Systematic Analysis of Shear-Wave Splitting in the Aftershock Zone of the 1999 Chi-Chi, Taiwan, Earthquake: Shallow Crustal Anisotropy and Lack of Precursory Variations

by Yunfeng Liu, Ta-Liang Teng, and Yehuda Ben-Zion

Abstract We analyze shear-wave splitting (SWS) in a high-quality waveform data set recorded at surface and downhole (0.2 km) seismometers in a region around the 20 September 1999 $M_w$ 7.6 Chi-Chi, Taiwan, earthquake sequence. The data set was generated by events in a 5-year period before, during, and after the mainshock. The purpose is to investigate the depth extent of the crustal anisotropy and its possible temporal evolution in relation to the occurrence of large earthquakes. Results from downhole records show a stable polarization direction of the fast shear wave that matches well the local Global Positioning System (GPS) velocity field. A slightly different polarization direction of the fast shear wave is obtained from surface data. This suggests a possible anisotropy change between the top 0.2 km structure and the deeper section of the crust. Measured time delays below the downhole station have an average value of 0.16 sec without systematic changes for sources from about 8 km to 20 km in depth. Estimates of time delays in the top 0.2 km of the crust based on shear waves reflected from the free surface give a constant 0.04 sec. A likely depth distribution inferred from these two types of measurements and an $S$-velocity model indicates that the crustal anisotropy in the region is dominated by the top 2 to 3 km. The measured polarization directions and time delays give essentially constant values over the 2.7-year premainshock and 2.3-year postmainshock periods in the region adjacent to the Chi-Chi rupture and within 10 km from the epicentral region of its two large $M \geq 6.0$ aftershocks. Analysis of SWS in waveforms produced by earthquake multiplets confirms further the lack of precursory temporal variations of crustal anisotropy in the immediate neighborhood of the Chi-Chi earthquake sequence. The results raise doubts on the general usefulness of SWS measurements for earthquake forecasting. An apparent coseismic increase in anisotropy time delay of approximately 10% is observed for depths $>0.2$ km; however, this value is clearly affected by spatial changes associated with different event locations before and after the Chi-Chi mainshock.

Introduction

The seismic anisotropy of the upper crust has been studied extensively, mostly based on analysis of shear-wave splitting from local earthquakes. The commonly used explanation for the upper-crust anisotropy is the assumed “extensive dilatancy anisotropy” (Crampin, 1978) or its modified version “anisotropic poroelasticity” (Crampin and Zatsepin, 1997; Zatsepin and Crampin, 1997). The model ascribes the upper-crustal anisotropy to a preferred orientation of vertical, fluid-filled microcracks aligned in a direction controlled by the in situ stress. In this hypothesis, shear waves with particle motions parallel to the plane of the cracks travel faster than those polarized in the orthogonal direction. Fluid-filled microcracks, oriented parallel to the maximum horizontal compressive stress direction ($\sigma_H$), will preferentially remain open and it is thus expected that the polarization direction of the fast shear wave will be parallel to $\sigma_H$. Other possible causes of shallow anisotropy include alignments of microcracks or minerals in fault vicinities (Leary et al., 1987; Zhang and Schwartz, 1994; Zinke and Zobak, 2000) and rock fabric anisotropy due to preferred mineral orientation (Kern and Wenk, 1990; Aster and Shearer, 1992).

In this study, we analyze shear-wave splitting (SWS) in seismic waveform data recorded by a downhole short-period station CHY and a surface strong-motion station CHY073.
located in the vicinity of the Chelungpu fault associated with the 20 September 1999 Chi-Chi earthquake (Fig. 1). Numerous aftershocks (with two having $M \geq 6.0$) occurred just beneath the stations about one month after the Chi-Chi mainshock. The purpose of the study is to investigate the depth extent of stress-induced crustal anisotropy and its possible evolution in time in relation to large earthquakes. High-quality borehole records are used to measure SWS more precisely than can be done with surface records. Clearly recorded surface reflections in downhole data along with surface records allow us to estimate the magnitude of near-surface crustal anisotropy. The aftershock sequence in this region lasted more than 2 years and many recorded events also occurred during a 2.7-year period before the Chi-Chi mainshock. We analyze SWS in this data set using both the aspect ratio method (Shih et al., 1989) and the cross-correlation method (e.g., Fukao, 1984). The results indicate that the crustal anisotropy in this region is dominated by the top 2–3 km and that the top 0.2 km accounts for about 20% of the total observed SWS time delay in the upper crust. The results do not show systematic premainshock or systematic postmainshock temporal changes of SWS parameters in the source regions of the examined large earthquakes over the 5-year study.

