

Review

A review of techniques for measuring shear-wave splitting above small earthquakes

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Received 9 November 2005; received in revised form 22 May 2006; accepted 15 June 2006

Abstract

Seismic shear-wave splitting is difficult to measure accurately because of the complexity of the shear-wave signal. A variety of techniques have been developed for measuring time-delays and polarisations of shear-wave splitting above small earthquakes. These range from ‘display’ techniques, where measurements depend on visual examination of rotated seismograms and polarisation diagrams, through a range of increasingly automatic techniques, to what are almost fully automatic processes. All techniques have disadvantages. Visual techniques are subjective and, although arguably the most accurate, are tedious and time-consuming. More automated techniques work well on noise-free impulsive near-classic examples of shear-wave splitting, but on typical records either require visual checking or need to pass stringent selection criteria which may severely limit the data and bias the results. The accompanying paper presents a combination of visual and automatic techniques to provide a user-friendly semi-automatic measurement technique. Such techniques are important because the new understanding of fluid-rock deformation suggests that shear-wave splitting monitors the low-level deformation of fluid-saturated microcracks in hydrocarbon production processes, as well as the accumulation of stress before earthquakes, and other applications.

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Keywords: Automatic techniques; Display techniques; Review of techniques; Shear-wave splitting; Small earthquakes

1. Introduction

Shear-waves propagating through anisotropic solids split into two fixed approximately orthogonally polarised phases which travel at different velocities. Throughout the crust, the splitting is typically caused by stress-aligned parallel vertical microcracks (Crampin, 1994, 1999), where the crack density, and crack alignment,

can be estimated from the time-delay between the split shear-waves, and the polarisations of the faster split shear-wave, respectively.

Note that the time-delays in shear-wave splitting (seismic birefringence) are the result of small (*second-order*) differences in the velocities of shear-waves. When split shear-waves are rotated into preferred polarisations, the arrival times of each split shear-wave can, in ideal circumstances, be read as accurately as a first arrival signal with approximately *first-order* precision. This means that time-delays are second-order quantities that have the potential to be read with first-order precision, which typically gives measurements of

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shear-wave time-delays a much higher precision than most other seismic measurements. However, shear-wave splitting above small earthquakes in the crust writes very small changes, usually less than 0.2 s, in the shear-wave arrival, and may be complicated by other *P*- and shear-wave phases and coda waves which can be difficult to recognise.

Table 1 summarises some of the difficulties of reading shear-wave splitting. These difficulties can usually be resolved by visual analysis of polarisation diagrams (PDs, hodograms, or particle-motion diagrams), and by rotating seismograms into preferred orientations, but these techniques are tedious and time-consuming, and

tend to be avoided by many authors. It is sometimes claimed that automatic techniques are preferred because of objectivity, consistency, and repeatability. However, objectivity, consistency, and repeatability are only useful if the measurements are meaningful geophysically. The typical complexity of the shear-wave signals means that automatic techniques have not been wholly successful, which is the justification for developing a combined visual and semi-automatic technique, the shear-wave analysis system (SWAS), in the accompanying paper Gao et al. (2006). This present paper reviews previous techniques for analysing and measuring shear-wave splitting.

Table 1

Reasons why shear-wave splitting above small earthquakes is difficult to measure

	References
Slow shear-waves propagate through higher impedance than fast shear-waves: effects of higher impedance	
1 Slow shear-wave is slower than fast shear-wave	Crampin (1981)
2 Slow shear-wave may be more attenuated than fast shear-wave and may sometimes be unobservable	Hudson (1981), Mueller (1991)
3 Slow shear-wave may have lower frequencies than fast shear-wave	Crampin (1981)
4 Slow shear-wave may be less impulsive than fast shear-wave	Crampin (1981)
Low signal-to-noise ratios: source and effects of noise	
5 Shear-waves are carried in the coda of <i>P</i> -waves	Volti and Crampin (2003a)
6 Slow shear-waves are carried in the coda of the fast shear-waves	Volti and Crampin (2003a)
7 Shear-wave splitting inserts very small (typically <0.2 s) time anomalies into seismic wave-trains which are easily hidden by noise	Volti and Crampin (2003a)
Many techniques mistakenly tend to assume orthogonality: sources of non-orthogonality	
8 Polarisation of shear-waves propagating at the group-velocity are not necessarily orthogonal except in a few isolated symmetry directions	Crampin (1981)
9 Except for normal incidence, any orthogonality of polarisations will be distorted by interaction with the free-surface	Booth and Crampin (1985)
10 Orthogonality may be seriously distorted for propagation near shear-wave singularities — see also Item 14, below	Crampin (1981), Volti and Crampin (2003a), Crampin (1991)
Interaction with surface and sub-surface topography	
11 The effects of the shear-wave window mean that irregular surface and internal topography may have severe effects on shear-wave polarisations, depending on the dip, orientation, distance from recorder, and direction and wavelength of the incident shear-wave	Booth and Crampin (1985), Liu and Crampin (1990)
Other sources of scattering	
12 Inhomogeneities of the geological structure	Crampin (1981)
13 Varying source-time functions. Note that multiple sources can usually be recognised and eliminated from analysis. Note also that since shear-wave splitting is controlled by the path rather than the source the effect of varying source functions can usually be eliminated	Crampin (1981)
14 Shear-wave singularities where the faster and slower split shear-waves have coincident phase velocities can be very common in some microcracked sedimentary sequences. Propagation within $\sim 10^\circ$ of a singularity may cause severe anomalies in time-delays and polarisations. Note that shear-wave singularities are comparatively common in sedimentary basins but are rare above small earthquakes	Crampin (1991)
15 Shear-waves from earthquake radiation patterns range over 360° in polarisation and vary the division of energy between split shear-waves	Crampin (1981)
16 90° -flips in shear-wave polarisations due to critically high pore-fluid pressures on all seismically active faults lead to the typically $\pm 80\%$ variations observed in time-delays between split shear-waves. These 90° -flips are believed to be the major source of scattering above small earthquakes	Crampin et al. (2004)

