Shear-wave anisotropy: spatial and temporal variations in time delays at Parkfield, Central California

Yun Liu,1,2,* Stuart Crampin1 and Ian Main1

1Department of Geology and Geophysics, University of Edinburgh, West Mains Road, Edinburgh, EH9 3JW, UK. E-mail: scrampin@ed.ac.uk; ian.main@glg.ed.ac.uk

2British Geological Survey, Murchison House, West Mains Road, Edinburgh, EH9 3LA, UK. E-mail: e.liu@bgs.ac.uk

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SUMMARY
Shear-wave splitting is analysed on data recorded by the High Resolution Seismic Network (HRSN) at Parkfield on the San Andreas fault, Central California, during the three-year period 1988–1990. Shear-wave polarizations either side of the fault are generally aligned in directions consistent with the regional horizontal maximum compressive stress, at some 70° to the fault strike, whereas at station MM in the immediate fault zone, shear-wave polarizations are aligned approximately parallel to the fault. Normalized time delays at this station are found to be about twice as large as those in the rock mass either side. This suggests that fluid-filled cracks and fractures within the fault zone are elastically or seismically different from those in the surrounding rocks, and that the alignment of fault-parallel shear-wave polarizations are associated with some fault-specific phenomenon.

Temporal variations in time delays between the two split shear-waves before and after a ML=4 earthquake can be identified at two stations with sufficient data: MM within the fault zone and VC outside the immediate fault zone. Time delays between faster and slower split shear waves increase before the ML=4 earthquake and decrease near the time of the event. The temporal variations are statistically significant at 68 per cent confidence levels. Earthquake doublets and multiplets also show similar temporal variations, consistent with those predicted by anisotropic poroelasticity theory for stress modifications to the microcrack geometry pervading the rock mass. This study is broadly consistent with the behaviour observed before three other earthquakes, suggesting that the build-up of stress before earthquakes may be monitored and interpreted by the analysis of shear-wave splitting.

Key words: anisotropy, fault gouge, pore-fluid pressure, shear-wave splitting, temporal variations.

1 INTRODUCTION
Shear-wave splitting, implying some form of effective elastic anisotropy, has been observed along almost all shear-wave ray paths in igneous, metamorphic and sedimentary rocks in a wide variety of tectonic regimes in many parts of the world (reviewed by Crampin 1994). The phenomenon of shear-wave splitting in the Earth's uppermost crust has been interpreted as the effect of distributions of stress-aligned fluid-filled microcracks and preferentially oriented pore space, known as extensive-dilatancy anisotropy or EDA (Crampin, Evans & Atkinson 1984).

* Now at: Cairn Energy plc, 61 Dublin Street, Edinburgh, EH3 6NL, UK.

It was anticipated that the geometry of EDA cracks would be modified by changes in stress before earthquakes, and temporal variations in shear-wave splitting were first positively identified at 99 per cent confidence levels before and after the M=6 North Palm Springs earthquake, Southern California (Peacock et al. 1988; Crampin et al. 1990, 1991). Temporal changes in time delays before smaller earthquakes in isolated swarms (Crampin 1991) have also been identified (Booth et al. 1990; Gao et al. 1997). Crampin & Booth (1989) and Meadows & Winterstein (1994) also identified temporal changes in shear-wave polarizations induced by hydraulic pumping.

Note that Aster, Shearer & Berger (1990, 1991) used an automatic technique to measure shear-wave splitting on the same data set used by Crampin et al. (1990) but could not identify temporal changes. To our knowledge, no automatic
technique has yet been devised that can reliably measure shear-wave splitting above local earthquakes, and Crampin et al. (1991) were able to show that the method of Aster et al. (1990) introduced errors in time delays of over 200 per cent. A further reason for the difference in results is that Aster et al. (1990) did not take account of the azimuthal variations in shear-wave splitting, which are an essential component of the analysis of Crampin et al. (1990).

A new theoretical model of how fluid-saturated porous rock responds to stress has been developed by Zatsepin & Crampin (1995, 1997) and Crampin & Zatsepin (1995, 1997a). This model shows that rock necessarily responds to small changes in stress, pore-fluid pressure and other phenomena by modifying the microscale geometry of the intergranular microcracks and intergranular pores known to be present in almost all rocks (Zatsepin & Crampin 1995, 1997). In this model, the evolution of fluid-saturated rocks under differential stress is controlled by microscale fluid migration along pressure gradients between neighbouring EDA cracks at different orientations to the stress field. This mechanism leads to an anisotropic poroelasticity (APE) model for the effects of stress on fluid-saturated cracked rock (Zatsepin & Crampin 1995, 1997; Crampin & Zatsepin 1997a). Numerical modelling predicts both the observed polarizations (an orthorhombic perturbation of hexagonal symmetry) and the magnitude (1.5-4.5 per cent) of shear-wave velocity anisotropy observed in intact rocks in the crust (Crampin 1994). Consequently, it is expected that variations in the stress field will modify the microscale geometry of intergranular microcracks and pores. If these effects can be observed in sufficient resolution by an appropriate controlled-source experiment, such as is suggested by Crampin & Zatsepin (1997b), then in principle the evolution of stress can be inferred from the observed temporal changes in time delays between split shear waves. Crampin & Zatsepin (1997a) have shown that APE modelling quantitatively matches, within the error bounds, the temporal variation of time delays associated with the $M=6$ North Palm Springs earthquake (Crampin et al. 1990, 1991) for a simple linear increase before and an abrupt decrease at the time of the earthquake. By implication, APE could also model behaviour at the other earthquakes where temporal variations have been noted (Booth et al. 1990; Gao et al. 1997).