Data Set and Geologic Background

Modern digital seismic monitoring in Taiwan began in the early 1970s. The Taiwan Central Weather Bureau Seismic Network (CWBSN) now operates 75 telemetered stations. Each station has short-period and strong-motion three-component sensors. A few of these stations are installed in boreholes reaching to about 0.2 km in depth. Since the beginning of the 1990s, the Taiwan Strong-Motion Instrumentation Program (TSMIP) added more than 650 additional free-field stations with modern digital instruments (Shin and Teng, 2001). The Chi-Chi earthquake sequence was unusually energetic, with many $M \geq 6.0$ aftershocks, two of which occurred in the study area. The parameters of these two events and the mainshock are listed in Table 1. The locations

Figure 1. A location map of the study region with the Meishan fault (MSF), the Chelungpu fault (CLF), and the Chukou fault (CKF). Solid triangles indicate short-period stations including a 200-m-deep downhole station CHY. Solid stars represent the 20 September 1999 $M_w$ 7.6 Chi-Chi earthquake and its two large aftershocks. Filled and open circles represent other small aftershocks recorded by the borehole station CHY and the strong-motion station CHY073, respectively.
of the mainshock (larger filled star), two large aftershocks (smaller filled stars), and other aftershocks (open and filled circles) are shown in Figure 1. The filled triangle indicates the location of short-period stations. One of these stations, CHY, is installed in a 0.2-km-deep borehole. There is also a surface strong-motion station, CHY073, in the same location as CHY. The sampling rate of strong-motion seismograms is 100 samples/sec (sps), and the short-period records of CHY are sampled at 50 sps. The data used in this study extend from January 1997 to March 2002.

Our study region is on the eastern boundary of the west coast Holocene alluvium plain, southwest of the southern

<table>
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<th>Longitude (E)</th>
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Table 1
The Parameters of the 1999 Chi-Chi Earthquake and Its Two Large Aftershocks

In general, particle motion for simple shear-wave splitting is elliptical but it will be linear under one of the following two conditions: (1) one shear-wave component is zero, or (2) the phase difference between the two components is $\pi n$. According to Shih et al. (1989), the AR is defined as:

$$AR_j = \frac{\sum_{i=1}^{n-1} d_i \cos(\theta_j - e_i)}{\sum_{i=1}^{n-1} d_i \sin(\theta_j - e_i)},$$

(1)

where $n$ is the total number of samples at equal time steps, $\theta_j$ is the angle between the fast trial axis and $x$ axis, and $j$ is the index of the angle increment. The parameters $e_i$ and $d_i$ are

$$e_i = \tan^{-1} \left( \frac{y_i - y_{i-1}}{x_i - x_{i-1}} \right),$$

(2a)

with $i$ being sample number and

$$d_i = \sqrt{(y_i - y_{i-1})^2 + (x_i - x_{i-1})^2}.$$  

(2b)

where $x$ and $y$ are two orthogonal velocity time series.

The maximum value of the aspect ratio in a time interval that contains only the early shear-wave arrival indicates the most linear particle motion (condition 1) and its direction indicates the PD of the FSW. The ideal time window for this aspect ratio calculation should start at the onset of the FSW and end right before the arrival of the SSW. Although it is not easy to determine such a window, multiple windows with different sizes can be tested to find a suitable one. Shih et al. (1989) also determine the TD by maximizing the aspect ratio as a function of time lag based on condition 2. Here we estimate the TD by cross-correlating the FSW with the SSW waveforms as determined by condition 1. According to Gledhill (1991), the measure error of the PD of the FSW is estimated by $\pm \tan^{-1} (1/AR)$. The confidence interval estimation for the CC method is discussed in Rau et al. (2000).

We adopt the same scheme to estimate the measurement errors for all parameters calculated in this study with a 95% confidence level.

Figure 2 illustrates an example of SWS analysis by using the AR method. Figure 2a shows original three-component seismograms from downhole velocity sensors at CHY, generated by an $M$ 1.88 event at 10.70 km depth with 1.70 km epicentral distance (see information above the figure). The shaded areas indicate the time window for data analysis chosen to bound the direct shear-wave signal. We estimate the PD of the FSW, which in this example gives 170° ± 1°, from the maximum value of the AR shown in Figure 2b. Then we rotate the horizontal seismograms into the determined PDs of the FSW and the SSW (Fig. 2c). The CCCs (Fig. 2d) of the two split seismograms are used to determine the TD of...
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Figure 2. An example of SWS analysis with the AR method using short-period data. (a) Original three-component seismograms. (b) A contour of the AR values. The maximum value indicated by the cross corresponds to the PD of the FSW (170° in this example). (c) Seismograms rotated to the fast and slow PDs. (d) The CCC between two split shear waves determined by b. The maximum value is related to the TD (0.16 sec in this example). (e) Waveforms of the fast and slow components shifted with the measured TD. (f) The horizontal particle motion of the original seismograms. (g) The horizontal particle motion of the fast and slow components shifted with the measured TD.

