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Key Points:

- We propose a time-evolving approach to conduct traveltimes seismic tomography
- The 1989 Loma Prieta earthquake and 2014 South Napa earthquake both occurred in between high- and low-velocity anomalies
- The time-evolving seismic tomography method is effective in revealing subsurface heterogeneous structures

Supporting Information:

- Supporting Information S1

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Time-evolving seismic tomography: The method and its application to the 1989 Loma Prieta and 2014 South Napa earthquake area, California

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Abstract We propose a time-evolving approach to conduct traveltimes seismic tomography in the 1989 M_w 6.9 Loma Prieta earthquake and 2014 M_w 6.0 South Napa earthquake area, California. The recording period of the chosen seismic data between 1 January 1967 and the day before the 2014 South Napa earthquake is divided into two time windows, separated by the 1989 Loma Prieta earthquake. In each time window the subsurface velocity structure is iteratively updated. Starting from the final model of the first time window, the velocity model has been successively improved throughout iterations in the second time window, indicating that the traveltimes data of later time windows have provided extra information to refine the subsurface images. Strong heterogeneities are observed in the final P wave velocity model. Both of the two large earthquakes occurred at transition zones in between high V_p and low V_p anomalies. In all, this study shows the effectiveness of the time-evolving seismic tomography method.

1. Introduction

Since the advent of seismic instruments that record earthquake ground motion as a continuous function of time in the late nineteenth century [Dewey and Byerly, 1969], seismic data have been critical in advancing studies of earthquake sources and structural heterogeneities of the Earth's interior [e.g., Rawlinson *et al.*, 2010; Liu and Gu, 2012; Thurber, 2014]. For example, the Berkeley Seismological Laboratory and its predecessors at UC Berkeley have operated seismic equipment and been monitoring earthquakes in California and the western U.S. since 1887 (<http://seismo.berkeley.edu/seismo.paper.records.html>). Likewise, the Japan Meteorological Agency has operated a nationwide network and archived seismic data for more than 130 years [Ohtake and Ishikawa, 1995]. In recent years, rapid deployment of seismic arrays worldwide, together with the development of new recording devices, has been increasing the total amount of seismic data at an unprecedented rate [e.g., Shang *et al.*, 2014; Tong *et al.*, 2014a]. The long-history and rapid accumulation of seismic data urge us to find new technologies and approaches to process and interpret the huge amount of data to provide new insights into earthquake source processes and subsurface structures.

Seismic tomography is one of the primary tools for imaging the Earth's interior by extracting information from seismograms [e.g., Liu and Gu, 2012; Tong *et al.*, 2014b; Zhao, 2015]. Since the pioneering work of Aki and Lee [1976], it significantly improved our knowledge of the structural heterogeneities and geodynamic processes of the Earth's interior [e.g., Di Stefano *et al.*, 2011; Tong *et al.*, 2012; Zhao, 2015]. Meanwhile, great strides have been made in improving almost every aspect of seismic tomography techniques in order to obtain higher-resolution subsurface images [e.g., Zhao *et al.*, 1992; Dahlen *et al.*, 2000; Tromp *et al.*, 2005; Tong *et al.*, 2014b]. The evolution of data usage in inversions presents a great illustration. In the early stage only the travel-time residuals of the first P or/and S phases were inverted for subsurface structure [e.g., Aki and Lee, 1976; Dziewonski *et al.*, 1977]; later surface waves, normal modes, reflected and refracted phases, and the full waveform content were progressively added to the data sets for tomographic inversions [e.g., Nakanishi and Anderson, 1982; Zhao *et al.*, 2005; Nicolson *et al.*, 2012; Tape *et al.*, 2009].

In an attempt to effectively utilize the constantly increasing seismic data and hence to obtain reliable tomographic images of subsurface structures, we propose a time-evolving approach to analyze seismic data in