In this paper, we present an analysis and interpretation of three years of data from the Parkfield seismic network across the San Andreas Fault, Southern California, displaying shear-wave splitting and temporal changes in the time delays between split shear waves. Liu et al. (1993) and Liu (1995) presented preliminary reports on this data set.

2 PARKFIELD HIGH-RESOLUTION SEISMIC NETWORK (HRSN)

The Parkfield High-Resolution Seismic Network (HRSN) was designed to monitor an expected $M \approx 6$ earthquake near Middle Mountain (MM) in the earthquake preparation zone near Parkfield on the San Andreas Fault (SAF), Central California (Bakun & McEvilly 1984; Bakun & Lindh 1985). The anticipated earthquake has not yet occurred. Nine sites were instrumented with three-component seismometers installed both in shallow boreholes at the surface and in deeper boreholes at 200-300 m depth. Locations are shown in Fig. 1 and are listed with other details in Malin et al. (1989). All instruments are 2 Hz Mark Products L22E seismometers recording at 500 samples per second. The triaxially mounted three-component seismometer packages are locked in pressure-sealed packages and oriented by levelling gimbals in wire-mesh cages (Malin et al. 1989). Triggered events were recorded with General Earthquake Observation System (GEOS) event recorders on loan from the United States Geological Survey (USGS), Menlo Park (Borcherdt et al. 1985).

Note that the two-letter station codes used in this paper have associated international three-letter codes: ED, FR, GP, JN, JS, MM, ST, VC and VR are equivalent to EAD, FRO, RMN, JCN, JCS, MMN, SMN, VCA and VAR, respectively.

3 LOCAL GEOLOGY

The geology of the Parkfield area is dominated by the strike-slip plate boundary of the SAF. Fig. 1 shows a map of the area with associated faults and the distribution of the Parkfield HRSN. The fault segment at Parkfield is generally understood to be the transition zone between the 170 km long creeping portion to the northwest and the 300 km long locked portion to the southeast. The average width of the zone of fault gouge varies from 100 to 200 m and it is bounded on both sides by a $\approx 400$ m transition zone (Leary & Ben-Zion 1992; Eberhart-Philips & Michael 1993). The SAF at the Parkfield segment is characterized by an abrupt cross-fault velocity gradient, with a 5 to 20 per cent lateral change in velocity in a 4 km wide zone parallel to the fault trace (Michelini & McEvilly 1991; Eberhart-Philips & Michael 1993). Analysis of seismograms recorded at Middle Mountain (MM) in terms of fault-zone trapped waves has shown that the shear-wave velocity of the fault gouge is about 1.1 km s$^{-1}$, and about 1.8 km s$^{-1}$ in the transition zone in the area close to the epicentre of the 1966 $M_L = 5.9$ main shock, whereas in the southeast segment approaching the locked portion of the SAF the shear-wave velocity is 1.8 km s$^{-1}$ in the gouge, and 2.5 km s$^{-1}$ in the transition zone (Li & Leary 1990; Leary & Ben-Zion 1992). The fault is dipping steeply at $86^\circ \pm 1.1^\circ$ to the southwest (Brown et al. 1967; Nishioka & Michael 1990).
The Salinian block on the southwest side of the SAF consists of Gabilan plutonic and metamorphic granite and granodiorite basement overlain by a maximum of 2 km of marine and non-marine sedimentary and volcanic rocks of Tertiary age with Quaternary deposits (Brown et al. 1967; Lees & Malin 1990). These stratified rocks are tightly folded along northwest-trending axes near the fault zone, but at approximately 1 km from the fault zone the folds are open and generally dip away from the Cholame Hills high, where Station VC is located. The Cholame Hills, corresponding to the basement uplift, parallel the SAF system several kilometres to the southwest, and are a high-velocity 'slice' on the southwest block of the SAF (Eberhart-Philips & Michael 1993). The northeast block of the SAF consists broadly of outcropping basement of Franciscan melange. Overlying this basement to the east are several kilometres of Cretaceous and younger sediments of the Great Valley sequence. This block is much more deformed than the Salinian block, and is complicated by numerous folds, thrust faults and strike-slip tear faults, demonstrating much internal deformation during its evolution. In general, Franciscan rocks are moderately to strongly deformed and, in the fault zone, slivers of various crystalline rocks appear to have been trapped within older branches of the fault (Brown et al. 1967). Some of the numerous thrust faults surrounding the SAF show Holocene movement.