Figure 2f shows the horizontal particle motion of the original seismograms. The abrupt change in the direction of particle motion starts at the 16th data point (indicated by an arrow). As a check, we advance the SSW with the estimated TD and plot the shifted seismograms in Figure 2e and their horizontal particle motion in Figure 2g. The linear particle motion in Figure 2g and well-matched seismograms in Figures 2e indicate that the preceding measurement is valid.

Figure 3 illustrates the SWS analysis with example using the CC method. The original strong-motion data are given in Figure 3a and the contours of the CCC are shown in Figure 3b. The maximum CCC value is indicated by the cross, and its corresponding azimuth and TD give the PD of the FSW (here, 177° ± 16°) and the TD (here, 0.20 ± 0.02 sec). Figure 3d is a cross section of Figure 3b at the determined PD of the FSW. The other panels in Figure 3 correspond to those of Figure 2. The nearly linear particle motion in Figure 3g and well-matched seismograms (with opposite polarity phases) in Figure 3e again indicate a valid measurement.

Estimation of Near-Surface Anisotropy from Borehole Data

Borehole-recorded seismograms at CHY clearly show (Fig. 4) a direct up-going phase (S') and a surface-reflected down-going phase (S'). Figure 5 illustrates schematically the geometry of the direct and surface-reflected waves in the borehole configuration. After the PD of the FSW is determined as discussed in the previous section, we project the horizontal seismograms into the fast and slow components as shown in the inset of Figure 6. The shaded regions indicate the direct (S') and reflected (S') phases, respectively. It is expected that the FSW and SSW travel to the free surface...
and are reflected back to the downhole receivers with separate velocities. We calculate the autocorrelation coefficients of both the FSW and SSW and plot them in Figure 6. The secondary peaks of the autocorrelation coefficients of the FSW and SSW correspond to the best matches between the S' and S phases and their time lags are indicated by $dt_f$ and $dt_s$, respectively. Therefore, $dt_f$ and $dt_s$ give the estimated two-way travel times between the receiver and the free surface for the FSW and SSW, respectively. We can thus calculate the TD of the SSW from the borehole receiver to the free surface within the top 0.2 km of crustal structure as

$$TD = (dt_s - dt_f)/2$$

(3)

The preceding estimate is based on the assumption that the PD of the FSW above the borehole station is the same as that below. Since the estimated TD is not very sensitive to the PD, the measured results will not be appreciably affected by slight variations of the PD.

Data Processing and Uncertainty in SWS Measurements

Figure 4 shows four examples of three-component records of the downhole station CHY. The events number, magnitude, focal depth, and epicentral distance are indicated in the upper left corner of each panel. These records are velocity seismograms bandpass filtered from 1 to 15 Hz. The estimated PD of the FSW and the TD are given over the fourth trace in each panel. We show the resolved FSW and SSW in the lower part of each panel for all four examples. The effect of $S$-$P$ waves conversion at the ground surface can be avoided by using steep rays within the shear-wave window (Nuttli, 1961; Booth et al., 1985). The latter is defined as having ray paths with incident $i_0$ angles less than $\sin^{-1}(V_s/V_p)$. For typical values of $V_s$ and $V_p$ near the top of the crust, $i_0 \approx 35^\circ$. Considering the high-velocity gradient near the surface, we use events with hypocentral distance less than focal depth to assure that their ray paths lie within the shear-wave window. The observed effect of $S$-$P$ wave con-
version is negligible for downhole records; however, we find that the incident angle does affect the accuracy of measurement. A total of 401 events within the shear-wave window satisfying these criteria are used in this study. Of these, 375 are recorded by the downhole station CHY, whereas the others are recorded by the surface strong-motion station CHY073. The strong-motion acceleration data are also band-pass filtered from 1 to 15 Hz and then integrated to velocity seismograms, which are used to measure shear-wave splitting.

We use both the CC and AR methods to analyze all these short-period records and show the results in Figure 7. We rejected measured results with a CCC value of less than 0.7. The open and filled circles give analysis results of 334 events based on the CC and AR methods, respectively. In general, both methods produce similar results. The average differences of the measured TD and PD between these two methods are 0.001 sec and 5.2°, respectively. This indicates that the measurements are robust because the AR and CC methods use different portions of seismograms in the PD calculation as discussed in the previous section. Figure 7 shows that the spread of the PD estimates of the FSW based on the AR method is smaller than that of the CC method, whereas both methods have a similar spread of measurements of the TD. We thus use mainly results from the AR analysis in the next sections.