seismic tomography. Contrary to most previous tomographic studies in which arrival times and/or waveform data are selected from a particular time period and simultaneously inverted [e.g., *Aki and Lee, 1976; Zhao et al., 1992; Tape et al., 2009; Tong et al., 2014c; Zhang and Lin, 2014*], the time-evolving approach first divides the time history of seismic recording in the study area into a few time windows (e.g., with lengths of tens of years). Subsurface parameters are iteratively updated through tomographic inversions in each time window, starting from the final model from the previous time window. One of the main advantages of the time-evolving approach is that the information embedded in the seismic data of previous time intervals can be used to construct the starting models for the subsequent time intervals. In this study we apply the time-evolving approach to the ray-based traveltime seismic tomography, the most straightforward, robust, and mature tomographic tool so far [*Zhao, 2015*]. It is worth noting that our newly written ray-based traveltime tomography code uses the highly accurate multistencils fast marching (MSFM) method to solve the eikonal equation [*Hassouna and Farag, 2007*] and then compute traveltime of phases by tracing the raypaths along the gradient direction of the traveltime field. We intend to apply the new time-evolving ray-based traveltime tomography method to investigate the subsurface structure of the 1989 M_w 6.9 Loma Prieta earthquake and 2014 M_w 6.0 South Napa earthquake area, California (Figure 1a).

As one of the most seismically active regions in the world, California is under constant threat from potentially damaging earthquakes [*Field et al., 2015*]. A new earthquake forecast model reminds us that damaging earthquakes are inevitable in California and estimates that the probability of a magnitude 6.7 or greater earthquake in the greater San Francisco Bay area is about 72% in the next 30 years [*Field et al., 2015*]. The 24 August 2014 south Napa earthquake with a magnitude of 6.0 is the largest earthquake in over 25 years to strike the greater San Francisco Bay area since the 17 October 1989 M_w 6.9 Loma Prieta earthquake (Figure 1a). The source area of the two large earthquakes is macroscopically within an 80 km wide set of major north-northwest-trending faults of the San Andreas fault system, marking the transform boundary between the Pacific and North American plates (Figure 1a) [*Brocher et al., 2015*]. The 1989 Loma Prieta earthquake was the first major event along the San Andreas Fault since the 1906 M_w 7.8 San Francisco earthquake [*Lin and Thurber, 2012*]. The 2014 South Napa earthquake occurred on the west Napa fault with a right-lateral strike-slip focal mechanism. It caused a surface rupture of approximate 12.9 km long and a maximum surface slip up to 46 cm [*Brocher et al., 2015*]. As a way of testing the proposed time-evolving traveltime seismic tomography method and also investigating the relationship between structural heterogeneities and large earthquake occurrence, we reveal the 3-D seismic velocity structure of the 1989 Loma Prieta earthquake and 2014 South Napa earthquake area with the proposed time-evolving technique in this study. The tomographic results may provide useful hints for potentially damaging earthquakes around the San Francisco Bay area in the future.

2. Methodology

For a given velocity model $c(\mathbf{x})$ and earthquake origin time and location $(\tau, \mathbf{x}_s)$, the arrival time field of a particular phase $t(\mathbf{x})$ can be implicitly described as $t(\mathbf{x}) = t(c(\mathbf{x}), \tau, \mathbf{x}_s)$. The traveltime field $t(\mathbf{x}) - \tau$ satisfies the eikonal equation under high-frequency assumption [*Aki and Richards, 2002*]:

$$|\nabla [t(\mathbf{x}) - \tau]| = \frac{1}{c(\mathbf{x})}. \quad (1)$$

Except for the arrival time field $t(\mathbf{x})$ at very limited number of receiver locations $\mathbf{x}_r^i (i = 1, 2, \dots, N)$, the structure parameters $c(\mathbf{x})$, and source parameters τ , and $\mathbf{x}_s$ are always unknown. To determine these parameters from arrival time measurements $t(\mathbf{x}_r^i)$, we usually begin with an initial estimates $c_0(\mathbf{x})$, τ_0 , and $\mathbf{x}_s^0$ and iteratively update the differences $\Delta c(\mathbf{x}) = c(\mathbf{x}) - c_k(\mathbf{x})$, $\Delta \tau = \tau - \tau_k$, and $\Delta \mathbf{x}_s = \mathbf{x}_s - \mathbf{x}_s^k$, $k = 0, 1, 2, \dots$, based on the variation formula

$$\delta t(\mathbf{x}) = \delta c(\mathbf{x}) \frac{\partial t}{\partial c} \Big|_{c(\mathbf{x})=c_k(\mathbf{x})} + \delta \mathbf{x}_s \cdot \frac{\partial t}{\partial \mathbf{x}_s} \Big|_{\mathbf{x}_s=\mathbf{x}_s^k} + \delta \tau \frac{\partial t}{\partial \tau} \Big|_{\tau=\tau_k}. \quad (2)$$

There are two key components in the above process: (1) accurately solving the eikonal equation (1) to obtain the coefficients in the relationship (2), and (2) preparing good estimates of $c_0(\mathbf{x})$, τ_0 , and $\mathbf{x}_s^0$ which are close enough to the “true” values so that the expansion in (2) is correct to first order.