4 SHEAR-WAVE SPLITTING

4.1 Shear-wave polarizations

Fig. 2 displays typical examples of shear-wave splitting at five stations of the HRSN network. The horizontal components of each seismogram have been rotated into faster and slower directions parallel and orthogonal to the polarization of the leading split shear waves. Note that the faster and slower horizontal components may or may not display similar waveforms, as the polarizations respond differently to the structure between source and receiver. For example, the slower arrival at VC in Fig. 2 has considerably higher frequency than the faster arrival. Consequently, although the arrivals and time delays may be relatively easily determined by visual inspection, automatic techniques face severe difficulties in routinely analysing shear-wave splitting. Note also that the vertical waveforms are typically very different from the horizontal waveforms. These are both characteristic features of shear-wave splitting above small earthquakes.

The distribution polarizations of the faster and slower split shear waves within the effective shear-wave window (Booth & Crampin 1985) at each station are plotted in Fig. 3, where (a) is a map of equal-area rose diagrams, and (b) shows equal-area polar projections. All triggered and located events were examined. Inverse weights 1–3 have been linearly applied to each measurement, where weight 1 (long bar) refers to the most reliable estimates and weight 3 (short bar) to the least reliable. The polarizations of the faster split shear waves display the approximately parallel alignments at ED, JN, MM, VC and VR that are typically observed above small earthquakes [see reviews by Crampin & Lovell (1991) and Crampin (1994)]. These polarizations are distributed about a N10°E direction, which is approximately parallel to the direction of the maximum principal stress (arrow in Fig. 3) near the SAF in central California (Zoback et al. 1987). The parallel polarizations at MM, however, are subparallel to N40°W, approximately parallel to the SAF and about 70° from the direction of maximum horizontal stress.

Of the five stations with most data, FR and ST show severe

![Figure 2](https://academic.oup.com/gji/article/130/3/771/677732/10)
4.2 Time delays between split shear waves

Temporal variations could be examined only at stations MM and VC, which recorded undisturbed shear-wave splitting from a sufficient number of small earthquakes within the shear-wave window. The greatest time delay between the faster and slower waves is about 80 ms at both MM and VC, corresponding to about 3 and 2 per cent differential shear-wave-velocity anisotropy, respectively. (Levels of normalized time delays at MM are abnormally high, probably owing to high pore-fluid pressures—see discussion in Section 6.2.) Figs 4(a) and (b) show the variation of time delays with hypocentral distance. If the anisotropy were uniformly distributed, the time delay in a given direction would be proportional to the hypocentral distance. The approximately linear variation of the maximum, and the median line passing close to the origin justify normalizing the time delays to unit path length in ms km$^{-1}$.

Note that time delays between split shear waves are expected to vary with angle of incidence and azimuth in 3-D patterns depending on the anisotropic symmetry. In particular, time delays will vary typically from the maximum value to zero within the effective shear-wave window (angles of incidence less than about 45$^\circ$). Consequently, wide variations in time delay are expected, and scatter similar to that in Fig. 4 is usually observed. Such scatter cannot be resolved without a much more accurate knowledge of the ray path and velocity structure than is generally available with earthquake data.

4.3 Depth extent of anisotropy

Figures 4(c) and (d) show the variation with focal depth of normalized time delays at MM and VC, representing the amount of seismic anisotropy. The delays at VC, on the block southwest of the SAF, are uniformly scattered with no clear variation with depth, suggesting that relatively uniform anisotropy extends from the surface to at least 14 km depth. At MM, however, there is a pronounced decrease, by a factor of two, in normalized time delays at focal depths greater than about 8 km. MM is close to the surface break of the SAF, and appears to be on the transition zone between the fault gouge and the surrounding rocks, so that the ray paths of shear waves to MM are likely to be in fault gouge (Li & Leary 1990; Leary & Ben-Zion 1992). Consequently, shear-wave splitting at MM may be attributed principally to the internal structure of the fault gouge, and hence the termination of gouge beneath MM may be indicated by the pronounced decrease at 8 km. This is in broad agreement with the results of Malin et al. (1989) who suggested that the fault gouge may extend from the surface to 10 km depth at the locked portion towards the southeast of the Parkfield segment and to about 5 km depth at the creeping portion towards the northwest of the segment in the fault zone. The range of time delays, between 2 and 6 ms km$^{-1}$, at VC is similar to that seen almost universally below $\approx$ 1 km elsewhere (Crampin 1994), whereas the delays at MM are about twice as large, and are probably associated with the internal structure of the fault gouge.