In many cases, the coda of shear waves makes it difficult to recognize the SSW arrivals, introducing an important source of ambiguity in TD measurements (Aster et al., 1990). However, even without significant effects from coda waves there is still a large scatter in our measurement results. Figure 8a shows the TDs versus incident angles represented by the ratio between epicentral distance and focal depth. We see clearly that the results from larger incident angles display larger scatter. For events with epicentral distance over focal depth less than 0.2, the measured TDs are within the range of 0.13 to 0.20 sec. Figure 8b indicates that results for events with magnitudes between 2.4 to 3.0 have less scatter. This is possibly due to a relatively higher signal level for these records. The measurement accuracy is correlated with both the AR and CCC values. As shown in Figure 7, the scatter
for both TDs and PDs gradually decreases as AR values increase. The TDs are restricted to the range 0.12 to 0.20 for measurements with AR > 50. The CCC represents the degree of similarity between the FSW and SSW. Factors that influence the similarity between the FSW and SSW may include scattering, attenuation, dispersion, and vertical variations of anisotropy. Figures 8c and 8d show the relations between hypocenter positions and TD measurements. We divide the measured TDs into the following three categories: (1) larger than 0.17 sec, (2) between 0.14 and 0.17 sec, and (3) less than 0.14 sec. Each category is distributed approximately over the entire study region. In other words, a scatter also exists in the measured TDs from similar paths. In addition to measurement uncertainty, there may be an inherent scatter in the TDs because of inhomogeneous distribution of the crustal crack density. However, we can not identify heterogeneous patterns of the crack density distribution because of the limited resolution. Such inhomogeneity will not affect the PDs of the FSW, which are controlled by the in situ horizontal stress. The PD measurement is also insensitive to coda “contamination.” Thus, the measured PDs of the FSW have less sources of error and scatter.

Near-Surface Crustal Anisotropy

Figure 9a shows the estimated TDs of split shear waves over the depth range 0–0.2 km from surface-reflected waves. The mean and standard deviation of TD values are 0.04 ± 0.003 sec. We thus have derived two sets of TDs from the same set of borehole records. The TD for the section deeper than 0.2 km (>0.2 km) estimated from the direct phase $S$ is related to the path from the source to the borehole station CHY. The TD for the depth section 0–0.2 km measured from the reflected phase $S'$ characterizes the path from the borehole station CHY to the free surface. Due to attenuation, dispersion and coda contamination, the individual analysis error of TD from the reflected phase $S'$ is much larger than that from the phase $S$. However, in contrast to the large scatter that exists in the measured TDs for >0.2 km, the estimated TDs for 0–0.2 km are much more consistent. Figure 9b shows TDs for >0.2 km versus TDs for 0–0.2 km. We find that there is no correlation between these two sets.
of TDs. This implies that the sources that cause a broader scatter in the measured TDs for >0.2 km do not affect the measurement of TDs over 0–0.2 km. Because the ray paths should be nearly vertical in the range 0–0.2 km, and all reflected waves should go through a nearly identical path, we expect to have a consistent measurement of TD over the range 0–0.2 km. The results suggest again that the scatter of TDs results mainly from inherent heterogeneity of anisotropic medium rather than measurement uncertainty. We find that the TD per kilometer of travel distance is 0.04 sec over the top 0.2 km of the crust, or 0.2 sec/km. Given the quality and quantity of the downhole data used in this study, the results demonstrate clearly that highly fractured rock in the near-surface crust plays a dominant role in seismic anisotropy observed at the surface.

Interpretations of SWS parameters in the presence of two anisotropic layers are discussed by several articles (e.g., Silver and Savage, 1994; Wolfe and Silver, 1998). Because the surface data are insufficient to resolve these different anisotropic layers, we adopt the following scheme to help interpret the SWS parameters derived from surface data. (1) We use the observed east–west seismogram produced by event 99-11121927 and shown in Figure 3a as “original” waveforms. (2) We project the assumed original seismograms into the FSW and SSW polarization directions of the lower layer in an assumed two-anisotropic-layer model (see caption of Fig. 11). (3) We advance the waveform of the FSW with the assumed TD of the lower layer. (4) We perform similar steps to the obtained waveforms of the FSW and SSW using the assumed PD and TD of the upper layer and get calculated waveforms (Fig. 11a) corresponding to the two-anisotropic-layer model. (5) We analyze the SWS parameters...
Figure 8. (a) TDs versus incident angles of ray paths represented by the ratios between epicentral distance and event depth. (b) TDs versus magnitudes. (c) The distribution of three categories of TDs in the north–south profile. (d) The distribution of three categories of TDs in the east–west profile.