In this study we use the multistencils fast marching (MSFM) method [*Hassouna and Farag, 2007*], a highly accurate and improved version of the isotropic fast marching method, to solve the eikonal equation. A detailed

Figure 1. (a) The tectonic conditions and surface topography around the San Francisco area. The study area is within the blue box. The local X axis and Y axis are along P_1P_4 and P_1P_2 , respectively. The black star represents the epicenter of the 2014 South Napa ($M_w = 6.0$) earthquake. The red star is the 1989 M_w 6.9 Loma Prieta earthquake. Major active faults denoted by grey curves are San Andreas Fault, San Gregorio Fault (SGF), Rogers Creek Fault (RCF), Hayward Fault (HF), West Napa Fault (WNF), Green Valley Fault (GVF), Calaveras Fault (CF), and Greenville Fault (GF). The 16 red dots indicate the surface centers of the 1° cells of CRUST1.0 around the study region. (b–c) Three-dimensional distributions of the selected earthquakes (red dots) in the two time windows. The seismic stations operated in each time window are denoted as blue inverse triangles.

description of the MSFM method and comprehensive numerical examples showing its accuracy can be found in *Hassouna and Farag* [2007]. After the traveltimes field $t(\mathbf{x}) - \tau$ is obtained, we trace the raypath $\gamma(\mathbf{x}_r, \mathbf{x}_s)$ from the receiver $\mathbf{x}_r$ to the source $\mathbf{x}_s$ along the negative gradient direction of traveltimes field.

A time-evolving approach is proposed as an attempt to build reasonably well starting or reference models $c_0(\mathbf{x})$. We assume that the seismic data recording period coincides with the time window of $[T_0, T_M]$ and divide it into M sequential windows as $[T_m, T_{m+1}]$, $m = 0, 1, \dots, M - 1$. The velocity model $c_k^m(\mathbf{x})$ ($k = 0, 1, \dots, k_m$) is iteratively updated from a starting model $c_0^0(\mathbf{x})$ by inverting the traveltimes data measured in the time window $[T_m, T_{m+1}]$. For the first time window, the starting model $c_0^0(\mathbf{x})$ should be carefully constructed by integrating information such as the well-established results of previous imaging studies, known geological structures, and well logs (recorded usually for exploration geophysical purposes). For the m 'th ($m > 0$) time window, the starting model $c_0^m(\mathbf{x})$ for the tomographic inversion is selected to be the final model $c_{k_{m-1}}^{m-1}(\mathbf{x})$ for the previous time window which explores the traveltimes information in all previous time windows. Seismic activities may slightly change over the time partially due to the changes in material properties. But the data sets in different time windows actually contain similar coverage on subsurface structures at the regional scale of this study. If the first starting model $c_0^0(\mathbf{x})$ is well prepared, the later starting models $c_0^m(\mathbf{x})$ ($m > 0$) should be close to the real velocity model. In addition, as data sets in different time windows are independent of each other, the time-evolving approach provides a way to image the subsurface starting from improved initial models for later time windows. The earthquake parameters τ_0 and $\mathbf{x}_s^0$ estimates can also benefit from an improved initial model if they are updated as well.

3. The Data and Starting Model

The U.S. Geological Survey has been operating the Northern California Seismic Network and providing high-quality earthquake data for a wide range of scientific research and hazard reduction activities since 1967 (<http://www.ncedc.org>). To test the proposed time-evolving tomographic inversion algorithm, we divide the recording period into two time windows. (1) 1 January 1967 to 16 October 1989 and (2) 17 October 1989 to 23 August 2014. The end dates of the two time windows are 1 day before the 1989 Loma Prieta earthquake and the 2014 South Napa earthquake, respectively. To ensure the quality of the selected seismic data used in tomographic inversions, earthquakes and corresponding P wave arrivals are carefully selected based on the following four criteria.

