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# Locating nonvolcanic tremors beneath the San Andreas Fault using a station-pair double-difference location method

Haijiang Zhang,[1] Robert M. Nadeau,[2] and M. Nafi Toksoz[1]

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[1] It has been a challenging task to locate nonvolcanic tremors because of the lack of impulsive wave arrivals. A station-pair double-difference (DD) location method is developed to determine absolute tremor locations by directly using the station-pair travel time differences measured from cross-correlating tremor waveform envelopes. Multiple tremors are located together for inverting for station corrections to take into account velocity model inaccuracy. The new method is applied to tremors in the Parkfield region of central California between 27 July 2001 and 21 February 2009. Compared to the tremor catalog locations determined from a grid search location method, most of the newly located tremors are located at depths between 20 and 35 km, well below the seismogenic zone in the area. The tremors beneath Cholame, CA are more clearly separated into two zones laterally distributed across the San Andreas Fault, with most tremors occurring to the southwest and exhibiting a periodic pattern of occurrence. The new tremor locations help better delineate the spatial and temporal distribution of tremor activity and therefore are helpful for better understanding tremor origin and process. **Citation:** Zhang, H., R. M. Nadeau, and M. N. Toksoz (2010), Locating nonvolcanic tremors beneath the San Andreas Fault using a station-pair double-difference location method, *Geophys. Res. Lett.*, *37*, L13304, doi:10.1029/2010GL043577.

## 1. Introduction

[2] Nonvolcanic tremor (NVT) activity has recently been observed deep (~20 to 40 km) in subduction zones off Japan [*Obara*, 2002], beneath the megathrust in Cascadia [e.g., *Rogers and Dragert*, 2003], and along the SAF in the Parkfield-Cholame region of California [*Nadeau and Dolenc*, 2005]. This NVT activity is characterized by low-amplitude seismic signals lasting continuously for a few minutes to several days with predominant frequency content generally between 1 and 10 Hz. Tremors are also emergent in character and generally do not contain any clear P- or S- phase arrivals. Tremor signals generally contain at least some short pulsating bursts of larger amplitude energy embedded in lower amplitude activity (present to varying degrees), considered to be low frequency earthquakes (LFEs) [*Katsumata and Kamaya*, 2003; *Shelly et al.*, 2007], and at any given seismic station the tremor signals are often similar in character to local cultural noise signals. Unlike cultural noise, however, tremors are observable on multiple stations, even those separated by many 10s of km, and their pulsating bursts are, to first order, coherently timed among the different stations. Seismic energy from tremors also appears to propagate at S-wave velocities.

[3] Because of lack of impulsive wave arrivals, it has been a challenge to accurately locate tremors. *Rubinstein et al.* [2010] give a review of various tremor location methods. Cross-correlation time alignments of the similarly shaped energy envelopes of the tremors, generally over a longer time window (e.g., 6 minutes), have been used to locate these events by converting station-pair differential arrival times into individual arrival times at different stations [*Obara*, 2002; *Nadeau and Dolenc*, 2005; *Nadeau and Guilhem*, 2009]. Station-pair differential arrival times are also directly used in a grid search location method to find the best tremor locations to minimize the residuals of the observed and theoretic differential times [*Suda et al.*, 2009]. Instead of minimizing station-pair differential time residuals, the tremors can also be located by searching for locations to maximize tremor signal coherency among seismic stations [*Wech and Creager*, 2008]. In nature, the above mentioned tremor location methods can be categorized as a single-event location method and the location accuracy relies heavily on an *a priori* velocity model. As a result, the tremor locations are generally scattered and absolute locations have strong dependence on velocity model.

