Imaging the Nonvolcanic Tremor Zone Beneath the San Andreas Fault at Cholame, California using Station-Pair Double-Difference Tomography

^{1*}Haijiang Zhang, ²Robert M. Nadeau, ¹Hao Guo

²Berkeley Seismological Laboratory and

Abstract

Nonvolcanic tremors have been detected deep beneath the San Andreas fault (SAF) near Cholame, California. The tremors are modulated by small stress changes induced by Earth tides and regional and teleseismic earthquakes, implying elevated pore fluid pressure in the tremor zone. In this area, tremors lie beneath the seismogenic zone, limiting the usefulness of local earthquakes for resolving seismic velocity properties in the tremor zone. Station-pair double-difference tomography that uses differential tremor arrival times is used, for the first time, to provide new detailed information about the complexity of the lower-crust and upper-mantle transition and the occurrence of deep tremors in zones of localized low shear wave velocity anomalies. The anomalies may represent zones containing high-pore pressure fluids that migrate into the tremor zone episodically from the southwest and that are ultimately derived from dehydration of deep serpentized mantle wedge material. These results improve our understanding of the tremor process, why tremors are concentrated in specific zones along the SAF, and possible relationships between tremors and large historic earthquakes.

1. Introduction

Nonvolcanic tremor (NVT) activity (long-duration seismic signals with no clear P or S waves) occurs primarily along subduction zones (Obara, 2002; Schwartz and Rokosky, 2007; Rubinstein et al., 2010) but has also been found along the transform San Andreas Fault (SAF) plate boundary (Nadeau and Dolenc, 2005; Gomberg et al., 2008). In most of these locations, NVT activity appears to be concentrated in locations adjacent to locked zones, to increase in activity and/or amplitude with detectable transient zone deformation (Rogers and Dragert, 2003), and to be stimulated by very small stress changes (a few kilopascals or less) from earth-tides (e.g., Rubinstein et al., 2008; Thomas et al., 2009), nearby large events (Nadeau and Guilhem, 2009) and even the passage of surface waves from distant regional or teleseismic earthquakes (e.g., Rubinstein et al., 2007; Peng et al., 2009; Guilhem et al., 2010). Embedded within the tremor signals are, in a number of cases, detectable larger amplitude, low frequency S-wave arrivals generated by low frequency earthquakes (LFEs) (Katsumata and Kamaya, 2003; Shelly et al., 2006). In subduction zones, these LFEs are located along the plate interface and above intraplate earthquakes (Shelly et al., 2006; Brown et al., 2009). In Japan as well as other subduction zones the LFEs also have focal mechanisms that are consistent with shear slip on the plate interface (Ide et al., 2007; Brown et al., 2009). The existence of over-pressured pore fluids in the tremor region in subduction zones is supported by the findings that tremors occur preferentially in regions having high compressional- to shear-wave seismic velocity ratios (Audet et al., 2009; Shelly et al., 2006) or ultralow shear wave velocity (Song et al., 2009). The fluids are presumably derived from dehydration of subducting oceanic crust, and appear to play an important role in generating NVT in subduction zones, by decreasing the effective normal stress (e.g. Obara, 2002; Shelly et al., 2006; Rubinstein et al., 2010). Therefore, NVT activity in

subduction zones may be manifestations of long-duration shear slip episodes occurring within subducting plate boundary zones in high pore-fluid pressure environments.

Along the transform SAF plate boundary, NVTs occur deep below the seismogenic zone (the upper ~15 km of Earth's crust where most SAF earthquakes occur) in the ductile lower crust or upper mantle (Nadeau and Guilhem, 2009; Shelly and Hardebeck, 2010; Zhang et al., 2010; Plourde et al., 2015). NVT activity is not, however, ubiquitous along the deep SAF, occurring instead in concentrated pockets often associated with known shallower geologic and geophysical complications along the plate boundary, e.g. at transition zones between freely slipping and locked fault or foreshock regions of past large earthquakes (Nadeau and Guilhem, 2009). Most of SAF tremors occur near Cholame in central California (Guilhem and Nadeau, 2012), but there are also a few tremors distributed in other places in central California (Peng et al., 2009) and northern California (Plourde et al., 2015). Like subduction zone tremors, SAF tremors often occur episodically (Guilhem and Nadeau, 2012) and contain LFEs (Shelly and Hardebeck, 2010). Inferential evidence suggests that the SAF tremors also occur under high (near lithostatic) pore-fluid pressure conditions (Nadeau and Guilhem, 2009; Thomas et al., 2009). Unlike subduction zones, however, a source of dehydration fluids from actively subducting oceanic plate material is not available in the transform SAF environment. However fossil slab material containing fluids may exist in distributed pockets beneath the SAF system (Fulton et al., 2009; Kirby et al., 2014). While the mere presence of fluids in the lower crust can only slightly decrease S-wave velocity (V_s) at NVT depths (Bourbie et al., 1987), extremely high pore-fluid pressures (near lithostatic) reduce V_s significantly (Nur and Simmons, 1969) to levels resolvable by seismic tomography.