1. We only choose earthquakes ($M_w > 2.0$) with more than eight P arrivals to reduce the influence of mislocation errors on tomographic inversion.
2. The focal depth of each chosen earthquake is greater than 2 km but less than 35 km.
3. We divide the research area (the brown box in Figure 1a down to the depth of 40.0 km) into $2.0 \text{ km} \times 2.0 \text{ km} \times 0.5 \text{ km}$ blocks and only one event per block recorded by the maximal number of stations is selected to avoid earthquake clustering and ensure a relatively uniform distribution of hypocenter locations.
4. The misfit between the observed arrival time and the synthetic arrival time in the starting reference model (discussed later) is required to be less than the minimum value of 4 s and 2 times the synthetic arrival time.

As a result, we selected 148,940 first P wave arrival times recorded by 156 stations from 4016 local earthquakes in the first time window (Figure 1b), and 261,560 first P wave arrival times recorded by 396 stations from 3915 local earthquakes in the second time window (Figure 1c). Generally speaking, the distribution of the selected earthquakes in each time window is similar to each other and mainly along the major faults (Figures 1b and 1c). A closer observation can be made that the seismicity along the southern segment of the San Andreas Fault in the current study area was more active in the second time window (Figures 1b and 1c), but the seismicity along the southern end of the Calaveras Fault was reduced (Figures 1b and 1c). The number of seismic stations has been stably increased in the past 50 years, giving much more first P wave arrivals with even less earthquakes in the second time window. Overall, a total of 410,600 first P arrivals generated by 7931 earthquakes are used in this time-evolving tomographic study.

A good starting model is essential for the convergence of the tomographic inversion [Virieux and Operto, 2009]. We build our starting model by inverting all the 1,124,538 P wave arrivals archived by the Northern California Earthquake Data Center (NCEDC) for earthquakes with magnitudes greater than 2.0 in the study area. Beginning with a five-layer model as shown in Table 1, we first invert all these P wave arrivals for a more accurate velocity in each layer. After that, we incorporate the Moho discontinuity into the layered model by extracting and interpolating the Moho depth information from CRUST1.0, a 1° by 1° global crustal model

Table 1. One-Dimensional Layered Velocity Models^a

Depth (km)	<3	[3, 7)	[7, 12)	[12, 18)	≥18
Initial	4.800	5.300	5.800	6.300	6.800
Inverted	4.859	5.4090	5.8551	6.2827	6.9256

^aStarting from the initial model, the inverted model is obtained by inverting all the 1,124,548 *P* wave arrivals archived by NCEC for earthquakes with magnitudes greater than 2.0 in the study area. The inverted 1-D layered model is further used to build up the starting model for the time-evolving tomographic inversion of this study.

(<http://igppweb.ucsd.edu/~gabi/rem.html>). The *P* wave velocity in the mantle is set as 8.0 km/s. To have a smoothly varying velocity model, the velocity across each discontinuity z_{dis} is further smoothed by using the interpolation function $v(z) = v_1 + (v_2 - v_1)/[1.0 + \exp(z_{dis} - z)]$ if $z < z_{dis}$ or $v(z) = v_2 + (v_1 - v_2)/[1.0 + \exp(z - z_{dis})]$ if $z > z_{dis}$. Here v_1 and v_2 are the *P* wave velocities above and below the discontinuity at z_{dis} , respectively. The smoothed velocity model will be used as the initial starting model (starting model of the first time window) for the time-evolving seismic tomography.

4. Tomographic Inversion and Results

To conduct the tomographic inversion, we need two sets of grid nodes: One is for representing the 3-D velocity model and solving the eikonal equation; the other is the inversion grid used to parameterize the velocity perturbation field $\delta c(\mathbf{x})$ in equation (2). The first set of grid should be much finer, and, for our case, uses an equal interval of 0.6 km to ensure the accuracy of ray tracing. The inversion grid has coarsely and regularly sampled nodes with a lateral spacing of 20.0 km and consisting of eight vertical layers located at the depths of −100, 0, 4, 9, 15, 22, 30, and 60 km.