[4] A different tremor location approach is to locate LFEs that are found within and may comprise much of the tremor signal [*Shelly et al.*, 2007, 2009; *Brown et al.*, 2009]. Compared to waveform envelope methods, the LFE location method only locates relatively impulsive events within tremor [*Rubinstein et al.*, 2010]. In comparison, LFEs have discernible S, and sometimes P arrivals when cross-correlation techniques are used. Thus they may be located using more conventional location methods. Double-difference (DD) location method [*Waldhauser and Ellsworth*, 2000] has been used to locate LFEs having similar waveforms with their templates. With this approach, previously scattered tremor locations resolve themselves into clustered and concentrated structures, for example, at the plate boundaries in multiple subduction zones [e.g., *Brown et al.*, 2009]. However, the absolute location accuracy of matched LFEs is determined by the location accuracy of the LFE templates.

[5] In this paper, we present a station-pair DD location method that directly uses the station-pair differential travel times to locate seismic events. This is a direct extension of the original DD location method of *Waldhauser and Ellsworth* [2000], which uses the event-pair differential travel times. The station corrections are also included in our inversion to compensate for the velocity model inaccuracy. Therefore,

[1]Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts, USA.
[2]Berkeley Seismological Laboratory, University of California, Berkeley, California, USA.


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the proposed approach is a 'multiple-event' location method. The new station-pair DD location method is specifically designed to locate tremors by directly using station-pair cross-correlation delay times measured from tremor waveform envelops at different stations. It can also be applied to locate LFEs if such travel time difference information is available. The new location method is applied here to locate nonvolcanic tremors beneath the San Andreas Fault around the Parkfield and Cholame area detected and located by *Nadeau and Guilhem* [2009].

## 2. Station-Pair Double-Difference Location Method

[6] The DD location method, developed by *Waldhauser and Ellsworth* [2000], has been widely used to locate earthquakes using differential arrival times at common stations from pairs of events. The same concept can be applied to the case where differential times on pairs of stations from common events can be accurately calculated.

[7] Assume a single event $i$ is recorded by two stations $j$ and $k$. For each event and station pair, a linearized location equation is used to relate the travel time residual to location and origin time perturbations. To compensate for the velocity structure uncertainty, a site correction term is also included, as follows:

$$\mathrm{r}_j^i = \sum_{\mathrm{m}=1}^{3} \frac{\partial \mathrm{T}_j^i}{\partial \mathrm{x}_\mathrm{m}^i} \Delta \mathrm{x}_\mathrm{m}^i + \Delta\tau^i + s_j, \tag{1}$$

$$\mathrm{r}_k^i = \sum_{\mathrm{m}=1}^{3} \frac{\partial \mathrm{T}_k^i}{\partial \mathrm{x}_\mathrm{m}^i} \Delta \mathrm{x}_\mathrm{m}^i + \Delta\tau^i + s_k, \tag{2}$$

where $\mathrm{r}_j^i$ and $\mathrm{r}_k^i$ are residuals from events $i$ at stations $j$ and $k$, T's are travel times, x and $\Delta$x are hypocenter coordinates and their perturbations, $\tau$'s are origin times, and $\mathrm{s}_j$ and $\mathrm{s}_k$ are station corrections for stations $j$ and $k$.

[8] By subtracting equation (2) from equation (1), we obtain

$$\mathrm{r}_j^i - \mathrm{r}_k^i = \sum_{\mathrm{m}=1}^{3} \left( \frac{\partial \mathrm{T}_j^i}{\partial \mathrm{x}_\mathrm{m}^i} - \frac{\partial \mathrm{T}_k^i}{\partial \mathrm{x}_\mathrm{m}^i} \right) \Delta \mathrm{x}_\mathrm{m}^i + s_j - s_k, \tag{3}$$

where $\mathrm{r}_j^i - \mathrm{r}_k^i = (\mathrm{T}_j^i - \mathrm{T}_k^i)^{\mathrm{obs}} - (\mathrm{T}_j^i - \mathrm{T}_k^i)^{\mathrm{cal}}$ is called station-pair double difference. Note that the origin time terms drop out since there is only one event involved. Based on equation (3), the observed station-pair differential times $(\mathrm{T}_j^i - \mathrm{T}_k^i)^{\mathrm{obs}}$, which can be calculated from waveform cross-correlation, are directly related to location perturbations.