To better understand the process of NVT generation, why NVTs are concentrated in specific zones along the SAF, and NVT's potential implications for seismic hazard along the

SAF (WGCEP, 2008; Akciz et al., 2010), a more detailed picture of deep SAF seismic velocity structure is needed. To date, attempts at imaging the seismic velocity structure containing the deep NVT in California have been limited, due largely to the lack of the deep local earthquake sources needed to resolve velocities at the sub-seismogenic depths of the NVTs (Nadeau and Guilhem, 2009; Zhang et al., 2010; Zeng et al., 2016).

An alternative to using earthquake sources to image velocity structure in NVT zones is to use the NVT signal itself to determine the velocity structure. Due to the lack of impulsive phase arrivals within the NVTs, however, it has been a challenge to accurately locate NVTs, let alone image the NVT zone. Cross-correlation time alignments of the similarly shaped energy envelopes of NVTs 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 have also been directly used in a grid search location method to find the tremor locations that minimize the residuals, i.e., the observed versus theoretical differential times (Suda et al., 2009). More recently a station-pair double-difference (DD) location method has been developed, a direct extension of standard (event-pair) DD location, and used for locating tremors beneath the SAF (Zhang et al., 2010). We present here a station-pair DD tomography method that uses the station-pair differential times directly to simultaneously image the NVT zone and locate the NVTs. We apply this method to NVTs detected and located by Nadeau and Guilhem (2009) beneath the SAF around the Parkfield and Cholame areas.

2. Station-pair DD tomography method

The DD tomography algorithm of Zhang and Thurber (2003, 2006), known as tomoDD, uses the differential arrival times from event pairs observed on common stations to simultaneously determine seismic event locations and velocity structure. It has the advantage of directly utilizing the accurate differential times derived from waveform cross-correlation and imaging the source region structure by cancelling some common model errors away from the source region. The same concept can be applied to the case where differential times on pairs of stations from common events can be accurately determined (Zhang et al., 2008; Eberhart-Phillips and Reyners, 2012).

For each event i recorded by a pair of stations j and k , the misfit between the observed and predicted arrival times is linearly related to the desired perturbations to the hypocenter location, origin time and velocity structure parameters, along with a station correction term, as follows:

$$r_j^i = \sum_{m=1}^3 \frac{\partial T_j^i}{\partial x_m^i} \Delta x_m^i + \int_i^j \delta u dl + \Delta \tau^i + s_j, \quad (1)$$

$$r_k^i = \sum_{m=1}^3 \frac{\partial T_k^i}{\partial x_m^i} \Delta x_m^i + \int_i^k \delta u dl + \Delta \tau^i + s_k, \quad (2)$$

where r_j^i and r_k^i are residuals from event i at stations j and k , T 's are travel times, $\mathbf{x}$ and $\Delta \mathbf{x}$ are hypocenter coordinates and their perturbations, τ 's are origin times, δu is perturbation to slowness, dl is an element of path length, and s_j and s_k are station corrections for stations j and k , respectively.