There are many published observations of shear-wave splitting above small earthquakes. Measuring time-delays requires identifying the arrival times of the fast and slow split shear-waves. The major difficulty in measuring shear-wave splitting above small earthquakes is that shear-waves write complicated signatures into three-component seismograms where polarisations, and particularly time-delays, are heavily scattered and vary widely in time and space (Crampin et al., 1980, 1985, 1990, 1999, 2002, 2003, 2004; Buchbinder, 1985, 1989; Peacock et al., 1988; Shih et al., 1989; Shih and Meyer, 1990; Gledhill, 1991, 1993a, 1993b; Crampin, 1999, 2003; Voli and Crampin, 2003a,b; Gao and Crampin, 2004; amongst many others). As a consequence of the scatter, shear-wave splitting is easily misread and misinterpreted and leads to a number of common fallacies summarised in Table 2. Since it is now claimed that shear-wave splitting monitors the low-level pre-fracturing deformation of *in situ* rocks (Crampin, 1999, 2003), reliable techniques for accurately assessing and measuring the defining parameters are urgently required.

Stratagems for identifying seismic shear-wave arrivals (phase picking) include analysis of: polarisations (Crampin, 1978; Vidale, 1986); energy (Earle and Shearer, 1994; Tong and Kennett, 1996); auto-regression, with the Akaike Information Criteria, AIC (Takanami and Kitigawa, 1988; Leonard and Kennett, 1999; Sleeman and van Eck, 1999); and wavelet transforms (Anant and Dowla, 1997; Tibuleac and Herrin, 1999), amongst many more-or-less minor modifications. Because of the complexity of shear-waveforms, phase picking, and phase identification, mixed techniques are often employed: Cichowicz (1993) presented phase picking techniques based on a filter combining polarisation and energy ratios; Withers et al. (1998) compared several trigger algorithms for phase picking; Bai and Kennett (2000) combined an autoregressive technique with energy analysis; Zhang et al. (2003) combined a wavelet transform with an AIC technique.

Besides these more-or-less deterministic techniques, several techniques using artificial intelligence, AI, techniques have also been developed for phase picking. Chiaruttini and Salemi (1993) developed a seismic net-

Table 2
Common fallacies in measuring and interpreting shear-wave splitting

	Fallacies	Actual behaviour
1	The polarisations of split shear-waves are orthogonal	Polarisations are strictly orthogonal only for <i>phase velocity</i> propagation. Since observations are of ray paths travelling at the <i>group velocity</i> , the polarisations of split shear-waves are strictly orthogonal only along ray paths in a few directions of anisotropic symmetry: see Table 1, Items 8, 9, 10
2	Polarisations of the fast split shear-waves are fixed parallel to cracks	The polarisations and time-delays of split shear-waves, always vary (three dimensionally) with azimuth and incidence angle, in three-dimensions, even when propagating through parallel cracks
3	There is the same percentage of shear-wave velocity anisotropy in all directions, through parallel cracks	The percentage of shear-wave velocity anisotropy always varies (three dimensionally) with azimuth and incidence angle. When propagating through parallel cracks, percentages may vary from positive to negative — there may be zero or negative time-delays in some directions of propagation
4	Polarisations observed at the free-surface are the polarisations along the ray path	Only at normal incidence to a horizontal free-surface are incident polarisations wholly preserved. For all other angles of incidence, projection on to the plane of the free-surface distorts incident polarisations, possibly very severely: see Table 1, Item 9
5	Temporal changes in shear-wave splitting (when monitoring the accumulation of stress before earthquakes, say) will cause changes in time-delays for all ray paths within the shear-wave window	Small changes of stress are only likely to affect crack aspect-ratios, which will change the average time-delay in Band-1 ^a directions of the shear-wave window. Changes in crack density will change <i>average</i> time-delays in Band-2 ^a
6	The shear-wave window, in which shear-waves can be observed at the free-surface undistorted by <i>S</i> -to- <i>P</i> conversions, is aligned normal to the horizontal plane	The shear-wave window is normal to the free-surface within about a wavelength of the seismic recorder. Since earthquakes are typically beneath irregular topography, shear-wave splitting observed above small earthquakes may be severely distorted unless these effects are recognized
7	Shear-wave splitting at the free-surface is confined to the uppermost few kilometers	Although there may be higher crack-induced anisotropy near the surface, widespread evidence suggests there is extensive anisotropy throughout the crust ^b

^a Band-1 is the double-leafed solid-angle of ray path directions making angles 15–45° to the average crack plane in distributions of parallel vertical cracks. Band-2 is the solid-angle of ray path directions $\pm 15^\circ$ to the average crack plane (Crampin, 1999).

^b A recent example is Hiramatsu et al. (2005) who report similar normalised (ms/km) time-delays at all depths down to 50 km.

work analyser, using a ‘blackboard’ technique to interpret signals from a local seismic network. Artificial neural network, ANN, and back-propagation neural network (BPNN) techniques have been applied to picking phase arrivals and identification of seismic arrival types by Murat and Rudman (1992) and Dai and MacBeth (1995, 1997a,b), and applied to shear-wave splitting (Dai and MacBeth, 1994). Expert system (ES) analysis has also been used for earthquake hazard assessment (Zhu et al., 1996), and Tong and Kennett (1996) used ES analysis to pick seismic phase arrivals. However, purely AI techniques are limited in accuracy and are usually adopted only when more deterministic techniques do not work effectively.