5 TEMPORAL CHANGES IN TIME DELAYS

5.1 Shear-wave splitting

In general, local earthquakes recorded by the HRSN network occur close to the fault plane with depths between 4 and 15 km, and magnitudes in the range $-0.5 \leq M_L \leq 2.0$. The largest earthquake during the period studied was a $M_L=4$ event recorded on 25th May 1989 in the Parkfield segment. The epicentre is marked by a star in Fig. 3. The event was approximately those observed elsewhere. Those at MM are abnormally high, probably owing to high pore-fluid pressures—see discussion in Section 6.2. (Figures 4(a) and (b) indicate the relative reliability of the measurements.)
Figure 4. Variation with hypocentral distance of time delays at (a) MM and (b) VC. Variation with focal depth of time delay normalized to ms km^{-1} at (c) MM and (d) VC.

Note also that we have examined other phenomena that might have caused variations in time delays. There appears to be no obvious correlation with hypocentral depth (which would have a direct effect on the normalized time delays), epicentral position, earthquake magnitude, focal mechanism or signal frequencies. We conclude that the observed changes are caused by changes in the response of shear-wave splitting along similar ray paths.

5.2 Statistical significance of observed temporal changes in time delays

The statistical significance of the observed changes in time delays may be analysed with respect to estimates of the experimental error. We therefore make an error analysis on data with weight 1 showing the most reliable measurements of shear-wave splitting. Errors are introduced in two principal ways. The faster and slower split-shear-wave arrivals are both superimposed on the P-wave coda and other signal-generated noise, and the energy of the arrivals may not be large enough to produce an obvious change of motion and so may be misidentified. The second source of error is the uncertainty in the correct start time of second split shear waves because the second arrival may be marked by elliptical motion when the time delay is not large enough to separate the faster and slower split shear waves completely. Under these circumstances, a measurement of observations of time delays and its

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The corresponding error may be estimated as follows:

\[ N = 0.5 \left( (N_{\text{max}} + N_{\text{min}}) \pm (N_{\text{max}} - N_{\text{min}}) \right), \]  

where \( N_{\text{max}} \) is the number of observations of possible maximum time delays and \( N_{\text{min}} \) is the number of observations of possible minimum time delays.

The temporal change in time delays at MM and VC is thus significant at the 68 per cent confidence level: strictly, the null hypothesis that no anomaly exists can be rejected at one standard deviation. This level of confidence is not high and has, statistically, a one in three chance of being wrong. Additionally, if the behaviour before the \( M_L = 4 \) earthquake is significant then the increase in time delays towards the end of the three-year period, particularly evident at MM in both Fig. 5 and Fig. 6, must also be considered as significant, and there was no larger earthquake at this time. The overall significance of these results will be considered in the discussion in Section 7.

### 5.3 Earthquake doublets/multiplets

Observations of time delays above earthquakes are typically based on the analysis of large sets of seismic events. Each event has different ray paths so that the data may be contaminated by random spatial variations which could be interpreted as temporal changes (Gledhill 1993; Kaneshima 1990). This range of ray paths means that the exact effects of temporal changes are difficult to identify. Analysis of clusters of earthquake doublets (multiplets) is one method of avoiding such artefacts due to spatial variations, and may be able to provide direct evidence of temporal variation of time delays along specific ray paths.

Doublets (or multiplets) are pairs (sets) of earthquakes with similar waveforms, and by implication, similar hypocentral locations, focal mechanisms, and similar ray paths to seismic stations. Detailed surveys show that multiplets are extremely common and that most, if not all, earthquakes are in clusters of multiplets (Lovell et al. 1987; Nadeau 1995). Poupinet,
Ellsworth & Frechet (1984) and Aster et al. (1990) used earthquake doublets to measure seismic-velocity changes. Got & Frechet (1993) used seismic doublets to measure temporal variations of attenuation in the crust before large-magnitude earthquakes. More recently, Nadeau et al. (1994) and Nadeau (1995) identified 294 clusters of from two to 12 multiplets during the period 1987–1992 at the Parkfield segment. In this study, we use earthquake multiplets to investigate the temporal changes in time delays observed at Parkfield.

In selecting clusters of multiplets, we examined earthquakes within the shear-wave window at MM and VC with a difference in depth of less than 2.5 km, and differences in both incidence and azimuth to MM and VC of less than 10°. For the chosen similar events, cross-correlation of each three-component trace is calculated for a 0.24 and a 0.3 s window length at MM and VC, respectively, starting from the beginning of the shear-wave arrival. Four distinct clusters of up to five-event multiplets were eventually selected, with an average maximum correlation coefficient greater than 0.63 at MM and 0.65 at VC. Table 1 lists the multiplets in each cluster.

We chose a central event in each cluster in relation to its geographical location, marked by asterisks in Table 1, and calculated correlation coefficients with respect to these events. Clusters of multiplets shared by both stations have been examined, but none was eventually used because they are either located outside the effective shear-wave window (angles of incidence less than 45°), or show complicated or poor-quality shear-wave trains with no satisfactory measurements of time delay at one of the stations. The similarity of the source mechanisms of the events in each cluster family was examined by comparing waveforms with all other stations in the network. Fig. 7 displays the three-component seismograms of the five selected clusters, where the horizontal components have been rotated to faster and slower directions, and arrows mark the arrival times of the split shear waves. The earthquakes in each cluster family have nearly identical P and S waveforms with average maximum correlation coefficients for the shear-wave train of 0.78 (Table 1).