By subtracting Equation (2) from Equation (1), we obtain

$$r_j^i - r_k^i = \sum_{m=1}^3 \left(\frac{\partial T_j^i}{\partial x_m^i} - \frac{\partial T_k^i}{\partial x_m^i} \right) \Delta x_m^i + \int_i^j \delta u dl - \int_i^k \delta u dl + s_j - s_k, \quad (3)$$

where $r_j^i - r_k^i = (T_j^i - T_k^i)^{\text{obs}} - (T_j^i - T_k^i)^{\text{cal}}$ is called the 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 $(T_j^i - T_k^i)^{\text{obs}}$, which can be calculated from waveform cross-correlation, are directly related to location and slowness perturbations. Different from the DD tomography algorithm of Zhang and Thurber (2003) and the tomography method of Eberhart-Phillips and Reyners (2012), the novelty of our proposed station-pair DD tomography method lies in directly using station-pair differential travel times to simultaneously solve for the slowness model and event locations. In addition, station corrections are also included to compensate for unmodeled velocity structure near stations.

3. Cholame tremor dataset and tomographic details

In this study, we apply the station-pair DD tomography method to 1243 relatively well-located NVTs beneath SAF that occurred between July 27, 2001 and February 21, 2009 (Nadeau and Guilhem, 2009; Zhang et al., 2010). Among them, more than 95% of the NVTs occurred near the Cholame tremor zone and to the north only few NVTs are distributed along SAF (Figure 1). Station-pair differential times were obtained by cross-correlating 6-minute RMS envelopes for all possible station-pair combinations having station separation distances of 100 km or less. Only differential times with maximum cross-correlation coefficients of 0.70 or greater were selected. Theoretically, differential times between a pair of stations will equal the sum of the differential times between each of the stations and a common third station (i.e., 2-leg estimates). Only differential times that differed by less than 0.5 seconds from the median average of their

corresponding 2-leg estimates were used in our analysis. There were 441,595 station-pair differential time measurements involving 64 stations used in the analysis, with an average of 305 measurements for each event. The distribution of differences between the differential times and the median of their corresponding 2-leg estimates were also used to estimate uncertainties for the differential times of $\sim \pm 0.30$ s at the 2-sigma level.

The 6-minute NVT signal used in the envelope cross-correlation most likely represents radiation from a source of extended dimension or from multiple distributed sources. The differential times, therefore, represent differences in the arrival times of the centroid of energy release (Rubinstein et al., 2010). Locations derived from the differential times, therefore, represent an average location of the NVT source. The LFE component of NVT sources sometimes migrates at speeds approaching 30 m/s (Shelly, 2015). However, this is small compared to the velocity of wave propagation (~ 4000 m/s) and, therefore, has little impact on RMS envelope shapes and the differential times derived from their cross-correlation alignments

For the tomographic inversion, we chose the same Cartesian coordinate system as Thurber et al. (2006), with the origin at ($35^{\circ} 57.60'$ N, $120^{\circ} 30.28'$ W), the Y-axis closely aligned with the SAF surface trace (139.2° rotation from the North), and the X-axis pointing positive northeast (Figure 1). Normal to the SAF, the inversion grid intervals range from 2 to 5 km near the fault and 7 to 25 km away from the fault. The grid nodes along the fault strike are denser (5 km) near the Cholame tremor zone and coarser in other regions (10-15 km). In depth, they are evenly distributed with an interval of 5 km.

As in the tomoDD algorithm, the residual weighting based on a bi-weight function is applied to down-weight large residuals (Zhang and Thurber, 2003, 2006). A first-order smoothing constraint is also applied to slowness perturbations to stabilize the tomographic

system. The damped least squares system is solved using the LSQR algorithm of Paige and Saunders (1982). The inversion starts from the NVT locations determined by the station-pair DD location method (Zhang et al., 2010) and a 3D Vs model converted from the Vp model of Thurber et al. (2006) using a constant Vp/Vs ratio of 1.75 for depths shallower than 25 km and the 1D S-wave velocity model of Nadeau and Guilhem (2009) for greater depths. The finite-difference travel time calculation method of Podvin and Lecomte (1991) is used to calculate travel times. The damping and smoothing parameters are chosen via a trade-off analysis of model perturbations and differential travel time residuals. The unweighted and weighted RMS station-pair differential time residuals decrease from 1.98 s and 1.57 s to 0.91 s and 0.61 s, respectively, after alternating 6 simultaneous and 6 location-only iterations and as well as 6 additional station-pair location iterations. These values are more than 10% smaller than the final residuals for the station-pair DD location of NVTs using the same data set (Zhang et al., 2010).