Note that there have been many studies of shear-wave splitting by the exploration industry and many sessions and hundreds of papers on shear-wave anisotropy at meetings of the Society of Exploration Geophysicists in USA and the Association of Exploration Geophysicists and Engineers in Europe, and elsewhere. Such techniques measuring shear-wave splitting by controlled-source experiments away from seismic zones are usually easier because there is typically much less scatter than above small earthquakes (Crampin et al., 2004). Shear-wave splitting in exploration studies can be displayed as separately timed arrivals on differently polarised record-sections and is almost universally interpreted as the result of propagation through distributions of approximately parallel vertical microcracks or fractures.

Although many papers describe different techniques for measuring shear-wave splitting, the methods are variations on a few basic mechanisms. Here, we review these basic mechanisms for measuring shear-wave splitting above small earthquakes. These range from what we call *display* techniques requiring visual analysis of unprocessed displays to *automatic* and *semi-automatic* techniques.

2. Difficulties in measuring shear-wave splitting above small earthquakes

Since measuring time-delays in shear-wave splitting requires differencing arrival times of fast and slow split shear-waves, reliable measurement of time-delays require more accuracy than merely picking individual phase arrivals. Difficulties include the following.

2.1. Fast and slow split shear-waves may not have similar waveforms

Slow split shear-waves are slower than fast shear-waves because the differently polarised waves are sub-

ject to different impedances. Consequently, slow waves tend to be more highly attenuated than fast shear-waves (Hudson, 1981) with the loss of higher frequencies, and the variation of the relative attenuation with direction tends to vary inversely with the relative velocity (Hudson, 1981; Crampin, 1984). This means that slow shear-waves tend to be less impulsive and more attenuated than fast shear-waves, and may have significantly different waveforms. In cases of heavy fracturing, slow shear-waves may have such low amplitude that they are difficult to observe (Mueller, 1991).

The shear-wave signals above small earthquakes will also be disturbed by *P*- and *S*-wave coda, other shear-wave arrivals, and surface-waves, which will affect the waveforms of the split shear-waves differently.

2.2. Shear-wave splitting may have low signal-to-noise ratios

Shear-wave splitting signals tend to be noisy as the shear-waves are disturbed by the *P*-wave coda, and the slower split shear-waves are disturbed by the coda of the faster split shear-wave. This may make it difficult to get accurate measurements of shear-wave splitting.

2.3. Shear-wave splitting is measurable only in a very small segment of the seismogram

Examination of polarisation diagrams, PDs, of horizontal particle-motion shows that, although there is typically evidence for approximately orthogonal changes in preferred polarisations at many places along the shear- and surface-wave wave-trains above earthquakes in the crust, the time-delays are small (less than 0.2 s, typically much less, Vosti and Crampin, 2003b). This means that clear measurable separation into two polarisations is confined to a very small segment of the seismograms. Since this is a very small disturbance to the complete shear-wave-trains of small earthquakes, measurable quantities are difficult to identify. Even earthquake doublets with more-or-less identical waveforms typically may show significant variations in time-delays (Lovell et al., 1987).

Note that the fast and slow split shear-waves of, particularly, surface observations above small earthquakes tend to have different waveforms (Section 2.1). This means that cross-correlations of the split shear-waves in most circumstances cannot give reliable measurements of time-delays.

2.4. Problems with non-orthogonality

There are several problems with the *non-orthogonality* of the polarisations of the fast and slow shear-

waves, which may be crucial, as two wave-trains, which are not orthogonally polarised cannot be uniquely separated into two independent signals without making several possibly erroneous assumptions. Many observation techniques assume the two split shear-waves are orthogonally polarised, but theoretically, only shear-waves propagating at the *phase velocity* are strictly orthogonal. Shear-waves propagating along seismic rays at the *group velocity* are only orthogonal in limited symmetry directions (Crampin, 1981). In some circumstances, such as propagating along ray paths close to shear-wave point singularities (where the phase velocity sheets touch), shear-wave polarisations may behave very irregularly including successive arrivals with dramatically non-orthogonal polarisations (Crampin, 1991). Additionally, simple geometry shows that incident orthogonal polarisations are only preserved in surface observations at strictly normal incidence. Orthogonality at all other incidence angles is both geometrically and geophysically distorted.

2.5. Interaction with surface topography

Shear-waves need to be recorded in the shear-wave window immediately above small earthquakes (Booth and Crampin, 1985). The waveforms of shear-waves outside an effective window of $\sim 45^\circ$ can be severely distorted from the polarisations of the incident wave by the effect of *S*-to-*P* conversions on shear-wave energy. Since earthquakes are generally below irregular topography, interpretation must always allow for the possible effects of the topography within about a wavelength of the surface recorder.

2.6. Scatter caused by 90° -flips in shear-wave polarisations

The principal difficulty in measuring shear-wave splitting above small earthquakes is the highly variable nature of the complicated shear-waveforms which vary both spatially and temporally with a scatter in time-delays which is typically as large as $\pm 80\%$ about the mean. Volti and Crampin (2003a) examined possible sources of scatter in a conventional non-critical crust, including: anisotropic variations with direction; errors in earthquake location; reading errors; complicated geology; complicated crack distributions, amongst others. Neither single source nor combination of sources can provide the consistent $\pm 80\%$ scatter in time-delays that is almost universally observed above small earthquakes, and a critical crust was inferred, but no specific source of scatter was suggested.