Fig. 8(a) shows the locations of the cluster events, and Fig. 8(b) shows the variation with time of the normalized time delays of the multiplets of each cluster at MM and VC. The interevent distance varies from a few tens of metres to hundreds of metres. The arrow indicates the time of the $M_L = 4$ earthquake. In view of the known scatter in observations of
Figure 7. The three-component seismograms at MM and VC from each of the five selected clusters of multiplets, A to E. Notation as in Fig. 2.
shear-wave splitting (see Discussion), not too much emphasis should be placed on this small selection of data; however, the behaviour is consistent with previous figures. Cluster A at MM shows an increase before and a decrease after the earthquake, and then one point indicating a further significant increase (as is also seen in Figs 5 and 6). At VC, we can only see the increase of delays before the event, since there is only one cluster of four events selected and they all occurred before the $M_L=4$ earthquake. Combining the two stations we see a change of time delays which is associated with the time of the $M_L=4$ event.

Note that we identify multiplets with temporal variations in split-shear-wave delay times. Zatsepin & Crampin (1995, 1997) and Crampin & Zatsepin (1995, 1997a) have shown that as
fluid-saturated EDA cracks evolve (in changing 3-D distributions of aspect ratios), thus the effects on shear waves will be anisotropic. In particular, shear waves in some directions may show changes, whereas shear waves (and P waves) along other ray paths may not. This means that the stability of multiplets for particular source-to-station ray paths does not necessarily imply overall stability of the rock mass, and temporal changes may be visible in other directions.

5.4 Summary of temporal changes
We have applied three independent tests to show temporal variations in shear-wave splitting at different levels of resolution:

1. we have used the whole data set of time delays irrespective of data quality (Fig. 5);
2. we have used a subset of high-quality (weight 1) data and independently estimated error bars (Fig. 6);
3. we have used a small subset of multiplet events (Fig. 8).

The results of each test suggest an increase in time delays between split shear waves before the $M_\text{w}=4$ earthquake and a (possibly abrupt) decrease near the origin time of the earthquake. Note, however, that Tests 2 and 3 show particularly high levels of time delays at MM towards the end of the three-year period, 1988–1990, which were not followed by an identifiably large earthquake.

6 INTERPRETATION

6.1 Shear-wave polarizations
Shear-wave first motions are aligned in the directions N10°E±20° at VC, ED, JN and VR in Fig. 2. Such scatter of directions is characteristic of shear-wave splitting above small earthquakes (Crampin & Booth 1985; Crampin 1994). This direction is subparallel to the direction of maximum horizontal principal stress near the SAF in Central California (Zoback et al. 1987), marked by an arrow in Fig. 3. VC is about 5 km away from the SAF, on the southwest block. The structure on the northeastern side of the SAF is much more complex (Brown et al. 1967; Eberhart-Philips & Michael 1993), and the essentially common pattern of nearly parallel polarizations either side of the fault clearly demonstrates that the cause of the parallel polarizations is independent of local geological conditions (Liu 1995).

In an isotropic medium, shear-wave polarizations recorded within the shear-wave window at a seismic station on the surface would depend on the polarization radiated from the source mechanism of the earthquake, modified only by interaction with local topography and internal interfaces (Booth & Crampin 1985; Liu & Crampin 1990). Shear-wave polarizations in the anisotropic crust typically exhibit parallel or subparallel alignment with the direction of maximum horizontal compressive stress (Crampin 1994; Crampin & Lovell 1991). A number of studies (Crampin et al. 1986; Gledhill 1990, 1991; Liu 1995

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on this data set) have compared the differences between observed polarizations and those expected from earthquake focal mechanisms. They find that in general shear-wave polarizations are consistent with the regional maximum horizontal compressive stress and can be interpreted as the vector decomposition of the radiated polarizations.

These nearly parallel polarizations at surface recorders are crucial elements in determining the cause of shear-wave splitting. The only anisotropic symmetry system that can yield parallel polarizations within the shear-wave window at the surface is hexagonal symmetry with a horizontal axis of symmetry, or a minor perturbation of such symmetry (Crampin 1981). Hexagonal symmetry, particularly aligned with a horizontal axis of symmetry, is rare amongst minerals or lithologies in the crust. The only common source of such symmetry is parallel vertical cracks (EDA cracks), which are typically aligned, like hydraulic fractures, perpendicular to the direction of minimum stress. Below a few hundred metres this minimum stress is typically horizontal, yielding the observed shear-wave polarizations (Crampin 1994). Zatsepin & Crampin (1995, 1997) and Crampin & Zatsepin (1995, 1997a), using APE modelling, showed that the evolution of intergranular fluid-filled microcracks and pores under stress conditions appropriate to the upper half of the crust necessarily leads to orthorhombic distributions of EDA cracks, consistent with the data presented here. In many circumstances, these orthorhombic symmetries are minor perturbations of hexagonal symmetry, and yield the observed parallel shear-wave polarizations within the shear-wave window. The polarizations observed here, subparallel to the direction of compressive stress at VC, ED, JN and VR, are consistent with stress directions near the SAF, and with APE modelling.