Although the 1D velocity model used in the station-pair DD location method of Zhang et al. (2010) is considerably different from the real velocity structure, the station corrections included in the inversion are able to compensate for most of the velocity heterogeneities and thus make the final differential time residuals at the same order as the station-pair DD tomography. In comparison, the station corrections resulting from DD tomography are generally smaller, indicating part of the velocity heterogeneities compensated for by station corrections in DD location are now resolved by DD tomography.

4. Velocity model resolution and uncertainty analysis

For evaluating the resolution of the Vs model around the tremor zone, we conducted the checkerboard resolution test (Humphreys and Clayton, 1988) and formal resolution matrix

analysis (Aster et al., 2005). For the checkerboard resolution test, positive and negative 5% velocity anomalies are added to the 1D velocity model used in the station-pair DD location method of Zhang et al. (2010) at alternating grid nodes in three directions to create the checkerboard velocity model. Synthetic station-pair differential travel times were generated for the checkerboard velocity model using the same source-receiver geometry as the real data, which subsequently were inverted using the station-pair DD tomography method. For the checkerboard resolution test, the starting model is the same 1D velocity model as used for creating the checkerboard velocity model and the inversion parameters are the same as those used for real data inversion. Figure 3 shows the across-fault cross-sections of recovered checkerboard patterns. The checkerboard resolution tests indicate that the Cholame tremor zone is relatively well resolved by station-pair differential times at depths of 15 to 30 km. Figure 4 shows the across-fault cross sections of diagonal resolution matrix values for station-pair DD tomography, based on Lanczos bidiagonalization with partial reorthogonalization (Zhang and Thurber, 2007). We find that the diagonal resolution matrix values for the Cholame tremor zone are generally greater than 0.6, indicating it is relatively well resolved (Figure 4).

To estimate location and velocity uncertainties, we applied a bootstrap analysis method (Efron and Gong, 1983; Efron and Tibshirani, 1991) by randomly resampling station-pair differential times such that the number of differential times remains the same but some differential times may be selected multiple times and some may not be selected at all. The resampled dataset is then used in a DD tomography inversion with the same inversion parameters as the original dataset. This process is repeated 50 times, which is enough to obtain a good estimation of uncertainties for location and velocity parameters (Efron and Tibshirini, 1993). For station-pair DD tomography, the RMS location uncertainties for the relocated NVTs were 2.89

km in X, 2.40 km in Y and 3.11 km in depth, respectively. There are about 73% of NVTs having absolute location uncertainties less than 3 km (Figure 2c). For the inverted velocity model (Figures 5 and 6) using the selected damping and smoothing parameters, the RMS uncertainty is about 0.10 km/s and the velocity uncertainty is up to 0.20 km/s on some grid nodes (Figure 2d). We also did bootstrap error analysis for station-pair DD location (Zhang et al., 2010). In comparison, the RMS location uncertainties for relocated NVTs from DD location are 3.16 km in X, 2.81 km in Y and 3.39 km in depth, respectively, which are systematically greater than the DD tomography results.

To further check the reliability of the inverted velocity model, we performed the inversion again using only NVTs with location uncertainties less than 2.5 km based on bootstrap analysis results. The newly inverted velocity model (Figure 7) is actually similar to the inverted velocity model (Figures 5 and 6) from the inversion using all the NVTs. The median value of absolute differences for two velocity models is 0.126 km/s, indicating that the inversion results are stable.

5. Results and Discussions

Bootstrap and resolution analyses indicate that the velocity structure of the Cholame tremor zone is well resolved by station-pair differential times at depths of 15 to 30 km (Figures 2-4.). In comparison, tomographic inversion for local P-wave velocity (V_p) structure using earthquakes in the region can only resolve the structure down to 20 km (Thurber et al., 2006). Recently, Zeng et al. (2016) assembled a new dataset including S arrivals from earthquakes and LFEs and thus constructed a 3D V_s model for the area using the DD tomography method. However, their checkerboard resolution test shows that the V_s model can only be resolved down

to around 21 km (Zeng et al., 2016). Our Vs inversion results provide the first direct estimates of the Vs structure within the deep roots of the SAF, made possible by the numerous deep NVT sources and our newly developed station-pair DD tomography method. Inaccuracies in assumed velocity models could also significantly affect event locations. Hence the use of our directly inverted Vs model provides a more accurate picture of NVT locations than previously reported locations based on inferred 1D or 3D Vs models.