The variability and scatter has recently been modelled theoretically by assuming critically high pore-fluid pressures on all seismically active faults modifying the local microcrack geometry (Crampin et al., 2002, 2004). In the presence of such high pore-fluid pressures, microcracks are no longer uniformly oriented in regional-stress directions. The shear-wave splitting indicates a critical-system of fluid-saturated microcracks (Crampin, 1999). Increased pore-fluid pressure rearranges microcrack geometry where one of the effects is to modify the geometry so that the faster and slower split shear-wave polarisations exchange directions in what are called 90° -flips.

Note that 90° -flips are comparatively well established. They have been demonstrated theoretically (Crampin and Zatsepin, 1997) and have been observed in over-pressurised oil fields by Crampin et al. (1996) and by Angerer et al. (2002) (who matched a field experiment exactly with modelled effects). They have also been observed above major faults such as the San Andreas and San Jacinto Faults in California and the Húsavík–Flatey Fault in Iceland (Liu et al., 1997; Peacock et al., 1988; Crampin et al., 2002, respectively). However, 90° -flips cannot be modelled in the laboratory as to achieve the 90° -flips, fluid pressures would have to approach confining pressures with possibly damage to the equipment.

Minor variations in the ratio of the length of the path of the ' 90° -flipped' polarisations (with negative time-delays, say) in the vicinity of the fault, and the length of the remainder of the normally pressurised path (with positive time-delays) to the recorder at the free-surface can easily lead to the observed $\pm 80\%$ variations in time-delays (Crampin et al., 2004). Crampin et al. (2002, 2004) show that shear-wave-splitting time-delays are extremely sensitive to details of the triaxial stress field and pore-fluid pressure. Since every earthquake releases stress and modifies stress and pore-fluid pressures, shear-wave splitting time-delays typically vary widely from earthquake to earthquake, even between earthquake doublets. The cause and effects of 90° -flips have been extensively discussed in Crampin et al. (2002, 2004).

Substantial differences between closely spaced surface recordings of polarisations (including near-orthogonal changes) and time-delays caused by 90° -flips in the fault zone (Crampin et al., 2002, 2004) are frequently claimed as indicating concentrations of anisotropy in the near-surface rocks (Savage et al., 1989; Aster et al., 1990; Gledhill, 1990, 1991; Aster and Shearer, 1992; Zhang and Schwartz, 1994; Liu et al., 2004; Peng and Ben-Zion, 2004), as opposed to pervasive anisotropy along the whole ray path. These authors usually attribute the different polarisations to different

mechanisms for anisotropy near the free-surface. The preferred interpretation is that, although there is frequently increased anisotropy as stress is relaxed near the free-surface, the major anomalies are due to the effects of 90°-flips in shear-wave polarisations due to the critically high pore-fluid pressures expected on all seismically active faults (Crampin et al., 2004).

Shear-wave velocity–anisotropy above levels of fracture criticality ($>4.5\%$, Crampin, 1994) indicates such heavily fractured rock that rocks disaggregate, shear strength is lost, and rocks tend to fracture (Crampin and Zatsepin, 1997; Crampin, 1999). When anisotropy is confined to the very near-surface, high values of shear-wave velocity anisotropy are often reported: (15–30%) Savage et al. (1989); (10%) Gledhill (1991); (16%) Zhang and Schwartz (1994); (10–14%) Munson et al. (1995); which are all well above levels of fracture criticality (Crampin, 1994). These are presumably due to the stress-release anomalies in near-surface rocks allowing high crack densities to exist in the absence of consistent tectonic stress. Typically, rose diagrams of polarisations in such circumstances still show comparatively linear (parallel) orientations, which would not be expected in the absence of shear strength. This suggests that pervasive anisotropy along most of the ray path is the major source of shear-wave splitting, although substantial near-surface anisotropy may still exist.

Note higher percentages of crack-induced anisotropy are sometimes reported from ultrasonic experiments in rock physics laboratories, for example, Rosolofosaon et al. (2000), amongst many others. True-triaxial stress is very difficult to organise in the laboratory. Consequently, typical rock physics stress-cells only model confining stress (and occasionally uniaxial stress), however, without true-triaxial stress there is less tendency to fracture, the fracture criticality limit can be exceeded, and high degrees of crack-induced anisotropy may be readily observed. Observations of shear-wave splitting in the field demonstrate that almost all *in situ* rocks are subject to true-triaxial stress that orients microcracks at levels of stress below those leading to fracture criticality and fracturing.

Above earthquakes, 90°-flips in shear-wave polarisations are directly observed only for recordings close to major (seismogenic) faults where most of the ray path is close to critically high pore-fluid pressures present on all seismogenic fault planes (Crampin et al., 2002, 2004). Shear-wave splitting in reflection surveys or vertical-seismic-profiles in seismic exploration typically varies smoothly, away from areas of seismicity and seismically active faults. The splitting may show even dramatic changes in polarisation or time-delays but, with-

out the presence of critically high pore-fluid pressures, time-delays do not display the $\pm 80\%$ scatter observed above small earthquakes. Note that Angerer et al. (2002) directly observed 90°-flips of shear-wave polarisations when critically high pressures CO_2 were injected into a carbonate reservoir at ~ 600 m-depth. The flips did not occur for a low-pressure injection (Angerer et al., 2002).

These numerous difficulties, summarised in Table 1, make it difficult to routinely monitor and measure shear-wave splitting above small earthquakes. The scatter, and spatial and temporal variations, above small earthquakes means that classic examples of shear-wave splitting, where two similar orthogonally polarised arrivals are separated by a time-delay, only occur in typically 20%, say, of earthquake records.