We introduce WiMj ($i = 1, 2, \dots, j = 0, 1, \dots$) to label the j th velocity model in the i th time window. For the first time period between 1 January 1967 and the 1989 Loma Prieta earthquake, the *P* velocity model is iteratively improved by inverting the traveltimes data picked over this time window (Figure 1b), using the smoothed velocity model discussed in the previous section as the starting model W1M0. A damping parameter is used to regularize the inverse problem and keep the perturbed model close to the starting model [Tong *et al.*, 2014b]. Rather than using the common *L* curve criteria to select an optimal damping parameter [e.g., Calvetti *et al.*, 2000; Tong *et al.*, 2014c], we simply choose a damping parameter which gives the maximum relative velocity perturbation $\delta c(\mathbf{x})/c(\mathbf{x})$ less than 2% at each iteration. This choice of 2% is subjective and empirical. With such a small perturbation, the raypaths only deviate slightly from the previous ones after one iteration. We stop at the twentieth iteration when no significant reduction in the root-mean-square value of the traveltimes residuals is observed. Model W1M20 is therefore considered to be the final velocity model from the first time window and used as the starting model W2M0 for the second time window (Figure 2). Repeating the iterative inversion process in the second time window, we obtain the final velocity model M2W20 of this time-evolving tomographic inversion study. The only difference is that in the second time window the earthquake locations (origin times and hypocenter locations) are simultaneously updated based on equation (2).

The planar view of the *P* wave velocity model at the hypocenter depth (11.23 km, the original depth is 11.12 km and the relocation is conducted in the final model W2M20) of the 2014 M_w 6.0 South Napa earthquake in Figure 2 demonstrates the process of the time-evolving tomographic inversion. We can observe that both high- and low velocity anomalies gradually appear along the north-northwest trending faults of the San Andreas fault system over iterations. The high- and low-velocity structures are roughly separated by West Napa Fault, the northern segment of Hayward Fault, and the southern part of San Andres Fault in the study area. The final model W1M20 (Figure 2e) obtained in the first time window is slightly different from the one W2M20 (Figure 2j) generated in the second time window. For example, the low-velocity anomalies in W2M20 span a wider area than those in W1M20. The differences between W1M20 and W2M20 may partially reflect the temporal *P* wave velocity variations, and they probably also indicate that seismic data of the second time window have provided extra information and further improved the velocity model. Examining the final model W2M20 (Figure 2j), we can also observe that earthquakes at the depths between 10 km and 13 km mainly occurred in low-velocity structures and along the faults. Specifically, the South Napa earthquake took place at the boundary of a relatively high velocity area and was very close to low-velocity anomalies as shown in Figure 2j.

Figure 3 shows the horizontal slices of the final 3-D seismic velocity model W2M20 at 10 representative depths from 3 km to 21 km. We can observe strong lateral heterogeneities with about $\pm 10\%$ velocity variations. At the

Figure 2. Map views of the iteratively updated P wave velocity model at the depth of 11.23 km, the same depth to which the 2014 South Napa earthquake was relocated in the final model W2M20. The absolute velocity is indicated by the color scale at the bottom. Blue and red colors denote high and low velocities, respectively. The velocity models are generated in the (a–e) first and (f–j) second time windows. Model W1M20 of the first time window is the starting model of the second time window. The black star represents the hypocenter of the 2014 South Napa earthquake. The white dots are the earthquakes at the depths between 10 km and 13 km occurred in the corresponding time window.

shallow depths, the velocity structure correlates well with the surface geology. For example, in Figures 3a–3c the low-velocity anomalies to the west of the San Andreas and Hayward Faults reflect the Cenozoic sedimentary rocks, the high-velocity bodies between the San Andreas and the Hayward-Rogers Creek Faults indicate outcrops of the Franciscan terrane, the high-velocity structure under the northern portion of the study area is associated with the Cenozoic volcanic rocks, and the low-velocity regions to the east of the Hayward-Rogers Creek Faults are the Cenozoic basins [Hole *et al.*, 2000]. To the east of the Calaveras Fault, the high velocities in the tomography model at 3–5 km depth (Figures 3a and 3b) can be explained as the existence of outcrops

Figure 3. Map views of the final *P* wave velocity model (W2M20) at 10 representative depths. The layer depth is labeled at the left top corner of each plot. The range of velocity change at each depth is shown in the color scale below. Red and blue colors denote low and high velocities, respectively. The black stars in Figures 3d–3f are around the hypocenter (at 11.23 km depth) of the 2014 South Napa earthquake. The red stars in Figures 3g–3i are around the hypocenter (at 16.78 km depth) of the 1989 Loma Prieta earthquake. Earthquake hypocenters within 1 km of the layer are shown as white dots.