Similar to what was obtained for the station-pair DD locations (Zhang et al., 2010), the NVTs are mostly located below the seismogenic zone when using station-pair DD tomography (Figure 1). Actually over 72% of the tremors in the Cholame area are located at depths between 20 and 35 km (Figure 1). This is well below the seismogenic zone in the region and spans the Mohorovicic discontinuity in this part of California (the transition from lower-crust to upper-mantle, estimated to be 22-29 km; McBride and Brown, 1986). Compared to the DD locations, considerable changes can be seen in the DD tomography NVT locations (Figure 2a,b), which are expected as the velocity model errors can affect the seismic event locations (e.g. Michelini and Lomax, 2004). Overall, the new NVT locations are less diffuse and show more sub-structure (Figure 2a, b). The NVT locations are also generally similar to relocated LFE locations (Shelly and Hardebeck, 2010), although the NVT locations are more broadly distributed (Figure 1).

The spatial distribution of NVT locations is notably larger than the distribution of LFEs in the area. Some of this is due to increased scatter in the locations resulting from larger timing uncertainties and the lack of P-wave timing information inherent in the differential time data used for NVTs. However, even after taking this into account, there are several lines of evidence supporting a distribution of NVTs that is larger in depth and across-fault than the distribution of LFEs.

291 Within the volume of interest, there are 35 known LFE family locations, each
292 representing the locations of 1000s of individual LFE events. Yet LFE activity from these
293 families overlaps with NVT activity for only a small fraction of the time (15 to 28%; Zhang et al,
294 2010; Guilhem and Nadeau, 2009). This indicates that considerably more NVTs are taking place
295 than is represented by the LFEs.

296 The repeating source assumption and template-pattern matching approach used for
297 defining LFEs is essentially the same as that used for defining repeating earthquakes (REQs),
298 and it biases the identification and locations of LFEs in a similar way. REQs represent the
299 repeating portion of the overall seismicity (generally less than 30%), which are primarily located
300 within the central core of fault zones (Nadeau and McEvilly, 2004). The remaining non-
301 repeating seismicity occurs both within and laterally distributed outside the central fault core
302 (Templeton et al., 2008). By analogy, LFEs likely represent the repeating portion of NVT
303 activity occurring within the central core of the fault zone at depth, with the non-repeating tremor
304 generating the larger fraction of NVT signal with sources both within and laterally distributed
305 outside the fault core zone (Guilhem and Nadeau 2012).

306 Owing to its more ductile nature, it is also reasonable to expect a broader fault-
307 perpendicular distribution of deformation accommodating relative plate motion in the lower-
308 crust and upper-mantle where tremor occurs. This is supported by the across-fault distribution of
309 our NVT locations and the contrasting occurrence patterns between NVTs occurring to the SW
310 versus those occurring to the NE of the SAF (Guilhem and Nadeau 2009; Zhang and Nadeau,
311 2010).

312 The surface and shallow SAF structure above the Cholame tremor zone is relatively
313 complex, owing to the intersection of the San Juan Fault (SJF) and SAF (Figures 1, 5 and 6) and

the presence of a localized body of serpentinite (Ellsworth, 2008). Our Vs structure and corresponding NVT locations reveal how this complexity manifests itself in the deep roots of the SAF and within the transition zone (Moho) at the boundary between the lower-crust and upper-mantle (Figures 5 and 6). The Vs structure and tremor locations show significant jogs in the deep SAF that indicate a complex non-planar structure, with relatively isolated bodies trending along or across the fault (Figures 5 and 6). Guilhem and Nadeau (2012) and Thomas et al. (2012) found that the recurrence patterns and episodic behavior of LFE families and family groups located along the SAF strike could vary significantly. Nadeau and Guilhem (2009) and Zhang et al. (2010) also found a clear contrast degree of periodic behavior of NVT activity laterally across the SAF in the Cholame tremor zone. These observations support the hypothesis that the temporal behavior of the NVTs are controlled by variations in fault zone properties that distribute themselves both along and across the fault, and such variations are evident in our results.

The Vs model determined from station-pair DD tomography reveals a low velocity anomaly between the SAF and SJF around the depth of 20 km (Figures 5 and 6). The NVTs are clearly associated with the relatively low Vs anomalies. For example, within the tremor generation zone, we see that about twice as many of the SAF NVTs occur where local Vs is below average (Figures 5H and 6G). It is possible that there are small pockets of low Vs within the higher Vs zones that are not resolved by the velocity inversion, and this, in combination with the uncertainties in the tremor locations, raises the possibility that an even greater percentage of tremor activity are located within anomalously low Vs fault zones.