Figure 9. (a) Histogram of measured TDs in the top 0.2 km of the crust. (b) Measured TDs in the top 0.2 km versus measured TDs in the deeper section.
from these generated data and compare the results with the parameters of the assumed model. The analysis results indicate that the estimated TD (Fig. 11d) is the sum of the TD values of the two layers and the estimated PD (Fig. 11b) is close to that of the upper layer. We also note that the shape of the resolved slow component is very similar to that of the original observed data given in Figure 3c and 3e. As shown in those figures, in both cases there are small phases (indicated by arrows) before the arrival of the SSW that are not observed in the downhole seismograms (Figs. 1c,e and 4). The earlier small phases are caused by the small vertical variation of anisotropy. As discussed earlier, the PDs from the surface observations represent primarily the PD in the upper layer, whereas the PDs from the downhole records give an estimate of the PD in the layer below the downhole station. The results imply that the anisotropy in the top 0.2 km of the crust differs from that of the deeper section. The average downhole PD of 170.2° is in good agreement with the direction of maximum compressional stress based on the regional Global Positioning System (GPS) data (Yu et al., 2001). The small change of PD in the surface data is possibly associated with a transition from a highly fractured, loosely cohesive material in the top 100–200 m of the crust to a more competent and cohesive rock below or short-scale fluctuations in the near-surface stress field.

Aster and Shearer (1991, 1992) analyzed the downhole and surface PDs and TDs using a similar experimental geometry in the Anza region of southern California with downhole arrays and local earthquake records. They noticed a PD change between the downhole and the surface receivers in station KNW-BH. They interpreted the anisotropy observed at this station as due to a fixed paleostrain alignment of anisotropic minerals and/or microcracks and ascribed the vertical change of PD to a highly weathered near-surface zone. Coutant (1996) studied very shallow shear-wave anisotropy in local earthquake seismograms recorded by a vertical array of accelerometers at Garner Valley in southern California. The results were interpreted in terms of two superposed anisotropy layers: an igneous-rock anisotropy due to microcrack or mineral alignment below 220 m and stress-induced anisotropy related to the San Andreas fault system above 220 m. Results on depth-varying crustal anisotropy were also found by VSP experiments (Winterstein and Meadows, 1991a, b).
Figure 11. SWS analysis of the calculated data based on a two-anisotropic-layer model using the CC method. The model parameters are PD = 170° and TD = 0.16 sec in the lower layer and PD = 177° and TD = 0.04 sec in the upper layer. (a) Waveforms generated from the east–west component of observed seismograms (Fig. 3a) using the assumed model. Panels b, c, f, and g correspond to respective panels in Fig. 3.

m/sec) in the near-surface material can have several wavelengths (e.g., four wavelengths with 10 Hz over 200 m). In this case, surface observations will be strongly affected by the depth-varying anisotropy that appears to commonly exist in the near-surface region.

Depth Distribution of Crustal Anisotropy

Observation of Anisotropy in the Crust Deeper than 8 km

Several studies have attempted to constrain the depth extent of anisotropy in the crust. Zhang and Schwartz (1994) claimed that the seismic anisotropy in the Loma Preita rupture zone is no deeper than 2 km. Gledhill (1991), Peacock et al. (1988), and Savage et al. (1989, 1990) concluded that anisotropy in their study regions must be confined to the upper few kilometers of the crust to explain their observations of very different PDs obtained for stations located only a few kilometers apart. In contrast, Shih and Meyer (1990) observed increasing TDs with propagation distance in the South Moat of the Long Valley Caldera in California, suggesting more pervasive anisotropy than only a few kilometers. Zinke and Zobak (2000) concluded that the shallow crust below their station does not influence the PD near the Calaveras fault in California. Li et al. (1994) observed anisotropy in the seismogenic layer beneath the Los Angeles basin. Because of the scatter inherent in TDs, it is often difficult to constrain well the depth extent of anisotropy. However, the high-quality seismic records observed in the downhole station CHY allow us to constrain it accurately in our study region.

Figures 12a and 12b show the TDs and the PDs of the FSW versus the depth of events. The filled triangles give the average values over 2-km intervals and the vertical lines indicate the associated standard deviations. For instance, the first triangle and vertical line on the left in Figure 12a represent the TD measurements from 6 to 8 km with the values 0.14 ± 0.04 sec. The results indicate that the PDs are very stable over the entire range from 6 to 18 km, and that the TDs are reasonably constant over the range 8 to 18 km. We conclude that the anisotropy in our study area occurs over the top 8 km of the crust. In other words, there is no appreciable crustal anisotropy from 8 to 18 km.
Estimation of Anisotropy in the Crust Shallower than 8 km

As shown in Figure 12a, the average value of TDs from several events with a depth of 6.0–8.0 km is smaller than that from the deeper section. We believe that the change in this range is mainly due to lack of sufficient data points. Unfortunately, no events are recorded with a depth less than 6.0 km, and thus we cannot constrain the depth extent of anisotropy directly in the range from 0.2 to 6.0 km. However, several lines of arguments can be used to infer that the observed crustal anisotropy is actually dominated by the top 2–3 km.

