994). The relative time shift length is established by cross-correlating the synthetics with the data. The final step is the actual inversion, using least squares regression and singular value decomposition.

Figures DR-5 – DR-11 summarize the moment-tensor inversion results for the two mainshock events and five aftershocks that were used in the stress inversion. Table DR-2 contains the full set of moment tensor parameters for these events. For purposes of interpretation, the hypocenter depths obtained from traveltimes inversion are taken as more definitive than the centroid depths estimated from moment-tensor inversion.

Figure DR-5. Moment-tensor inversion results for the first mainshock (2017/05/01 12:31:54). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is ± 7.5 km.

Figure DR-6. Moment-tensor inversion results for the second mainshock (2017/05/01 14:18:15). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is ± 3 km.

Figure DR-7. Moment-tensor inversion results for the aftershock afs30 (2017/05/05 05:21:30). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is +/- 8 km.

Figure DR-8. Moment-tensor inversion results for the aftershock afs32 (2017/06/22 01:35:08). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is +/- 2 km.

Figure DR-9. Moment-tensor inversion results for the aftershock afs33 (2017/09/16 23:38:02). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is +/- 3 km.

Figure DR-10. Moment-tensor inversion results for the aftershock afs34 (2017/12/19 00:10:16). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is +/- 2 km.

Figure DR-11. Moment-tensor inversion results for the aftershock afs35 (2018/01/18 02:47:47). Focal mechanism parameters including double-couple (DC), isotropic (ISO) and compensated linear vector dipole (CLVD) components are indicated in the top left. Goodness of fit is plotted for a range of focal depths, top right. Waveform fitting results are also plotted for selected stations (bottom). Observed waveforms are plotted in red, and best-fitting synthetic waveforms are plotted in blue. Each station is annotated with the time shift of the observed waveforms by CAP and the squared correlation coefficient as a percentage. Depth uncertainty at 95% fit is +/- 2 km.

Text DR4. Stress inversion

Linear stress inversion was proposed by Michael (1984) and involves the Bott hypothesis (Wallace, 1951; Bott, 1959), which postulates that fault slip is in the direction of the maximum resolved shear stress on a fault. In this study, we applied the stress inversion algorithm of Vavryčuk (2014), which automatically distinguishes the fault plane by iteratively evaluating the Mohr-Coulomb failure criterion. Here, the fault plane choice algorithm analyzes orientations of the two nodal planes in the stress field and finds which of the nodal planes are more unstable and thus more susceptible to shear faulting. The stress inversion result obtained using this method for 7 earthquakes from the St. Elias earthquake sequence is summarized in Figure DR-12.

Figure DR-12. Stress inversion result for 7 earthquakes from the St. Elias earthquake sequence (Figures DR-4 to DR-10). The moment tensor parameters for these events are given in table DR-2. The determined principal stress directions are indicated by the green symbols, while the P/T axes of the earthquakes are marked by red circles and blue plus signs, respectively.

Text DR5. Coulomb Stress Calculation

Coulomb failure criterion is one of the most widely used criteria for rock-failure stress condition. The application of it has been appeared in many field observations as wells as laboratory experiments (e.g. Jaeger et al. 2009). The program Coulomb 3 is published by USGS to calculate static displacements, strains and stresses at any depth caused by fault slip, magmatic intrusion or dike expansion/contraction (Toda et al. 2011). It implements calculations assuming an elastic half-space with uniform isotropic elastic properties, and the effective coefficient of friction, given by a user, is applied to all of the calculating space. The program is based on a grid-search algorithm, which lets the user specify the scale of the area to calculate. Every grid point used in the calculation is assigned the fault orientation based on an assigned receiver fault, and a user can assign multiple source faults. Source fault positions, strike, dip and rake are required.

A receiver fault can be one of three types: a specified fault with provided orientation information, a fault optimally oriented for failure in the calculation area and a nodal plane from a provided focal mechanism file. Specifying a receiver fault is widely used by researchers and provides the simplest calculation among the three methods. Using this method, the fault strike, dip and rake of the receiver plane are specified. Based on specified regional stress parameters, the stress imparted by the source fault is calculated based on the given friction coefficient on the receiver fault. In this study, the receiver fault orientation was specified based on nodal plane information from a focal mechanism. Additional details about the Coulomb 3 algorithm can be found in Toda et al. (2011).

Table DR-1. Catalog of seismicity with re-located hypocenters based on double-difference relocation method.

Table DR-2. Moment tensor parameters for 7 events considered in this study.

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