The low velocity zones at depths above 25 km are mostly located between the projected SJF and SAF surface traces (Figure 6A and D) and shift toward the northeast across the projected

SAF trace at greater depth (Figure 6C and F). Given the location uncertainties, a low velocity and active tremor (NVT and LFE) zone appears to extend for at least 8 km vertically and is bounded on the northeast by the downward projection of the SAF surface trace and to the northwest and southeast by relatively higher velocity vertical bodies (Figures 5A,C, D and F). There also appears to be a greater concentration of tremors in a low V_s zone bounded to the southwest by the SAF and to the northeast by a high V_s zone (Figure 5B and E).

The existence of fluids and high fluid pressures can reduce V_s (Nur and Simmons, 1969), and high fluid pressures can also facilitate brittle failure and thus tremors within an otherwise ductile deformation regime (Thomas et al., 2009; Shelly and Hardebeck, 2010). This may explain the association we observe between anomalously low V_s and the NVT locations. Furthermore, bounding high V_s structures around the pockets of tremor activity could act as barriers that may confine fluid migration and allow for the formation of high pore-pressure pockets associated with the tremors.

Evidence of near lithostatic pore-fluid pressures in the generation zone of the Cholame tremors has been inferred from the preferential activation of tremors by external shear-stresses and their relatively insensitive response to normal-stresses imposed from earth tides (Thomas et al., 2009), regional earthquakes (Nadeau and Guilhem, 2009), and surface waves from distant regional and teleseismic earthquakes (Peng et al., 2009; Ghosh et al., 2009; Guilhem et al., 2010). This supports our interpretation of ultra-high fluid pressure within the localized pockets of tremor generation and anomalously low V_s .

A recent magnetotelluric survey around Parkfield and Cholame clearly images low resistivity anomaly zones in the mantle and lower crust adjacent to and southwest of the SAF and Cholame tremor zone (Becken et al., 2011). The anomalous zones appear to represent aqueous

fluid networks whose ultimate fluid source is partially serpentinized deep mantle wedge material from a relict slab of the fossil subduction system that existed in the region before the creation of the SAF transform system (Kirby et al., 2002; Fulton and Saffer, 2009; Kirby et al., 2014). Various geophysical and geological evidences for fossil forearc mantle along the California Coast Ranges have been summarized in Kirby et al. (2014), including the identification of strong PmP reflections related to fossil slab (Brocher et al. 1999), the presence of intermediate Vp structure in the lower crust of the Coast Ranges (Fuis and Mooney, 1990) and the finding of magnetic anomalies in the northern California (Jachens et al. 1995) that are consistent with the presence of a partially serpentinized forearc mantle, and the distribution of many serpentinites along Coast Ranges including the Parkfield/Cholame segment of SAF that could be weakened and mobilized by the release of water from the fossil serpentinized forearc mantle.

Becken et al. (2011) proposed that ongoing fluid generation by dehydration reactions of the deep serpentinized mantle material might result in high-pressure fluids trapped beneath an impermeable cap to the southwest of the tremor zone. A sub-horizontal wedge-shaped low-velocity and high reflectivity zone between these trapped fluids and the tremor zone suggests pathways for the high-pressure fluids to migrate laterally to the northeast into the tremor zone, which continue migrating to the northeast across the SAF in the deeper region based on the shift of LVZ to northeast at the depth of 30 km (Figure 6C). The tremor source region separates itself across the SAF into a southwestern zone having periodic episodes and a northeastern zone with steady, non-episodic behavior (Nadeau and Guilhem, 2009; Zhang et al., 2010). The more periodic tremor episode occurrences are located closer to the fluid source and their episodic behavior could reflect a failure mechanism involving cyclic fluid pressurization and seismic energy release (Becken et al., 2011). Additional support for this model comes from the

horizontal low V_s structure at ~ 25 km depth at the southwestern edge of our V_s model (Figure 5A, B, C). Along the central SAF elevated $^3\text{He}/^4\text{He}$ isotope ratios were detected in surface fluids, indicating some fluids are originated from the mantle (Kennedy et al., 2007). This further shows that mantle and crustal permeability is coupled with the flow of deep mantle derived fluids through the ductile lithosphere (Kennedy et al., 2007).