The depth extent of anisotropy may be estimated through the coefficient of anisotropy $k$, which is defined as (e.g., Tadokoro et al., 1999)

$$ k = \frac{(V_{fs} - V_{ss})}{V_{fs}}, \quad (4a) $$

where $V_{fs}$ and $V_{ss}$ are the velocity of the FSW and SSW, respectively. The coefficient of anisotropy $k$ can also be calculated by

$$ k = \frac{(T_{ss} - T_{fs})}{T_{ss}}, \quad (4b) $$

where $T_{fs}$ and $T_{ss}$ are the travel time of the FSW and SSW, respectively. We have two boundary conditions for $k$. In the range deeper than 8 km,

$$ k = k_0 = \frac{(T_{ss} - T_{fs})}{T_{ss}} = \frac{0.04 \text{ sec/0.52 sec}}{0.077}. \quad (5a) $$

The values $T_{ss} = 0.52$ sec and $T_{fs} = 0.48$ sec are obtained from surface-reflected waves, associated with the TD measurement. The shear-wave velocity in the top 0.2 km can be estimated from the measured $T_{fs}$, which gives 0.2 km/0.48 sec = 0.42 km/sec. The velocities in the deeper crust were estimated from surface-wave dispersion analysis by Chung and Yeh (1997). The station CHY is located at the northern boundary of their study area. We combine the shear-wave velocity in the top 0.2 km measured in this study with their shear-wave velocity model (Fig. 13a). The surface-wave dispersion data may not resolve the layer structures accurately. However, it gives a good estimate for the average velocity distribution over the depth range under consideration, which is important for the following analysis.

To estimate the depth distribution of TD data subjected to the boundary conditions (5a) and (5b), we assume two models that relate $k$ to the depth $h$:

$$ k = \begin{cases} k_0 & h < h_c \\ 0 & h \geq h_c \end{cases}, \quad (6a) $$

where $h_c$ is a given depth above which the crustal anisotropy follows a constant $k_0$, and

$$ k = \begin{cases} k_0 (1 - h/8) & h < 8 \text{ km} \\ 0 & h \geq 8 \text{ km} \end{cases}. \quad (6b) $$

We calculate the travel time (beginning at the surface) using the shear-wave velocity model and plot the results in Figure 13a. Then we calculate the TD (again beginning at the sur-
TD is distributed in the top 4 km. Because the actual
in Figure 13b. The results for model (6a) indicate that the
face) based on assumptions (6a) and (6b) and plot the results
k

f

0, this result gives an upper
limit for the TD distribution. Considering a gradual closure
of microcracks with increasing depth (due to increasing of
confining pressure), model (6b) having a linear decaying k
with depth is more realistic than model (6a). The correspond-
ing TD distribution shown in Figure 13b is similarly more
realistic. The TD gradually approaches the observed value
of 0.2 sec when the depth reaches 8 km. Model (6b) is con-
sistent with results of Boness and Zoback (2004) from the
SAFOD pilot hole in Parkfield, California. They found that
the shear-wave velocity anisotropy decreases with depth
from about 10% to 2%. We note that, although the aniso-
tropy may extend beyond 4 km, for both models (6a) and
(6b) 65% to 70% of the TD is distributed within the top 2
km and 20% is in the top 0.2 km. This is not surprising
because the travel time in the top 2 km (Fig. 13a) contributes
nearly half of the total travel time in the 8-km section of the
crust.

Figure 13. (a) A model of S-wave velocity from
0 to 8 km depth and corresponding travel times be-
ing from the surface. (b) Inferred TD beginning
from the surface based on two models of k, two
boundary conditions of k, and the S-velocity model
of a.

Normalization by Travel Distance versus by
Travel Time

Because of the existence of low-velocity layers, the
shallow structure dominates the travel time and it also plays
an important role in the TD distribution. In this regard, we
can see that an inherent drawback exists in the commonly
used method, in which the TD is normalized by travel dis-
tance to describe the degree of anisotropy. For instance, if
our observed total 0.20-sec TD is normalized by 8 km, we
get 25 msec/km. The travel time per kilometer in the deep
layer with a shear-wave velocity of 3.29 km/sec is about 0.30
sec and the corresponding k is 0.083. In contrast, the travel
time per kilometer in the shallow layer with a shear-wave
velocity of 0.90 km/sec is 1.11 sec and the corresponding k
is 0.023. In this case, the consequence of the preceding nor-
malization is that the coefficient of anisotropy k in the deep
layer is 3.6 times the value of k in the shallow layer. Obvi-
ously, this could cause confusion in interpreting measured
results. Based on our shear-wave velocity model, the total
travel time in the top 8 km of crust is 3.7 sec (Fig. 13a).
Thus, the average k = 0.2 sec/3.7 sec = 0.057. In this case,
the corresponding TDs per kilometer over the deep and shal-
low sections are 0.017 and 0.063 sec, respectively, which
are much closer to the real observations. Considering the
strong variations in shear-wave velocity, normalizing the TD
by travel time provides a more physical way of describing
the degree of anisotropy.