[9] Following the analysis of *Wolfe* [2002], if the differences of location partial derivatives in equation (3) are not too small, or close to zero, the station-pair differential times should be able to solve for absolute event locations. The two location partial derivatives are approximately the same only if the two stations are very close. However, it is rarely the case that all pairs of stations are close together. Compared to the event epicenter, event depth may be determined less accurately by using station-pair differential times when the network of stations is located directly above the seismic event. In this case, the partial derivatives of the arrival times with respect to event depth for two stations are close and therefore are insensitive to depth variations. For the station correction terms, similar to origin time terms in the event-pair case, only relative values are determined by station-pair differential times because of the same partial derivatives of 1 [*Wolfe*, 2002].

## 3. Application to Nonvolcanic Tremors Beneath San Andreas Fault

[10] Between July 27, 2001 and February 21, 2009, 2198 tremors were detected using borehole seismometer data from the High Resolution Seismic Network (HRSN) at Parkfield, California [*Nadeau and Guilhem*, 2009]. The migrating grid search code BW_RELP of *Uhrhammer et al.* [2001] was used to locate the tremors from the best fitted relative arrival times from station-pair differential times measured from waveform envelop cross correlation. For this process tremor energy was assumed to propagate at S-velocities. The velocity model used was a gradient layer over a half space with velocities and gradients based on the regional velocity model used by the USGS for their NCSN earthquake catalog. Specifically, they used a surface S velocity of 2.644 km/s with a gradient down to 40 km depth of 0.05968 (km/s)/km and an S-velocity of 5.0316 km/s below 40 km. No station corrections were applied in the location process. The catalog locations of the "well located" 1246 tremors are shown in Figure 1a. There are 441,595 station-pair differential times for 64 stations, with an average of 305 pairs for each event. For the station-pair DD location method, we started from catalog tremor locations determined by *Nadeau and Guilhem* [2009] and used the same S velocity model. The travel times are calculated using the finite-difference travel time calculation method of *Podvin and Lecomte* [1991]. The inversion is performed in the same Cartesian coordinate system as *Thurber et al.* [2006], with the origin at (35° 57.60′N, 120° 30.28′W), the *Y*-axis almost aligning with the SAF surface trace (139.2° rotation from the North), and the X-axis pointing positive northeast (Figure 1a).

[11] We performed 6 iterations of simultaneous determination of tremor locations and station corrections. At each iteration, data residual weighting based on a bi-weight function was applied to down-weight large residuals [*Waldhauser and Ellsworth*, 2000]. For example, at the start of the inversion, station-pair differentials with residuals greater than 7.9 s are removed from the inversion. The LSQR method was used to solve the weighted linear systems with the damping value selected via a trade-off analysis. The unweighted and weighted root-mean-square (RMS) residuals for initial catalog locations were 2.0 s and 1.7 s and they decreased to 0.96 s and 0.65 s for the final locations, respectively. Figure 2a shows the comparison of histograms of station-pair differential time residuals for catalog tremor locations and final tremor locations. It can be seen that the station-pair differential travel time residuals both show a Gaussian-like distribution but the station-pair DD location method shows a narrower distribution.

[12] We evaluated tremor location uncertainties using two methods. The first method is known as the "restoration method" used in evaluating velocity model resolution in seismic tomography. We followed a similar strategy here by calculating synthetic travel times based on the final tremor locations and then constructed synthetic station-pair differ-