Figure 4. Vertical cross sections of the relative P wave velocity perturbation field (in percent) $\delta V_P/V_P^{1D}$ along profile AA', BB', CC', DD', and EE' through (a–e) the final model of the first time window and (a'–e') the final model of the second time window. The locations of the five profiles are indicated on the (f) inset map. (g) The 1-D velocity model V_P^{1D} calculated by averaging the initial starting model W1M0 at each depth. The black and red stars denote the South Napa earthquake and Loma Prieta earthquake, respectively. Earthquake hypocenters in each time window and within 5 km of the profile are shown as white dots on the corresponding plot.

of the Franciscan terrane as well. Starting from the upper crust and persisting to about 11 km depth, strong velocity contrasts can be observed across the northern and middle portions of the Hayward Fault (Figures 3a–3e), which reflects the boundary between the west high-velocity Franciscan terrane and the east low-velocity Great Valley Sequence [Zhang and Thurber, 2003; Hardebeck *et al.*, 2007]. In the southern part of the study area and from about 7 km depth to the lower crust, a general high- to low-velocity change is observable across the San Andreas Fault (Figures 3c–3j). This is mainly because of the Salinian terrane to the west of the San Andreas Fault is faster than the Franciscan terrane to the east [Hole *et al.*, 2000]. The prominent feature of our tomographic results is that a low-velocity structure exists to the east of the Hayward-Rogers Creek Faults and almost passes through the whole crust (Figure 3). In the middle and lower crust, this low-velocity anomaly extends westward below the San Andreas Fault (Figures 3f–3j). Hole *et al.* [2000] and Thurber *et al.* [2007] also mapped a region of relatively low velocity around the Hayward Fault in the middle crust and reaching the western edge of the Great Valley in the lower crust, which was explained by the eastward thickening of the crust. Meanwhile, in our tomographic results we can observe that earthquakes distributed mainly along active faults or within the low-velocity zone. We can also find that almost all earthquakes occurred in regions with P wave velocity less than 6.3 km/s. This is generally consistent with the finding of Hole *et al.* [2000] that no earthquakes were found in regions with P wave velocity greater than 6.3 km/s and the finding of Thurber *et al.* [2007] that only about 11% of the earthquakes occurred at places with $V_p > 6.3$ km/s. The 2014 South Napa earthquake was initially located at a depth of 11.1 km and then relocated at 11.2 km depth. In Figures 3d–3f, we can observe that the 2014 South Napa hypocenter is located at the boundary of a relatively high velocity zone and almost in between a high- and a low-velocity anomaly. The hypocenter location of the 1989 Loma Prieta earthquake (relocated from 17.21 km depth to 16.78 km depth) also shows a strong velocity contrast feature (Figures 3g–3j), indicating that the 1989 Loma Prieta earthquake may similarly occur in between a high- and a low-velocity anomaly.

Figure 4 shows the vertical views of the relative velocity perturbation field $\delta V_p/V_p^{1D}$ across five different profiles (Figure 4f). Here $\delta V_p = V_p^{3D} - V_p^{1D}$, V_p^{3D} is the final three-dimensional model W1M20 of the first time window (Figures 4a–4e) or the final model W2M20 of the second time window (Figures 4a'–4e'), and V_p^{1D} is calculated by averaging the initial starting model W1M0 at each depth (Figure 4g). The final models of the first and second time windows show similar features. For example, across all the five profiles significant low-velocity anomalies exist in the lower crust. High-velocity structures are mainly distributed in the upper and middle crust. The majority of seismic activities are in and around the high-velocity regions, which may be explained as the seismogenic zone. The hypocenter areas of the 2014 South Napa earthquake and the 1989 Loma Prieta earthquake have strong velocity variations with obvious high- and low-velocity anomalies around. Specifically, the 2014 South Napa earthquake occurred at the edge of a high-velocity body with a significant low-velocity anomaly below (Figures 4a', 4b', and 4e'). The 1989 Loma Prieta earthquake took place immediately below a high-velocity anomaly and at the top boundary of a low-velocity anomaly which has a deep root into the mantle (Figures 4d–4e). Similar findings about the 1989 Loma Prieta earthquake were reported by previous studies [e.g., Lees, 1990; Takauchi and Evans, 1995; Lin and Thurber, 2012]. Based on these results we may claim that transition zones in between high V_p and low V_p anomalies are apt to generate large crustal earthquakes. The differences between the final models of the first and the second time windows are also observable in Figure 4. We can find that the shapes and amplitudes of the velocity anomalies have been slightly changed after the traveltimes inversion in the second time window. The P wave velocity structure around the 2014 South Napa hypocenter seems to be almost the same in the final models of the two time windows (Figures 4a and 4a', 4b and 4b', and 4e and 4e'). But around the source area of the 1989 Loma Prieta earthquake, the velocity and size of the low-velocity anomaly become slower and bigger, and significant low-velocity anomalies near the surface also appear in the second time window (Figures 4d and 4d', and 4e and 4e'). There are probably two reasons for the velocity change around the 1989 Loma Prieta hypocenter.