Temporal Change of SWS Associated with Stress
Changes Induced by Large Earthquakes

Temporal changes of SWS parameters were reported be-
fore and after the M 6 North Palm Springs earthquake in
southern California (Peacock et al., 1988; Crampin et al.,
1990, 1991). However, Aster et al. (1990, 1991) applied an
automatic technique and similar event analysis of SWS for
the same data set used by Crampin et al. (1990) and did not
identify temporal change in TD. Temporal changes in TDs
before smaller earthquakes were reported by other SWS stud-
ies (e.g., Booth et al., 1990; Liu et al., 1997; Gao et al.,
1998). Crampin et al. (1999) claimed that they successfully
“stress forecast” an M = 5 earthquake in Iceland by using
variations in TDs of the SWS. It is difficult to identify pos-
tible temporal changes from measured TDs with a large scat-
ter. Because analysis of temporal variations often uses TDs
normalized by travel distance, inferred temporal patterns in
TDs could be contaminated by different travel paths, that is,
spatial changes. In many cases it is difficult to measure TDs
unambiguously because of the complex nature of shear-
wave seismograms recorded at the surface. All these factors
contribute to the existing controversy regarding temporal
changes of seismic anisotropy.

As shown in Figure 1, station CHY is located off the
southern end of the Chelungpu fault, on which the M 6.7
Chi-Chi earthquake occurred. The M 6.4 Chiayi earthquake
and another large M 6.0 aftershock occurred just beneath the
station about one month after the Chi-Chi mainshock. Thus,
CHY is well situated to detect possible changes of SWS as-
associated with crustal stress adjustment before, during, and
after the Chi-Chi mainshock and two nearby large after-
shocks. Most of the well-recorded events by CHY are after-
shocks of those large earthquakes, and about 30 events also
occur before the Chi-Chi mainshock. Figures 14a and 14b
show the measured TDs and PDs of the FSW, respectively, in the crust below the 0.2-km-deep borehole station versus time, and Figure 14c shows the calculated TDs in the crust above the 0.2-km-deep station versus time. The circles in Figures 14a and 14c give estimated TD values based on the maximum CCC, and the associated vertical lines mark the 95% confidence interval. The circles in Figure 14b give estimated PD values based on the maximum AR, and the associated vertical lines are error estimates discussed in the Analysis Methods section. The average and standard deviation of the TD and PD values during the 2.7 years before and 2.3 years after the Chi-Chi mainshock are plotted with solid and dashed lines, respectively. None of the results show systematic changes either before or after the mainshock. In addition, the PD values for depth >0.2 km and TD values for the 0–0.2 km depth range do not change significantly over the entire period, while the TD values for depth >0.2 km changed from 0.144 ± 0.028 to 0.161 ± 0.027 sec at the time of the Chi-Chi mainshock. The observed apparent coseismic temporal change of TDs for depth >0.2 km is, however, affected by spatial changes associated with different event locations before and after the Chi-Chi mainshock.

Figure 15 shows the hypocenter locations of events before (triangle) and after (circle) the Chi-Chi mainshock. Because the hypocenter distributions of these two sets of events are quite different, the seismic waveforms generated by earthquakes before and after the mainshock sample, in general, different regions of space. To estimate the possible impact of the different event locations on the TD values, we assign for each event before the Chi-Chi mainshock the closest event after the mainshock. Fifteen pairs of such events with catalog distances less than 2 km are marked in Figure 15 with filled symbols. The mean and standard deviation of the TD values from these 15 pairs are 0.146 ± 0.031 before and 0.155 ± 0.034 after the mainshock. The difference between those values is significantly smaller than that associated with the entire data set (Figure 14a, solid lines). If we decrease the maximum separation distance to 1.5 km, the number of event pairs drops to 11 and the TD values are characterized by 0.146 ± 0.026 before and 0.151 ± 0.038

**Figure 14.** (a) Measured TDs in the crust below 0.2 km versus time. Arrows indicate the times of the Mw 7.6 Chi-Chi mainshock and two M 6.4 and M 6.0 Chiayi aftershocks. (b) Measured PD of the FSW below 0.2 km versus time. (c) Measured TDs in the top 0.2 km versus time. The mean and standard deviations of the results before and after the Chi-Chi mainshock are indicated by solid and dashed lines.
after the Chi-Chi mainshock. The results indicate that spatial variations of parameters affect strongly the apparent coseismic temporal change of TDs for the section deeper than 0.2 km.