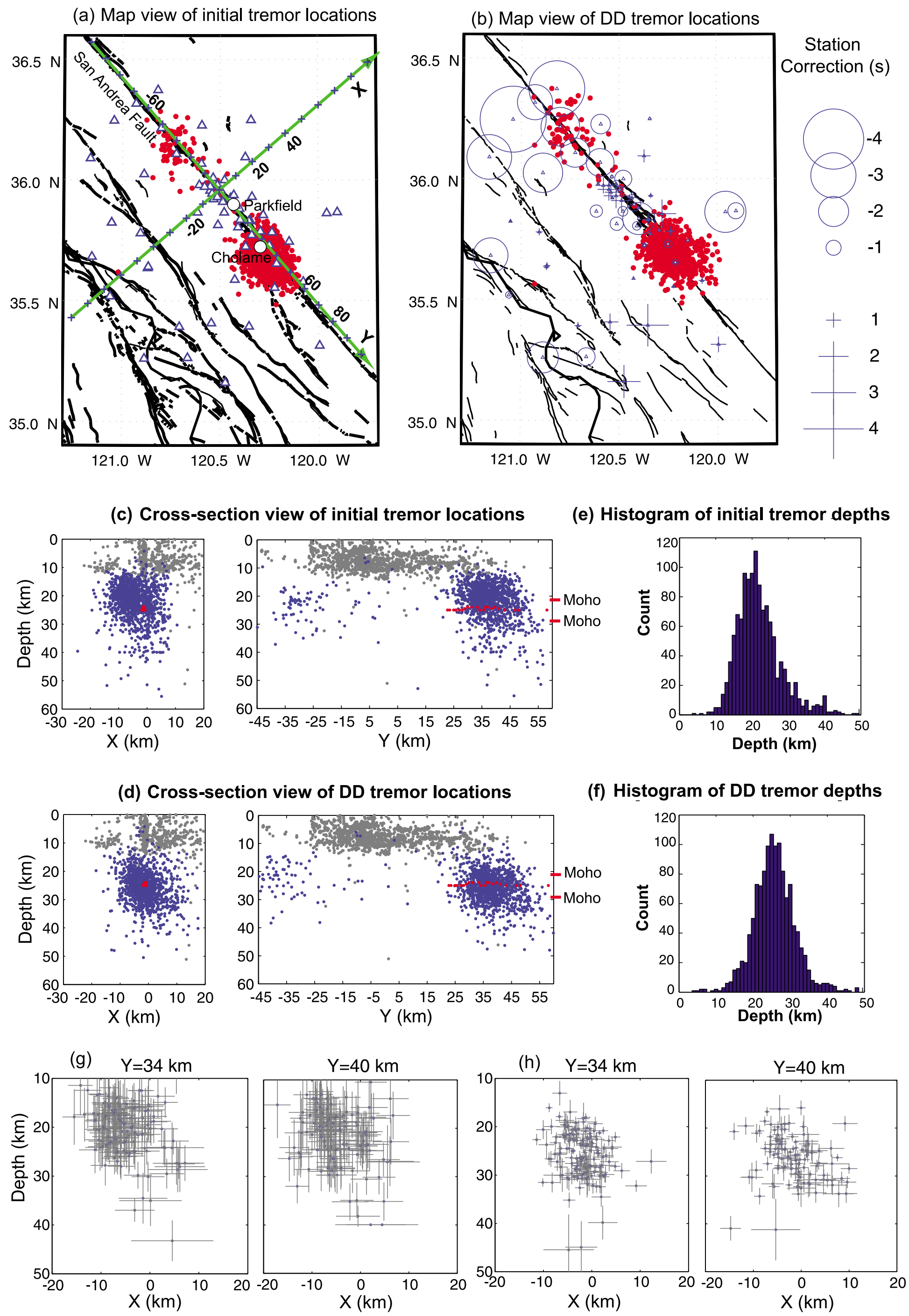


**Figure 1.** Comparison of catalog and DD tremor locations. (a) Map view of 1246 tremors (red dots) and 64 stations (blue triangles) used in this study. Tremor catalog locations are from *Nadeau and Guilhem* [2009]. The Cartesian coordinate system used in location is shown with a tick (marked as "+") interval of 10 km on X and Y axes (green). Faults are shown as black lines. Parkfield and Cholame are marked as white dots. (b) Map view of new tremor locations determined by station-pair DD location method. Station corrections in seconds are also shown. (c) Catalog tremor locations are shown in X-depth and Y-depth sections. (d) DD tremor locations are shown in X-depth and Y-depth sections. (e) Depth distribution of catalog tremor locations. (f) Depth distribution of DD tremor