3. Display techniques for measuring shear-wave splitting

3.1. Polarisation diagrams

Polarisation diagrams, PDs, or hodograms, are (usually horizontal) sections of the particle-motion typically displayed for successive time-intervals on three-component seismograms. The first use of PDs to display the effects of anisotropy on seismic particle-motion is probably that of Crampin and King (1977), who used PDs to display the effects of upper mantle anisotropy on the coupling of higher mode surface-waves propagating across Eurasia. The PDs showed elliptical coupled Rayleigh–Love wave motion indicating pervasive anisotropy in a thin layer in the uppermost mantle over most of Eurasia.

Crampin (1978) used PDs to display the effects of shear-wave splitting in synthetic body-wave seismograms propagating through models of fluid-saturated microcracks. Crampin et al. (1980) used PDs to display shear-wave splitting in the first confirmed observations of shear-wave splitting above small earthquakes in the crust. Crampin et al. (1985) also used PDs (without rotating seismograms) to measure shear-wave splitting in the Turkish Dilatancy Projects, which was the first comprehensive examination of shear-wave splitting above small earthquakes.

The problems discussed in the previous section mean that shear-wave splitting writes complicated signatures into three-component seismograms. These may be difficult to interpret, but the diagnostic feature of anisotropy-induced shear-wave splitting in PDs is abrupt *nearly orthogonal* changes in the directions of the horizontal particle-motion. These are always visible as long as the wave-trains have not been low-pass filtered too aggres-

sively. Consequently, it is usually easy to demonstrate shear-wave splitting and measure polarisation directions and time-delays (number of samples) directly from PDs. As a consequence, PDs and linearity tests may be used as convenient tools to display the effects of shear-wave splitting in almost all seismograms.

Linearity tests are what we call displays of PDs of the horizontal particle-motion, where the two preferred orientations have been shifted by removing the time-delay. These are frequently used to demonstrate the quality of the measurements of shear-wave splitting by the linearity of the resulting PD.

One of the important features of PDs is the insight they can give into the behaviour of shear-wave splitting. For example, clusters of earthquakes frequently show broadly comparable signatures (see note on doublets in Section 2.3, above), with patterns of similar overall behaviour, but with minor-to-major differences in the usually very small time-interval containing the shear-wave splitting (Lovell et al., 1987). This is thought to be the result of the sensitivity of shear-wave splitting to the changes in triaxial stress and pore-fluid pressure leading to 90°-flips modifying crack geometry after every earthquake. Such repeating behaviour sometimes leads to clarification of how to interpret the shear-wave splitting.

3.2. Rotated seismograms

Rotating seismograms into the observed polarisations of the split shear-waves is another convenient tool for all investigations of shear-wave splitting. Ando et al. (1980), in the first observations of shear-wave splitting in the upper mantle (above intermediate-depth earthquakes), systematically rotated horizontal seismograms and selected preferential polarisations that indicated most clearly the separation of the different shear-wave arrival times.

The combination using PDs to determine preferred polarisations for seismogram rotation soon became standard, both above small earthquakes in the crust (Buchbinder, 1985, 1989; Gao et al., 1998; Volti and Crampin, 2003a,b; Gao and Crampin, 2003, etc.) and in upper mantle studies (Bowman and Ando, 1987; Silver and Chan, 1988, 1991).

Note however the problems with non-orthogonality discussed in Section 2.4, above. If the horizontal projections of the polarisations are not strictly orthogonal, as is typically the case in horizontal PDs, the polarisations cannot be separated into uniquely independent wave-trains. However, distinct abrupt changes of particle-motion direction in PDs will typically still be observed.

3.3. Vectorial polarisation diagrams

Bernard (1987) and Bernard and Zollo (1989) developed a shear-wave display where the horizontal particle-motion vectors are plotted from an origin moving linearly with time. Although these elegant, even beautiful, plots (Iannaccone and Deschamps, 1989; Zollo and Bernard, 1989) – often reminiscent of the flight of birds with feathered wings – clearly demonstrate the presence of shear-wave splitting, neither polarisation angles nor time-delays are easy to interpret or measure.

3.4. Complex-polarisation analysis

Vidale (1986) developed a complex-polarisation analysis of particle-motion where the variation of the eigen values and eigen vectors of complex covariance matrices can be interpreted in terms of the variations of the dip, strike, angular and linear polarisation of the shear-wave (and surface-wave) particle-motion. Vidale's is a basic technique that can be used for automated processing (see Section 4.5, below), as well as to visually examine and measure polarisations and time-delays of shear-wave splitting in plots of the various parameters (Silver and Chan, 1988, 1991; Zhang and Schwartz, 1994; Munson et al., 1995; Shih et al., 1989, 1991). However, these plots are usually complicated to interpret and only provide measurements that are more easily and directly obtained from PDs.

4. Automatic techniques for measuring shear-wave splitting

Automatic measurements of shear-wave splitting are usually variations of a few basic automatic techniques.

4.1. Cross-correlation techniques

Fukao (1984) was the first to use automated analyses of shear-wave splitting on teleseismic shear-wave reflected from the Earth's core. Using horizontal PDs to indicate preferred polarisations, Fukao (1984) estimated shear-wave splitting time-delays from the lags of cross-correlations of the preferentially rotated wave-trains. Fukao displayed the results in PDs, and used linearity tests to demonstrate the quality of the measurements. He showed that fast split shear-wave polarisations over the whole of Japan are uniformly NE to NNE and parallel to the subduction direction of the Philippine and Pacific Plates beneath Japan. Since Fukao (1984), cross-correlation techniques have been used extensively to measure time-delays in most automatic measurement

techniques, both above small earthquakes in the crust, and in upper mantle anisotropy.