Shear-wave first motions at MM, however, show pronounced alignment in the direction N40°W, approximately parallel to the SAF about 70° from the stress direction. Such fault-parallel polarizations have previously been observed in California at seismic Station KNW on the San Jacinto Fault, California (Crampin et al. 1990, 1991), and at seismic stations in the Loma Prieta segment of the SAF (Zhang & Schwartz 1994). Station MM is sited in the immediate vicinity of the SAF, and 95 per cent of the shear waves from the earthquakes recorded within the shear-wave window (angle of incidence less than 45°) at this station propagate along ray paths at an angle of less than 15° to the proposed vertical EDA crack plane. The average width of the fault gouge of the SAF in the Parkfield segment varies from 200 to 400 m, with about 400 m of transition zone on both sides of the fault (Li & Leary 1990; Leary & Ben-Zion 1992). It is thus likely that the stress-oriented gouge in the fault zone plays the most significant role in controlling shear-wave polarizations at MM.

The polarizations at MM are at 60° to those at VC. They are also approximately orthogonal (within 20°) to the direction of compressive stress, which would tend to close cracks normal to the stress. Consequently, the approximately orthogonal polarizations at MM are anomalous. Crampin et al. (1996) have shown that APE modelling implies that in some conditions high pore-fluid pressures lead to shear-wave polarizations within the shear-wave window becoming orthogonal to the strike of EDA cracks. This is supported by observations in a high-pressure reservoir in the Caucasus Oil Field (Crampin et al. 1996). High pore-fluid pressures are very common in the Earth (Neuzil 1995). Consequently, since high pore-fluid pressures are probably associated with heat-flow anomalies along the SAF (Rice 1990) and are likely to be necessary to relieve high frictional forces on most earthquake fault planes, it is probable that the anomalous polarizations at MM are due to high pore-fluid pressures within the fault gouge, rather than having to appeal to cracks nearly perpendicular to the stress field. Since some of the field evidence for fault-parallel cracks is based on shear-wave data, fast shear-wave propagation with polarizations perpendicular to the crack face, caused by high pore-fluid pressures, could affect other results indicating fault-parallel cracks.

### 6.2 Temporal changes in time delays

Liu (1995) has shown that, as has been found elsewhere, this data set showed no correlation of temporal changes in time delays with seismic moment or with focal depth (apart from the reduction in time delays below 8 km at MM attributable to the bottom of the fault gouge, see Section 4.3 above). The most plausible explanation for temporal changes in shear-wave splitting is stress-induced changes to EDA crack geometry (Peacock et al. 1988; Crampin et al. 1990, 1991; Booth et al. 1990; Gao et al. 1997). Zatsepin & Crampin (1995, 1997) showed that the evolution of fluid-filled intergranular EDA cracks leads to the observed behaviour of shear-wave splitting in the crust. Crampin & Zatsepin (1997a) used APE modelling to match observations of temporal changes at Anza (Crampin et al. 1990, 1991). The effects are due to increasing crack aspect ratios of cracks distributed about the strike of the stress as the maximum stress increased before the earthquake.

The overall polarizations of the shear waves, and overall alignment of EDA cracks, in Fig. 3 do not appear to change with time. Theoretical studies of crack distributions (Peacock et al. 1988; Crampin et al. 1990) suggest that changes of aspect ratio of parallel vertical distributions of cracks will marginally vary the normalized time delays in the band of wave paths between 14.5° and about 45° to the average EDA crack plane within the shear-wave window. Whereas changes of crack density in such geometries, caused either by changes in the number of cracks or by changes in the size (radius) of cracks, will effectively increase the normalized time delays between faster and slower split shear waves over the whole of the shear-wave window, including the band of wave paths at less than 14.5° to the crack plane that are not affected by changes in aspect ratio. This means that the overall pattern of changes in time delays can, in principle, distinguish the character of the changes in crack geometry. The interpretation at Anza was that the behaviour of shear waves could be explained by the combined effects of the following: a small increase in crack density and a more marked increase of aspect ratio (bowing) of the supercritical fluid-filled parallel vertical EDA cracks throughout the rock mass as the stress built up before the North Palm Springs earthquake (Crampin et al. 1990); a rapid decrease of aspect ratio (flattening) at the time of the earthquake; and a more gradual reduction of crack density by partial healing as stress relaxed following the earthquake. Note that the optimum division between the two bands across the shear-wave window may vary with the data set, that is with the history of the rock mass. At Anza, 14.5° was appropriate (Peacock et al. 1988), whereas 13° seems to be more appropriate for the Parkfield data set.