Studies of LFEs and their moment tensors in Japan and polarization analysis in Cascadia argue that tremor is generated primarily by double-couple shear slip involving little volumetric change. In contrast, focal mechanism analysis of NVT in Japan appear to show that NVT signal is the result of a single-force source mechanism, consistent with fluid flow and volumetric change (Rubinstein et al., 2010). Despite these apparent contrasts, laboratory results have confirmed that volume change alone is capable of generating tremor-like seismic signals (Burlini et al., 2009). And, the mineralogical transitions resulting from the presence of fluids may produce fault zone weakening materials such as chlorite (Imber et al., 1997; Fagereng and Diener, 2011). Combined with high pore pressures, the fault zone may be very weak and can emit seismic signals with the dominant shear-slip component, like LFEs. This could explain the presence of the NVTs and LFEs near Cholame.

We have shown that a strong association exists between the location of NVT activity and localized and distributed zones of low V_s beneath the SAF at Cholame, CA, on a scale of a few kilometers. The degree to which V_s is low in the tremor generation zone likely suggests the presence of high-pressure pore fluids. These findings are consistent with the inferential evidence of the involvement of high-pressure pore fluids in tremor genesis discussed above. The complex yet systematic distributions of the V_s anomalies and NVT locations suggest the existence of pathways and boundaries that may confine the fluids and NVTs to within specific zones.

The fact that strong NVT is not ubiquitous along the deep SAF (Nadeau and Guilhem, 2009; Peng et al., 2009; Shelly and Hardebeck, 2010) suggests that deep high-pore pressure fluid conditions may not exist uniformly at depth along the SAF. Of particular note is the fact that strong NVT and LFE activity has not yet been observed directly beneath the central portions of the Parkfield and Cholame segments where larger earthquakes have occurred. Instead, the NVTs and LFEs tend to concentrate in regions bounding these locked segments, where stress-concentrations accumulate through time (possibly opening up fluid pathways).

The presence of high-pressure fluids in localized zones is consistent with the great sensitivity of tremor activity to subtle deformation and stress changes. This and the tendency for tremor to concentrate more strongly at the boundaries of locked fault segments suggests their potential for signaling changing stress conditions beneath the nucleation zones of large earthquakes with great sensitivity (e.g. Kato et al., 2012; Uchida and Matsuzawa, 2013).

To date, monitoring of SAF tremor activity within the northwest portion of the Cholame tremor zone has revealed anomalous precursory changes in tremor activity rates and migration in the nucleation region of the 2004 M6 Parkfield earthquake (Nadeau and Guilhem, 2009; Shelly, 2009). In the southeast portion of this tremor zone, the locked Cholame segment of the SAF also terminates and is believed to have ruptured as part of and possibly nucleated the great M7.8 Ft. Tejon earthquake of 1857 (WGCEP, 2008). It is possible that anomalous precursory NVT activity may also precede the rupture of this segment. Previous estimates of the average rupture recurrence time of this segment range between 85 and 142 years (WGCEP, 2008), and a more recent study suggests the average recurrence time may be on the short end of this range (88 $\pm$ 41 years) (Akciz et al., 2010). It has now been over 150 years since the segment last ruptured, in

excess of either estimate, indicating the importance of continued close monitoring of tremor behavior in this region.

6. Conclusions

A new station-pair DD seismic tomography method has been developed to directly use the differential tremor arrival times to image the tremor generation zone beneath the San Andreas fault near Cholame, California. The resulting tomographic image shows that the tremors are associated with regions of anomalously low V_s , which are likely caused by high-pressure pore fluids from dehydration reactions. This is the first time that a detailed velocity model has been imaged for the tremor generation zone beneath the SAF and within the transition zone associated with the Mohorovičić discontinuity. Our tomographic results support the hypothesis that the tremors are associated with high pore fluid pressures and they are consistent with observations of tremor modulation by small stress changes induced by regional and teleseismic large earthquakes and Earth tides.