Gao and Zheng (1995) developed a cross-correlation technique, where contoured plots of the cross-correlation coefficient against azimuth and time-delay allow polarisation and time-delay to be estimated. PDs and linearity tests were used to evaluate restively isolated, as in the measurements of shear-wave splitting above swarms of small earthquakes (Gao et al., 1998), but is not generally applicable because to the specific limitations discussed in the next section.

Generally, cross-correlation techniques are frequently used to check the quality of estimated polarisations and time-delays determined by other techniques, however, cross-correlations suffer from at least two major difficulties. One difficulty is that shear-wave splitting writes sometimes very small, rather subtle anomalies into a comparatively large shear- and surface-wave signal, and cross-correlations are not very sensitive to such small changes, unless very severe selection criteria are followed often rejecting 50–70% of the data (Liu et al., 2004; Teanby et al., 2004a). Such rejections are invidious as they may strongly bias the interpretation. The other difficulty is the sensitivity of cross-correlation techniques to the choice of window end points, particularly when the window has to end within the surface-wave-train which frequently overlaps the shear-wave signal.

4.2. Problems with cross-correlations

- (1) It is often difficult to choose suitable ends to windows for cross-correlation. The effect of different end points, and different methods of ending the window, in the middle of the relatively large-amplitude shear- and surface-wave coda can easily dominate the comparatively small effects of shear-wave splitting on a small segment of the wave-train.
- (2) The waveforms of the fast and slow split shear-wave arrivals are seldom wholly similar (Section 2.1), where even good examples of shear-wave splitting would have low cross-correlation functions. For example, Figs. 2 and 3 of Gao et al. (2006) show figures with rotated seismograms showing what are, we suggest, clear unambiguous examples of shear-wave splitting. However, the initial fast and slow arrivals are so different in waveforms, with different frequencies, and different kinks in the waveforms (as other phases arrive), that cross-correlations coefficients would be low and effects would be dominated by the effects of the ends of the windows.

- (3) The polarisations of the fast and slow split shear-wave are seldom strictly orthogonal and cannot uniquely be rotated into independent fast and slow shear-wave wave-trains. This means that direct cross-correlation of independent fast and slow shear-wave-trains is frequently impossible.

These various difficulties apply to almost all surface observations of shear-wave splitting above small earthquakes in the Earth's crust. Teleseismic shear-waves propagating through the mantle usually have much lower frequencies (typically 0.1–2 Hz, as opposed to usually greater than 5 Hz above crustal earthquakes), and fast and slow shear-waves usually have similar waveforms, and do not have the heavy shear- and surface-wave coda typical of crustal earthquakes. Teleseisms also tend to arrive close to normal incidence and thus can be directly separated into independent wave-trains by rotating horizontal axes. Consequently, cross-correlation is a useful technique for estimating values of shear-wave splitting in the mantle (Silver and Chan, 1988, 1991) that cannot always be reliably applied to surface observations above earthquakes in the crust.

- (4) Teanby et al. (2004a), recognising the problems of window selection, use cluster analysis to select optimum windows for Silver and Chan (1991) cross-correlation measurements of shear-wave splitting time-delays and polarisations. Teanby et al. applied this technique to subsurface measurements of time-delays and polarisations of subsidence-induced events in the Valhall North Sea oil field. These subsurface recordings are without the surface complications of surface-wave coda and lack of orthogonality, and are more comparable to the upper mantle arrivals in the previous section than to surface observations of small earthquakes. Consequently, Teanby et al. (2004a) are able to obtain very effective almost wholly automated measurements consistent with (laborious) manual visual techniques. Nevertheless, this success is at the expense of rejecting ~50% of the data for variety of reasons (low signal-to-noise ratio, interfering phases, low energy, etc.), and visually examining each record for quality control, which led to a ~40% rejection.

Note that such cluster analysis techniques cannot be applied to surface observations above small earthquakes (the source of the majority of non-exploration measurements of shear-wave splitting). At the surface, interfering body-wave phases and surface-waves severely disturb the results of the Silver and Chan technique.

4.3. Linearity techniques: aspect-ratio

Several automatic techniques measure the linearity of the fast split shear-wave arrival. Shih et al. (1989) and Shih and Meyer (1990) developed an ‘aspect-ratio’ technique by sequentially rotating seismograms for maximum linearity. The particle-motion of the possible fast shear-wave is projected onto orthogonal axes in, typically, the horizontal plane and the ratio of the projections calculated as the function of the azimuth of the projection axes. The polarisation is the azimuth at which the maximum aspect-ratio occurs where the aspect-ratio is the ratio of the two projections (Vidale, 1986; Shih and Meyer, 1990). Clearly the aspect-ratio technique will not work well if the time-delay between the split shear-waves is too small (although Gledhill, 1991, was able to measure parameters for as little as three samples), or when the polarisations are not orthogonal. Shih et al. (1991) used this technique very effectively in a comprehensive analysis of signals from the ~ 160 km-deep seismicity of the Bucaramanga Nest swarm in Columbia, where time-delays were large enough (0.19–0.39 s) and the arrivals were close enough to vertical to overcome many of the difficulties of non-orthogonality in Section 2.

Note that the use of ‘aspect-ratio’ by Shih et al. (1989, 1991) as length/width is the inverse of the ‘aspect-ratio’ usually used for crack distributions which is width/length (Hudson, 1981; Crampin, 1984).