The SWS analysis of clusters of earthquake doublets (multiplets) may be the most robust way to identify temporal variation of anisotropy. Doublets and multiplets are sets of earthquakes with similar waveforms, and indications of similar hypocentral locations, focal mechanisms, and ray paths to the stations. Temporal changes extracted from analyzing cluster of earthquake doublets or multiplets should have less contamination due to spatial variations. Figure 16 shows a set of eight multiplets from borehole records at CHY spanning a period from about 1.5 years before the Chi-Chi mainshock (25 February 1998) to seven months after it (13 April 2000). The pertinent parameters of these events and the measured TDs and PDs are given on the left of Figure 16b above the traces of the north–south components. The arrows mark the traces before the mainshock. Figures 16c and 16d display the resolved FSW and SSW of these events in a stacked form. The results from these similar events clearly demonstrate a lack of significant TD and PD changes. The well-synchronized reflected waveforms in Figure 16 also indicate that there are no significant changes in near-surface crustal anisotropy during this period (e.g., a change of TD larger than about 10 msec).

An important result of this study is that there were no systematic changes of SWS parameters observed within the 2.7 years before the Chi-Chi earthquake. The measured lack of temporal changes in anisotropy characterizes the vicinity of several large earthquakes during a highly active tectonic period, which involved a 100-km rupture with as much as 8 m of surface slip and a very energetic aftershock sequence, including many $M \geq 6$ events. The results do not support claims that SWS measurements can provide a general tool for forecasting impending large earthquakes.

Conclusions

We have systematically analyzed SWS of seismic data recorded in the aftershock zone of the 20 September 1999 $M_w$ 7.6 Chi-Chi earthquake at a 0.2-km-deep borehole short-period station and a surface strong-motion station. The analysis uses both the AR and CC methods. The results show that the two methods provide almost the same TDs and similar PDs of the FSW. Because the AR and CC methods use different portions of seismograms, the measured SWS parameters are robust. We have also measured the TDs between the fast and slow shear waves in the top 0.2 km of the crust by analyzing surface-reflected waves in the borehole records.

We found that the incident angle of a ray path might influence the accuracy of SWS measurements, even when the events are within the shear-wave window. The measured PD of the FSW in the crust below the 0.2-km-deep borehole station is $170.2^\circ \pm 4.1^\circ$. This matches very well the surface deformation field during 1992–1999 based on GPS data (Yu et al., 2001). The results indicate that the anisotropy in the section deeper than 0.2 km is controlled by local tectonic stress field. The mean and standard deviation of TD measurements in the top 0.2 km of the crust are $0.04 \pm 0.003$ sec. The analysis reveals strong near-surface (0.2 km) anisotropy in the study region with a coefficient of anisotropy $k = 0.077$. The mean and standard deviation of PD measure-
ments from the surface strong-motion data are $177.5^\circ \pm 4.3^\circ$. The change of PDs between the downhole and surface observations implies that the anisotropic axis of symmetry varies slightly with depth.

The measured TDs in the crust below the 0.2-km-deep borehole station show essentially a constant value of 0.16 sec for sources from 8 to 20 km, implying a zero coefficient of anisotropy in this depth range. We use measured $k$ values (zero below 8 km and 0.077 at the top 0.2 km) as boundary conditions to infer a possible depth distribution of crustal anisotropy based on an $S$-wave velocity model. The assumption of a constant $k$ leads to the result that the anisotropy is limited to the top 4 km of the crust, whereas the assumption of a linearly decaying $k$ with depth gives a slightly deeper extent of anisotropy. Both results indicate that the top 2 km of the crust produces 65% to 70% of the total TD. The dominance of the shallow crust in shear-wave anisotropy is partly due to the existence of low shear-wave velocity layers. Considering the strong variation of shear-wave velocity over depth, we suggest that measured TDs should be normalized by travel time rather than by travel distance.
The observed PDs of the FSW and TDs in both the near-surface and deeper crust, and the great similarity of waveforms generated by multiplets, show no appreciable systematic temporal changes over the 2.7-year period before the 1999 Mw 7.6 Chi-Chi mainshock and the 2.3-year period after. The observed apparent coseismic temporal change of TDs is affected strongly by spatial variations of seismic anisotropy, and at most indicates increasing crack density associated with the mainshock.

We note that our data set is recorded in the vicinity of large rupture zones that experienced stress changes that are as big as expected to occur in the brittle crust during large earthquake cycles. Nevertheless, these stress changes apparently did not leave precursory or postseismic temporal signatures that can be mapped by our SWS analysis of the observed high-quality borehole seismograms. The lack of such temporal changes of SWS parameters and the dominant influence of the shallow structure on the data suggest that the seismic anisotropy in our study region may be associated with a preferred closure of randomly distributed cracks due to the existing anisotropic stress field. A similar cause of seismic anisotropy is suggested by Boness and Zoback (2004).

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