APE modelling suggests that wholly parallel vertical systems
EDA cracks are only a first approximation of the behaviour of fluid-saturated rocks under stress (Zatsepin & Crampin 1995, 1997; Crampin & Zatsepin 1995, 1997a). However, the interpretation of Crampin et al. (1990, 1991) appears to be valid for the isolated swarms at Enola (Booth et al. 1990) and Hainan Island (Gao et al. 1997). In order to assess whether the temporal variations at Parkfield can also be interpreted by simple changes in crack density or changes in aspect ratio, or possibly a combination of both, ray paths within the shear-wave window were split into two bands, 13° to 45°, and ±13° to the crack plane. Since some 95% of the data received at MM are from waves propagating within ±13° to the crack plane and thus largely independent of changes in aspect ratio, this examination was only relevant to data recorded at VC.

Fig. 9 shows equal-area projections of time delays and the variations of time delays against time in the two bands of data in the shear-wave window at VC. Solid lines are five-point moving averages. [Note again that moving averages may not be wholly appropriate as they tend to smooth over the apparently abrupt change at the time of the earthquake.] Both bands in Fig. 9 show a large scatter, but both show an increase before the $M_L=4$ event, a decrease just after the event, and a further increase thereafter. Since changes of crack density modify time delays in the whole of the shear-wave window, the marked increase in crack density implied by the central band, ±13° to the crack plane, in Fig. 9(a) masks any possible change in aspect ratio. (Note that the temporal change at MM in Fig. 5 is wholly within this central band.) This suggests that crack density increased beneath VC before the earthquake, dropped at the time of the earthquake, and that cracks started to heal once the stress had relaxed following the earthquake.

Existing data sets are too few to generalize, but the behaviour at VC appears to be atypical. The previous examples (Crampin et al. 1990, 1991; Booth et al. 1990; Gao et al. 1997) show time delays in the central band ($\pm 14.5^\circ$ to the crack plane) varying only marginally, so that the increases in the outer band (14.5° to 45° to the crack plane) suggest that an increase in aspect ratios is the dominant effect of increasing stress on EDA crack geometry before earthquakes. Note that APE modelling (Zatsepin & Crampin 1995, 1997) shows that the behaviour of shear waves sampling the evolution of EDA crack geometry (principally modifying the distribution of crack aspect ratios) depends on the stress history of the rock mass. Currently, there are too few observations of stress-induced changes in shear-wave splitting to indicate whether each particular seismic zone needs to be characterized differently.

It is worth noting that in both bands in Fig. 9 (including the outer band thought to be largely due to changes in crack density), the average halving of time delays at the time of the earthquake is typical of that seen below 1 km in temporal changes elsewhere (Crampin 1994; Crampin & Zatsepin 1997a), where it has been interpreted as being related to changes in aspect ratio (Crampin et al. 1990, 1991; Booth et al. 1990; Gao et al. 1997). It may also be relevant that the change in time delays at MM in Fig. 5 is a reduction of not a half, but about a quarter (25 per cent) of the maximum time delay. This is the

**Figure 9.** Variation of normalized time delays within the shear-wave window at VC. Equal-area polar projections out to 45° of time delays (top), and variation with time of time delays (bottom). (a) Plots of time delays in the band at $\pm 13^\circ$ to the average crack strike; and (b) time delays in the band at 13° to 45° to the crack strike. The solid line is a five-point moving average, and the arrow marks the time of the $M_L=4$ event.
only reported example to date of temporal changes wholly within the fault zone, where fluids are likely to be at high pore-fluid pressures (Rice 1990; see previous section). It might be expected that such high pore-fluid pressures near the fault zone would accelerate stress-corrosion cracking (Atkinson 1984), which would modify crack density during stress increases before the impending earthquake by changing both the number and size distribution of cracks. Fluid–rock interactions are critical to many deformation processes, and the development of the pre-existing cracks by stress corrosion, dissolution and diffusion in the fault zone would be influenced by the chemical effects of high-pressure pore fluids in the crustal environment. Consequently, the propagation rate of the cracks in the fault zone with high-pressure pore fluids is likely to be greater than that in the more intact rock beneath VC. This is consistent with the differential shear-wave anisotropy at MM being about twice as large as that at VC and elsewhere (Crampin 1994). This suggests that the fluid-filled fractures within the fault zone are more extensive and will modify more readily than that in the surrounding crust.

An alternative explanation for the larger time delays at MM than at VC is that, since MM is sited on fault gouge, lower seismic velocities would increase the time delay for given crack configurations and a given percentage of shear-wave anisotropy. However, in a review of shear-wave splitting, Crampin (1994) found comparable ranges of percentage shear-wave anisotropy in sedimentary, igneous and metamorphic rocks, regardless of seismic velocities. This suggests there may be some compensatory phenomenon, such as rock strength, that acts to equilibrate the effects of cracks in different rock types.