1. The velocity is indeed changed after this great earthquake because of the dynamic process of the hypocenter area.
2. The velocity remained almost the same, but the increased number of earthquakes after the Loma Prieta main shock improves the illumination of the source area.

Additionally, a resolution test is conducted to estimate the reliability of the obtained tomographic results as well as the time-evolving approach. We assume that the true model has a uniform 6% P wave velocity perturbation from the initial starting model W1M0. Following the same procedure, the data sets in the two time windows are sequentially inverted to recover the high-velocity structure. The only difference is that the

“observed” traveltimes here are calculated in the true model. Random errors with a standard deviation of 0.1 s are added to the observed data to mimic the noises in the real data. The velocity model is iteratively updated in each time window. The inverted results of the two time windows are shown in Figures S1–S2 in the supporting information documents. Generally speaking, thanks to the good starting model provided by the first time window, the result obtained in the second time window is much closer to the true velocity model. This suggests the effectiveness of the time-evolving approach. However, due to the incomplete data coverage, the uniform 6% *P* wave velocity perturbation is not uniformly recovered. For example, near the boundaries, anomalies are only partially recovered to 3% (Figure S2). But structures in most regions especially around the hypocenters of the 1989 Loma Prieta earthquake and the 2014 South Napa earthquake are almost fully recovered. Overall, this resolution test indicates that the tomographic results obtained by using the time-evolving approach are reliable and can be used for interpretation.

5. Discussion and Conclusions

As an attempt to fully explore the large amount of seismic data spanning a period of about 50 years, a time-evolving traveltime seismic tomography method is proposed and applied to recover the crustal structure of the 1989 Loma Prieta earthquake and 2014 South Napa earthquake area, California. One of the advantages of the time-evolving approach is that models that explore traveltime data from previous time windows can serve as the starting models of later time windows. The data sets of previous time windows provide strong constraints on the subsurface structure and are very essential for constructing a “closer” starting model to the true model for the tomographic inversion in the subsequent time window.

Our tomographic results are generally consistent with previous studies in overlapped regions [e.g., Lees, 1990; Hole et al., 2000; Zhang and Thurber, 2003; Hardebeck et al., 2007; Thurber et al., 2007]. Strong heterogeneities exist in the crust, especially near the surface. The crustal seismogenic zone is characterized as a high-velocity zone. We find strong velocity variations around the source areas of both the 1989 Loma Prieta and 2014 South Napa earthquakes. Significant low- and high-velocity anomalies are around the two hypocenters (Figure 4). We may infer that the occurrences of the two great earthquakes in our study area have a close relationship with the crustal heterogeneities.

This time-evolving seismic tomography technique can be applied to other regions with long histories of seismic recordings, such as Southern California, Alaska, and Japan. With the fast accumulation of seismic data at an unprecedented speed nowadays, we believe that the time-evolving approach will have wider applications in the future. The time-evolving approach is procedurally similar to the 4-D time-lapse seismic imaging [Asnaashari et al., 2015]. We also find obvious velocity changes around the source area of the 1989 Loma Prieta earthquake in the two time windows separated by this large earthquake. However, so far we are lack of supporting evidences to claim that the time-evolving seismic tomography can be used to monitor the subsurface dynamic processes and related changes of velocity structures as the 4-D time-lapse seismic imaging does. Whether this time-evolving technique can be used effectively to capture the time-varying subsurface structures in regional tomography studies remains to be further investigated.

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