Shih et al. (1991) independently calculated time-delays by aspect-ratio techniques and by cross-correlating rotated seismograms. The average time-delays at nine stations from the aspect-ratio and cross-correlation techniques differed by an average of 11%. Since the wave-trains were noisy and complicated this was an excellent demonstration of the effectiveness both aspect-ratio and cross-correlation techniques for these near-normal shear-wave arrivals.

Gledhill (1991, 1993a,b) used two techniques: the Chen et al. (1987) visual display technique (albeit in an interactive graphics display); and the Shih et al. (1989) and Shih and Meyer (1990) aspect-ratio technique to make a very detailed examination of shear-wave splitting observed by seismic stations on the Wellington Peninsula, New Zealand. The peninsula is very hilly and Gledhill observes a wide variety of shear-wave polarisations, some variations are associated with topographic or tectonic irregularities, and some are associated with local topographic lineations. However, the largest anomaly is three recorders in a ~ 7 km line, with similar topography, parallel and equidistant from two major faults, where the polarisation of one recorder is orthogonal to the others. Neither Gledhill nor we offer any firm

explanation, but we suspect it is due to irregular surface topography.

4.4. Linearity technique: linearity interval

Aster et al. (1990), in a very detailed analysis of temporal changes in shear-wave splitting, monitored the polarisations and time-delays of shear-wave splitting by assessing the linearity interval by the magnitude of the vector linearity in the direction of the largest eigen value of the variance tensor. This paper was essentially a critical examination of the waveform data set used by Peacock et al. (1988) and Crampin et al. (1990), who had respectively found temporal variations in shear-wave splitting time-delays before the 1986 $M=6$ North Palm Springs Earthquake. Aster et al. (1990) “...find a hint of the temporal variations which Peacock et al. and Crampin et al. noted but with no clear association of these variations with the North Palm Springs earthquake.” However, many of their figures do not show what is claimed. For example, their Fig. 15 shows “similar pairs” of events, sometimes separated by up to 5.5 years which have been “aligned on the maximum of the three-component cross-correlation function” which “limits temporal variations in the shear-wave splitting delay time to less than 0.004 s.” In fact, the events are not similar. The pairs of three-component records show substantial differences in character (only one pair, out of 19, show nearly identical fast and slow shear-wave arrivals, and first arrivals are not aligned).

This is a demonstration of the difficulty of measuring shear-wave splitting discussed in Section 2, trying to measure a very small element of the seismogram in a very noisy record, where the crucially important window is difficult to define. Objective measurements are only useful if they measure appropriate phenomenon. These do not. Crampin et al. (1991) showed several other examples where the arguments of Aster et al. (1990) are not correct. This application of the linearity technique clearly has serious errors, and Crampin et al. (1991) showed that some of the time-delays of Aster et al. (1990) are in error by up to 200% (factor of 3).

4.5. Linearity techniques: singular covariance matrix

Adapting the complex-polarisation technique of Vidale (1986), Silver and Chan (1988, 1991) estimate polarisations and time-delays by searching for non-zero eigen values of singular covariance matrices. That is they search for separate linear polarisations. This technique is highly effective for low-frequency arrivals from the

upper mantle where signals are well separated, but is much less useful in the typical highly variable shear-wave splitting above small earthquakes in the crust where the shear-waves arrivals typically are part of a substantial shear- and surface-wave coda.

Note that [Teanby et al. \(2004a,b\)](#) used the covariance technique of [Silver and Chan \(1991\)](#) to automatically measure shear-wave splitting from hydrocarbon reservoir-induced events recorded by borehole recorders. As mentioned above, the subsurface recordings of very small very local earthquakes were simple signals without near-surface and surface-wave contamination, and the records successfully showed temporal variations in shear-wave splitting.

In contrast, [Peng and Ben-Zion \(2004\)](#) also used the program of [Silver and Chan \(1991\)](#) to monitor and measure shear-wave splitting in the pronounced crustal seismicity along the Karadere–Düzce branch of the North Anatolian Fault in the 6-month period following the 1999 M_w 7.4 Izmit and M_w 7.1 Düzce earthquakes. In an attempt to treat crustal earthquakes, [Peng and Ben-Zion \(2004\)](#) pass the $\sim 20,000$ event data through 10 ‘objective’ quality criteria which left only $\sim 30\%$ of the records as acceptable data. The results are extremely scattered, and likely to be biased. The shear-wave arrivals begin complicated shear- and surface-wave waveforms and do not have the simplicity of upper mantle arrivals. However, the major problem is local topography. Most recorders are located in a comparatively narrow valley on the edge of the Almacik Block which rises from ~ 100 to ~ 1500 m in ~ 3 km (Fig. 10 of [Peng and Ben-Zion, 2004](#)). In such irregular topography, the shear-wave polarisations are heavily distorted by topography immediately around the recording station, and the records cannot easily be interpreted ([Booth and Crampin, 1985](#); [Evans et al., 1987](#); [Crampin and Gao, 2005](#)).

4.6. Linearity techniques: single-value decomposition

[Shieh \(1997\)](#) used a single-value decomposition technique to decompose an orthogonal matrix of (time-windowed) three-component seismograms into preferred polarisations, and estimate the time-delay as the lag of cross-correlation of the faster and slower split shear-waves. As far as we know, this comparatively simple technique has not been widely used. However, it seems very promising, as it appears to give stable results even in the presence of noise, although it would still be subject to the restrictions listed in Section 4.2, above.