7 DISCUSSION

APE modelling suggests that fluid-saturated microcracked rocks, that is virtually all rocks, are highly compliant and respond to small changes in stress by redistributing the pore fluid along pressure gradients between neighbouring intergranular EDA cracks at different orientations to the stress field. Shear waves are very sensitive to such changes in EDA crack geometry, and temporal changes in shear-wave splitting as the stress builds up before earthquakes must be expected and can be modelled (Crampin & Zatsepin 1997a,b). Crampin et al. (1990, 1991) showed changes before and after a M = 6 earthquake some 30 km from the seismic station, and Booth et al. (1990) and Gao et al. (1997) showed changes in shear-wave splitting at seismic stations within the shear-wave window of isolated swarms of earthquakes which are very quiet environments for monitoring small changes in behaviour (Crampin 1991).

As yet there are comparatively few records of temporal changes before earthquakes. This is due to the stringent recording requirements. What are needed are the following:

1. swarms of small earthquakes to provide a source of shear waves;
2. a recording network of three-component seismometers at high enough digital sampling rates to record comparatively high-frequency shear waves within the shear-wave window of the earthquake swarm; and
3. a large earthquake nearby.

Alternatively, if an isolated swarm is monitored, requirement (3) is relaxed, and temporal changes appear to be visible before the larger earthquakes of the sequence, known as typical events (Crampin 1991), which may still be comparatively small. However, isolated swarms are comparatively scarce, and there may be differences between the behaviour before typical events and that before large earthquakes.

The temporal variations in shear-wave splitting at Parkfield reported in this paper (a M1 = 4 earthquake at $\approx 15$ km depth) are an intermediate case, where the earthquake is not very large and the seismicity is part of continuous seismic activity along many tens of kilometres of the San Andreas Fault and is not an isolated swarm. APE modelling suggests that a characteristic of the temporal changes in shear-wave splitting before earthquakes is that they are very sensitive. Chen, Booth & Crampin (1987) examined temporal changes in shear-wave splitting at the Turkish Dilatancy Projects, which monitored a seismic zone on the North Anatolian Fault at least 15 km in diameter. Chen et al. (1987) observed changes in shear-wave splitting, but were unable to associate the changes with specific earthquakes. The changes could have been associated with comparatively small earthquakes nearby, or larger more distant earthquakes. Although the statistical significance of the Parkfield data set presented here is low (68 per cent), it does display behaviour broadly consistent with (the small number of) observations elsewhere. To our knowledge, there are no reports of observations of small earthquakes within the shear-wave window before a large earthquake where temporal changes in shear-wave splitting have not been identified.

Such indeterminacy is always likely to be present in monitoring programmes where the shear-wave source is small earthquakes. An extended zone of seismicity will require a large earthquake to organize the geometry of EDA cracks sufficiently to show temporal changes in shear-wave splitting and provide the resolution necessary to identify the build-up of stress reliably. Crampin & Zatsepin (1997b) suggest that the build-up of stress before earthquakes can be monitored by controlled-source experiments using an airgun source in cross-hole measurements between deviated boreholes.

8 CONCLUSIONS

Analysis of shear-wave splitting in a three-year data set from the Parkfield HRSN Network has shown polarizations and time delays broadly consistent with observations above small earthquakes elsewhere. These observations include parallel or subparallel polarizations at each station, where, with one exception, station averages are subparallel to the direction of the regional compressive stress. The exception, Station MM, is situated on the fault-zone gouge, where the polarizations are fault-parallel and there are exceptionally high levels of normalized time delays. These are probably associated with high pore-fluid pressures in the fault gouge.

Time delays between split shear waves can be normalized to unit path length. At the two stations with most data, MM and VC, the normalized time delays show an increase before, and an abrupt decrease near the time of a $M1 = 4$ earthquake some 15 km from MM and VC. This is followed by the suggestion of a further increase in time delays. The initial change is statistically significant at one standard deviation (68 per cent confidence), and can be interpreted in terms of increasing crack density before the earthquake and an abrupt decrease near the time of the earthquake as the stress relaxed. This interpretation differs from the other three examples of

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temporal changes in shear-wave splitting before earthquakes, where the behaviour can be interpreted as being dominated by the changing aspect ratio of distributions of parallel vertical EDA cracks. These analyses are consistent with the predictions of APE modelling that stress-induced modifications of highly compliant fluid-filled intergranular EDA cracks are sensitive to marginal changes in stress distributions before earthquakes, and can be monitored with shear-wave splitting.

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REFERENCES


© 1997 RAS, GJI 130, 771–785
Temporal variations in splitting


Nadeau, R., 1995. Characterization and application of microearthquake clusters to problems of scaling, fault zone dynamics and seismic monitoring at Parkfield, California, *PhD Dissertation*, University of California, Berkeley, CA, USA.