locations. (g) Across-fault cross sections of catalog tremor locations and the associated 95% uncertainty bounds at Y = 34 and 40 km (within 1 km on both sides). (h) The same as Figure 1g but for DD tremor locations. In both Figures 1c and 1d, background seismicity from *Thurber et al.* [2006] is shown as grey dots, the inferred LFE locations of *Shelly* [2009] are shown as red dots, and the Moho depths of 22 and 29 km from *McBride and Brown* [1986] are marked as red lines.

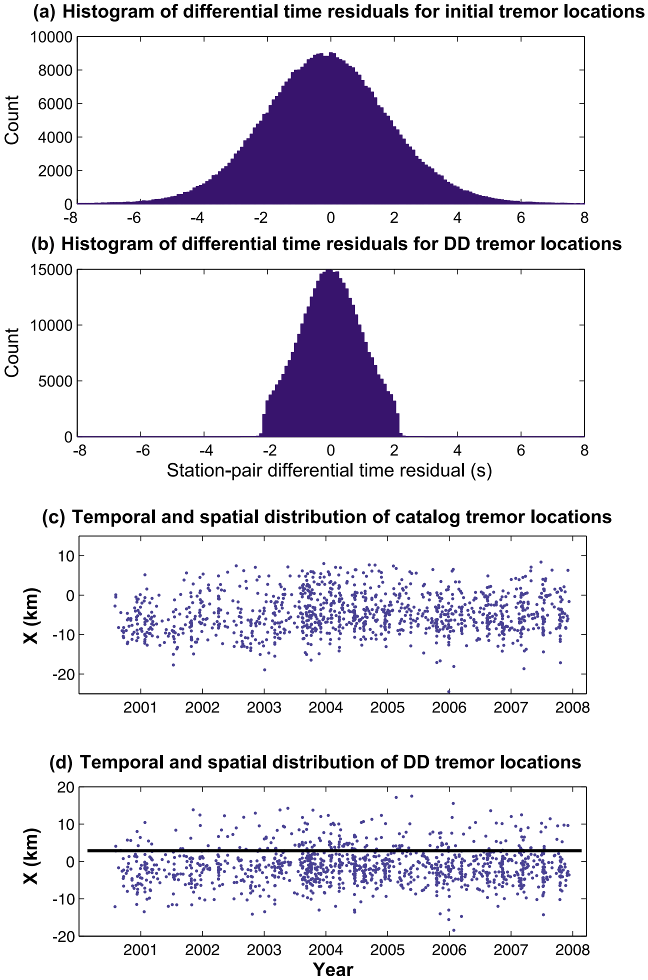


**Figure 2.** Histogram of station-pair differential time residuals and spatial-temporal distribution for (a and c) catalog tremor locations and (b and d) DD tremor locations. The black line located around X = 3 km in panel d separates the tremors into southwestern (with periodic episodes) and northeastern (aperiodic) zones.