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Figure Captions

Figure 1. (A) Map view of 1243 relocated tremors (red dots) by station-pair DD tomography method with stations (blue triangles) used in this study. Station corrections in seconds are shown as circles (negative) or pluses (positive) centred on each station. The Cartesian coordinate system used in the tomography is shown with a tick (marked as “+”) interval of 10 km on X and Y axes (green). The Cartesian coordinate system is the same as Thurber et al. (2006), with the origin at (35° 57.60’ N, 120° 30.28’ W), the Y-axis closely aligning with the SAF surface trace (139.2° rotation from the North), and the X-axis pointing positive northeast. Faults and California coast are shown as black lines. Parkfield and Cholame are labeled yellow dots. Green box shows the Cholame tremor zone that is resolved using station-pair DD tomography. SAF: San Andreas Fault. SJF: San Juan Fault. (B) Cross-section views of relocated tremors (red dots) with background seismicity from Thurber et al. (2006) shown as grey dots and the 88 LFE locations of Shelly and Hardebeck (2010) shown as blue stars. (C) Histogram of tremor depths.

Figure 2. Comparison of tremor locations and the associated uncertainty bounds as well as Vs uncertainties for the volume of interest (VOI; Defined as [-15, 17] in X and [25, 50] in Y and [15, 40] in depth). Across fault cross sections of tremor locations and uncertainties at (a) Y=32 km and (b) Y=36 km resulting from (left) the initial tremor catalog, (middle) station-pair DD location and (right) station-pair DD tomography.

Figure 3. Across-fault cross sections of recovered checkerboard anomalies for station-pair DD tomography. Tremors within 2.5 km of each cross section are shown as red dots. The true

checkerboard model has alternating positive and negative velocity anomalies in three directions. For the checkerboard resolution test, the starting model is chosen as the 1D velocity model. Only areas showing clear checkerboard patterns are well resolved.

Figure 4. Across-fault cross sections of diagonal resolution matrix values at $Y=25$ km to 50 km for station-pair DD tomography. Tremors within 2.5 km of each section are included as black dots.

Figure 5. Across-fault vertical cross sections of (left) the resolved V_s model within the Cholame tremor zone using station-pair DD tomography and (right) its percent difference with respect to an average V_s depth profile at $Y=30$ km (A and D), $Y=35$ km (B and E), and $Y=40$ km (C and D), respectively. Tremors within 1.0 km of the sections and with bootstrap location uncertainties in the X, Y, and Z directions of less than 2.5 km are shown as black dots. Green stars are locations of LFE families (Shelly and Hardebeck, 2010) within 2.5 km of the sections. The color saturation at each model location is proportional to the corresponding diagonal model resolution matrix value, which is rescaled to range from 0 to 1 from the original value. (G) Depth profiles of V_s (red line refers to V_s from the station-pair double-difference tomography inversion averaged over $[-15, 17]$ in X and $[25, 50]$ in Y at each depth; USGS layered and gradient layer models inferred from V_p using $V_p/V_s = 1.78$ are shown as black and dashed gray lines, respectively); (H) Normalized distribution of the V_s percent difference at tremor hypocenters for tremors located within 1 km of each Y slice. Among 258 tremors, 198 of these, approximately 76.7%, are associated with negative velocity changes,

Figure 6. Depth slices of (left) the V_s model and (right) its percent difference with respect to an average V_s value at each depth for depths 20 km (A and D), 25 km (B and E), and 30 km (C and F). Fault traces of San Andreas (right) and San Juan (left) faults are shown as white lines. Tremors having location uncertainties less than 2.5 km that are located within 2.5 km of the sections are shown as black dots. The color saturation at each model location is proportional to the corresponding diagonal model resolution matrix value, which is rescaled to range from 0 to 1 from the original value. (Bottom, G) Normalized distribution of the V_s percent difference at tremor hypocenters located with 1 km of each depth slice. Among 254 tremors, 176, or 69.3% of these are associated with negative velocity changes.

Figure 7. Resolved V_s model within the Cholame tremor zone using station-pair DD tomography by tremors with location uncertainties less than 2.5 km. (Left) Across fault vertical cross sections of the V_s model at (A) $Y=30$ km, (B) $Y=35$ km, and (C) $Y=40$ km, respectively. (Right) Depth slices of the V_s model at depths (D) 20 km, (E) 25 km, and (F) 30 km, respectively. Other figure notations are the same as those in Figures 5 and 6.

Figures

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