4.7. Combination of cross-correlation and linearity techniques

[Liu et al. \(2004\)](#), in a comprehensive study of aftershocks of the 1999, $M_w = 7.6$ Chi-Chi Earthquake, Taiwan, were able to successfully use both cross-correlation and aspect-ratio techniques. This was because the very low near-surface velocity (~ 400 m/s) meant that all arrivals were incident nearly vertically at the surface and within the shear-wave window. Consequently, the fast and slow waves were almost orthogonally polarised and there was minimal P -wave or shear-wave coda. In these circumstances, aspect-ratio and cross-correlation techniques gave virtually identical results.

[Liu et al. \(2004\)](#) specifically studied earthquakes for 2.7 years before the main shock and 2 years of aftershocks associated with a major $M_w = 6.4$ aftershock 1 month after the main shock. Their analysis showed, as expected, no variation of shear-wave splitting during the aftershock sequence. However, despite the claim in the title of [Liu et al. \(2004, 2005\)](#), [Crampin and Gao \(2005\)](#) showed that time-delays calculated by Liu et al. before the main shock displayed the characteristic pattern of temporal variations seen before some 15 earthquakes worldwide ([Crampin and Gao, 2005](#)).

4.8. Artificial neural networks for analysis of shear-wave splitting and other seismic arrivals

Artificial neural networks (ANNs) attempt to recognise complex non-linear patterns by simulating the complex interconnections and interaction between neurons in the brain ([McCormack, 1991](#)). [Dai and MacBeth \(1994\)](#) examine the application of ANNs to analysing shear-wave splitting. The general conclusion is that although the technique has promise, shear-wave splitting is so highly non-linear that developing appropriate training sets is too time-consuming and restrictive to be useful.

[Dai and MacBeth \(1995\)](#) also review other techniques and develop ANNs for picking seismic arrivals on three-component seismograms. It is clear that, although there is promise, neither ANNs nor other AI techniques ([Takanami and Kitigawa, 1988](#); [Cichowicz, 1993](#); [Chiaruttini and Salemi, 1993](#); [Roberts et al., 1993](#); [Tong, 1995](#); [Sleeman and van Eck, 1999](#)) are going to provide sufficient accuracy for reliable routine measurements of shear-wave splitting.

5. Discussion

Sections 3.1–3.4 report several *display* techniques where various plots are visually examined for polarisa-

tions and time-delays. Examination of particle-motion of shear-wave splitting in PDs, for example in Gao et al. (2006) indicates extremely complex non-linear behaviour. Since there is no wholly objective way to measure shear-wave splitting, display techniques are probably the most accurate, and certainly the most flexible, techniques but are time-consuming and tedious. Sections 4.1–4.8 report *automatic* techniques, which automatically determine polarisations and time-delays. However, these automatic techniques can only produce even relatively accurate measurements of polarisations and time-delays for near-classic examples of shear-wave splitting. Consequently, wholly automatic measurement can only be used if the input data is subjected to rigorous elimination of all except nearly classic examples of shear-wave splitting involving orthogonal polarisations of impulsive arrivals well separated from shear- and surface-wave coda. Such rigorous elimination of most of the data is undesirable in seeking temporal changes in shear-wave splitting before earthquakes where there is typically minimal data. Such rigorous elimination may also strongly bias the interpretation which could lead to incorrect conclusions.

5.1. Subjectivity and objectivity

Display techniques involving visual examinations are frequently alleged to be subjective, due to the necessary visual judgements of the polarisations, arrival times of shear-wave arrivals, and what observations to apply and how to interpret them. At various times, subjectivity of display techniques to measure shear-wave splitting has been tested by several analysts who, given the procedures listed in Chen et al. (1987), say, obtained consistent overall measurements on the same data set. Note that the invariable large $\pm 80\%$ scatter in time-delays above small earthquakes (see Section 2.5 and Crampin et al., 2004) means that the effects of subjectivity may be less important than is sometimes implied.

Note also that all automatic techniques also require visual evaluations or judgements, albeit only in setting window lengths, minimum signal-to-noise ratios, rigorous rejection criteria, and other phenomena, as in the two most successful techniques of Liu et al. (2004) and Teanby et al. (2004a).

5.2. Suspect objectivity

Visual examination of shear-wave splitting is subjective, and it is comparatively easy to develop a computer program in an attempt to eliminate subjectivity, as we have shown above. Numerous papers claim objectivity

by subjecting seismograms to an automatic technique with minimal human interaction. However, using a computer program to measure some parameter, is only useful if the program can be adequately demonstrated as measuring the intended parameter. Without such demonstration, the ‘objectivity’ is worthless, may be misleading, and has suspect objectivity. We suggest that, without exception, the automatic techniques, reported in Section 4, have suspect objectivity.

5.3. Development of the semi-automatic shear-wave analysis system

We suggest that no existing display techniques or automatic techniques are wholly satisfactory. Combining the advantages of both display and automatic techniques in the shear-wave analysis system (SWAS), suggested in the accompanying paper (Gao et al., 2006), is probably the best option for measuring the polarisations and time-delays of shear-wave splitting above small earthquakes.

Since it has been argued that shear-wave splitting, by opening a window into the pre-fracturing deformation of the fluid-saturated microcracks in almost all *in situ* rock, is the key to a ‘New Geophysics’ of a monitorable, calculable, predictable, and in some circumstances even controllable crust (Crampin, 2003, 2004; Crampin et al., 2003; Crampin and Peacock, 2005). We hope that SWAS will make it easier to measure shear-wave splitting above small earthquakes. This paper demonstrates it is needed.

Acknowledgements

This work was supported partly by: European Commission PREPARED Project EVG1-CT2002-00073; China NSFC Project 40274011; China MOST Project 2004BA601B01; China MOP Key Project 2003; a UK Royal Society Joint-Project with China.

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