ential times that have the same distribution as the real data. To assess if the tremor locations can be determined under the current observation system, no noise is added. We then relocated tremors using the synthetic station-pair differential times following the same inversion strategy as the real data inversion. The RMS differences in X, Y and depth are 1.58, 0.75 and 1.48 km with standard deviations of 0.47, 0.55 and 0.48 km between relocated and "true" tremors, respectively. In this synthetic test, the S velocity models are the same for calculating synthetic times and inversion. To simulate the case where the velocity model used for inversion is different from the real velocity structure, we calculated the synthetic travel times based on the S velocity model converted from the P velocity model of *Thurber et al.* [2006] using the empirical Vp-Vs relationship of *Brocher* [2005]. The converted Vs model is expected to have strong velocity variations across the San Andreas Fault, as seen in the Vp model [*Thurber et al.*, 2006]. In comparison, the RMS location differences are 1.71 km in X, 1.18 km in Y, and 2.45 km in depth between relocated and "true" tremor locations, with standard deviations of 1.06, 0.99, and 1.65 km, respectively. As expected, the tremors are located less accurately when the velocity model used for location is (greatly) different from the true model. However, because of the incorporation of station corrections in the inversion, part of the velocity model uncertainty is compensated for and therefore the tremors are still located reasonably well. We also added Gaussian noise with a zero mean and a standard deviation of 2 s to the station-pair differential times based on differential time residual distribution for the real data (Figure 2a). In this case, the RMS location differences are 1.54 km in X, 0.99 km in Y and 2.30 km in depth, with standard deviations of 1.19, 0.99, and 2.04 km, respectively.

[13] Another method to estimate the location uncertainty is to use the covariance matrix constructed from the linearized inverse problem [*Aster et al.*, 2005]. At the final iteration, after event locations and station corrections are determined, it is assumed that the data residuals are errors caused by data noise and velocity model uncertainty. For each event, we constructed the Fréchet derivative matrix using the associated station-pair differential times by excluding station correction terms. The singular value decomposition (SVD) is used to construct the model covariance matrix. Assuming data residuals are independent, we can estimate the location uncertainty using the covariance matrix [*Aster et al.*, 2005]. The RMS tremor location uncertainties are 1.21 km, 1.01 km, and 1.27 km, respectively, in an order similar to what estimate using the "restoration method."

## 4. Results and Discussion

[14] Figure 1 shows the comparison of tremor locations from the station-pair DD location method and the catalog. The notable differences between two sets of tremor locations are that the newly located tremors are shifted to northeast and are deeper (Figures 1c and 1d). On average, the shift is 3.4 km in X and 3.7 km in depth. The location shift to northeast could be caused by biased catalog tremor locations due to using a simplified velocity model in a region of strong lateral heterogeneity [*Nadeau and Guilhem*, 2009]. The shift in depth is most likely due to the fact that the new station-pair DD location method better determines the tremor depths by avoiding the coupling effect of depth and origin time. Overall, the location uncertainties of DD tremor locations are about half those of the catalog tremors (Figures 1g and 1h; also see auxiliary material for other sections).[1] Furthermore, the new tremor locations also appear to be more clustered and some substructure can be seen (Figure 1h).

[15] Figures 1e and 1f show histograms of tremor depths from the catalog and station-pair DD locations. It can be seen that most of tremors from the catalog are located at depths between ~15 and 30 km. However, the tremors from the station-pair DD location method mostly occurred at depths between ~20 and 35 km. For the tremors in the Cholame area, 84% of newly relocated tremors are located at depths between 20 and 35 km, where the location uncertainties are much smaller than for those located deeper (see auxiliary material). In comparison, only 54% of selected catalog tremors are located at depths between 20 and 35 km. In the selected tremor catalog, 9% of tremors are located above 15 km,

[1]Auxiliary materials are available in the HTML. doi:10.1029/2010GL043577.

within the seismogenic zone. In comparison, only 2% of the newly relocated tremors are located above 15 km. In this area, the earthquakes are located in the upper ~15 km of the Earth's crust (Figures 1c and 1d) [*Thurber et al.*, 2006] and the Moho depth is 22–29 km [*McBride and Brown*, 1986]. Therefore, our results indicate that the tremors predominate below the seismogenic zone, in the ductile lower crust. Statistically speaking, the new tremor locations are strongly suggestive of a distinct gap in depth between the seismogenic and tremor zones.

[16] *Nadeau and Guilhem* [2009] noticed that periodic episodes of tremors at Cholame after the 2004 Parkfield M6 earthquake did not occur ubiquitously throughout the central part of the SAF. Instead they are more concentrated and periodic in the southwestern portion of the tremor zone (Figure 2c). The new tremor locations determined by station-pair DD location method more clearly show this feature compared to the tremor catalog of *Nadeau and Guilhem* [2009] (Figure 2d). The boundary separating the tremors into southwestern and northeastern tremor zones at Cholame is at around X = 3 km, located to the northeast of the SAF trace at surface. In comparison, the boundary is located slightly to the southwest of the SAF surface trace in the *Nadeau and Guilhem* [2009] catalog. The concentration and periodic behavior of the tremors to the southwestern side of the SAF in the Cholame tremor zone indicates that the process generating tremors across the fault zone may be different and structurally controlled.

[17] Station corrections are generally greater at stations farther away from the Cholame tremor zone, indicating the accumulation of velocity model uncertainties along longer ray paths (Figure 1b). The station corrections are mostly negative, indicating that the used 1D S velocity model is on average faster than the real velocity structure. The absolute value of station corrections can be as great as 4 s, however, only station correction differences between stations are resolvable by the station-pair DD location method.

[18] Compared to the catalog tremor locations of *Nadeau and Guilhem* [2009], the absolute locations are better determined for individual tremors using the station-pair DD location method. However, they still look scattered and "cloudy" in both horizontal and vertical sections (Figure 1). This could be partly due to the fact that waveform envelope location methods result in some average location over a longer time period [*Rubinstein et al.*, 2010]. It is shown that in some cases LFE activity might migrate considerable distances over a period of several minutes [*Shelly*, 2009]. The primary effect of the migration of the source of radiated energy within the measured time window is a temporal expansion or contraction of the tremor envelope shape arriving at the different stations. In the more extreme cases and assuming a wave propagation speed of 4 km/sec and migration of 10 km in 6 minutes (used in this study), this would result in relative stretching or contraction of the shapes on the order of ~1% (a few seconds over the entire 6-minute window) between station pairs. The net effect on the station-pair alignments of the envelopes, therefore, is expected to be relatively small.

[19] To better understand the internal structure of tremor locations, the relative event location method using differential times between pairs of LFEs could be used [*Shelly et al.*, 2007, 2009; *Brown et al.*, 2009]. *Shelly et al.* [2009] showed that the locations of LFEs in a 24-hour period of tremor form a near-linear structure striking parallel to the SAF, indicating that at least some of the tremors in the Parkfield region occur on such a structure (Figures 1c and 1d). However, it is noted that only a small fraction of the most energetic tremor activity (~15%) in the Parkfield region appears to be composed of LFEs (see auxiliary material). Accurately locating the remaining 85% of the activity is important for determining the structural conditions responsible for generating NVTs and for monitoring the evolution of the NVT process in this non-subducting tectonic environment. If station-pair differential times can be obtained for LFEs, the station-pair DD location method can be helpful for determining the absolute locations of additional LFE templates in this area, which to date have been largely inferred from travel time moveout information and a priori assumptions of SAF parallel alignment.

## 5. Concluding Remarks

[20] A station-pair DD location method has been developed and applied to locate nonvolcanic tremors beneath the SAF around the Parkfield and Cholame area of central California by directly using station-pair differential times measured from cross-correlation of tremor signal envelops. Station corrections are also incorporated in the inversion to compensate for velocity model uncertainty used for location. Compared to tremor catalog locations determined from the grid search location method used by *Nadeau and Guilhem* [2009], the new tremor locations are more concentrated deeper, at depths between 20 and 35 km and the Cholame tremor zone is more clearly divided into southwestern and northeastern zones of activity, indicating a broad across fault distribution the tremor generation zone in the region.

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R. M. Nadeau, Berkeley Seismological Laboratory, University of California, 215 McCone Hall, Berkeley, CA 94720, USA.

M. N. Toksoz and H. Zhang, Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, 77 Massachusetts Ave., Cambridge, MA 02139, USA. (hjzhang@mit.edu)