A four-year study of shear-wave splitting in Iceland: 1 – background and preliminary analysis

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Abstract: This is Part 1 of a report of a four-year study of seismic shear-wave splitting in Iceland designed to seek temporal variations before earthquakes. Shear-wave splitting is observed routinely in Iceland whenever shear-waves arrive within the shear-wave window of seismic stations, and whenever adequate data are available, temporal and spatial variations in shear-wave splitting are observed before both earthquakes and volcanic eruptions. Shear-wave splitting is caused principally by the stress-aligned fluid-saturated microcracks and pore throats in almost all in situ rocks. Fluid-saturated microcracks are the most compliant elements of the rock mass, and changes in splitting can be directly interpreted and modelled as the effects of changing stress on the microcrack geometry in the rock mass often at considerable distances from the immediate earthquake source zone. Such changes were found and are reported in Paper 2. This paper presents the background, preliminary observations, and analysis of shear-wave splitting in Iceland.

Keywords: Iceland, shear-wave splitting, stress, stress-forecasting earthquakes, volcanoes.

This paper and Paper 2 (Volti & Crampin 2002) summarise the results of a four-year study of seismic shear-wave splitting data in Iceland. The European Commission funded PRENLAB Projects, 1996 to 2000, used Iceland as a natural laboratory for earthquake prediction research. PRENLAB extended the SIL seismic network (South Iceland Lowland network of seismic stations, Stefánsson et al. 1993; Böðvarsson et al. 1999), and made various geophysical and geological studies throughout Iceland, including studies of shear-wave splitting above small earthquakes. Iceland is on an offset of the Mid-Atlantic Ridge, and this study concentrates on the major concentration of seismicity over the transform zone which is onshore in SW Iceland (described by Menke et al. 1994) and is sometimes known as the South of Iceland Seismic Zone, or SISZ. Shear-wave splitting is observed whenever shear-waves were recorded within the shear-wave window below seismic stations.

We suggest that the observed changes in time-delays are not earthquake precursors in the usual sense of the word. Shear-wave splitting is monitoring the build up of stress before larger earthquakes and hence is monitoring the primary driving force of all earthquakes. Without the stress build up, there would be no earthquakes. Consequently, with appropriate
observations, changes in shear-wave splitting can always be observed before earthquakes, and inferred changes in crack geometry can be directly interpreted in terms of changes in stress and the approach of fracture criticality and earthquakes.

In this paper, we describe the background of shear-wave splitting and measurement techniques, and the preliminary observations in Iceland. Changes in the time-delays between split shear-waves were identified before both earthquakes and eruptions at the few stations where there were adequate temporal and spatial data to monitor temporal variations. The time and magnitude of an $M_5$ earthquake in Iceland was successfully stress-forecast (Crampin et al. 1999a). This and other temporal variations are reported in Paper 2.

**Forewarnings**

The major problem with this studies is that we are unable to resolve the exceptionally large ($\pm 80\%$) scatter in the measured time-delays between the split shear-waves (see for example Fig. 1 of Paper 2). The five possible sources of scatter identified in conventional (non-critical) geophysics cannot explain the large scatter, and we have to appeal to the geophysics of critical systems. This we suggest is plausible. The 1.5\% to 4.5\% shear-wave velocity anisotropy observed in most rocks in the uppermost 15km of the crust (Crampin 1994) imply distributions of very closely-spaced fluid-saturated microcracks which are near to levels of fracture-criticality at $\sim 5.5\%$ shear-wave velocity anisotropy (Crampin & Zatsepin 1997). This suggests that the microcrack distributions in most rocks are close to a critical state. Critical states imply heterogeneities and clustering at all scales in both time and space (Bruce & Wallace 1989; Crampin & Chastin 2001). Consequently, seismic waves travelling through rocks on the verge of criticality will be scattered and in particular the scattering will vary with time. Since it can be shown that shear-wave splitting is controlled by and is extremely sensitive to changes in low-level deformation (Crampin & Zatsepin 1997), the effects will be most severe on the scattering of shear-wave splitting.

**Background**

In anisotropic elastic solids, seismic shear-waves travelling at the group velocity split into two nearly-perpendicular polarisations, which travel at different velocities and have polarisations and velocities that are fixed for each direction of travel in each anisotropic symmetry system (Crampin 1981). Shear-waves with approximately orthogonal polarisations separate in time and write characteristic signatures into the particle motion that can be identified and analysed either in polarisation diagrams (hodograms) or by rotating
horizontal seismograms parallel and perpendicular to the preferred polarisations. The distinctive signature of azimuthally varying shear-wave splitting, where the faster polarisation is approximately parallel to the direction of maximum horizontal stress, is widely observed in almost all \textit{in situ} rocks below a critical depth, usually between 500m and 1km below the surface (Crampin 1994). At shallower depths, shear-wave splitting may be disturbed by near-surface stress-release anomalies and the are effects often dominated by lithological and petrological phenomena.

Such stress-aligned shear-wave splitting was first positively identified in the shear-wave window above small earthquakes in the Turkish Dilatancy Projects (Crampin \textit{et al.} 1980, 1985). Since then, stress-aligned shear-wave splitting has been observed in a great variety of sedimentary, igneous, and metamorphic crustal rocks below the critical depth in controlled-source exploration seismology, as well as above small earthquakes (reviewed by Crampin 1994; see also Crampin 1996, & Winterstein 1996). The splitting is caused by propagation through distributions of fluid-saturated grain-boundary cracks and low aspect-ratio pores that become aligned in the stress field (Crampin 1994; Crampin & Zatsepin 1997). These distributions of aligned “cracks” are known as \textit{extensive-dilatancy anisotropy (EDA)} and the individual inclusions as EDA-cracks (Crampin \textit{et al.} 1984). There are only a few well-understood exceptions where \textit{in situ} rocks do not contain EDA-cracks (Crampin 1994, 1999).

\textit{Evolution of fluid-saturated cracks}

Fluid saturated EDA-cracks are stress-sensitive and highly compliant. This means that cracks respond immediately if the stress in the rockmass changes however marginally. The mechanism for the response (the mechanism for deformation) is fluid migration by flow or diffusion along pressure-gradients (Brodie & Rutter 1985; Rutter & Brodie 1991) between neighbouring grain-boundary cracks, and low aspect-ratio pores at different orientations to the stress field (Zatsepin & Crampin 1997; Crampin & Zatsepin 1997). Since shear-wave splitting is sensitive to details of microcrack geometry, changes in shear-wave splitting can be used to monitor stress-induced changes to microcrack geometry. The evolution of such stressed fluid-saturated microcracks under changing conditions has been modelled by \textit{anisotropic poro-elasticity (APE)} (Zatsepin & Crampin 1997; Crampin & Zatsepin 1997). APE-modelling of fluid-saturated crack evolution in typical stress fields in the Earth leads to crack distributions of approximately-parallel approximately-vertical cracks below the critical depth (Crampin & Zatsepin 1997). Such cracks have hexagonal anisotropic
symmetry (transverse isotropy) with a horizontal axis of symmetry (TIH-anisotropy), or minor perturbations thereof.

APE modelling matches a very wide range of some 15 different static and dynamic phenomena associated with shear-waves, cracks, and the occurrence of earthquakes (Crampin 1999). APE shows that the dominant effect of increasing stress, in the low-level pre-fracturing stage before earthquakes or eruptions occur, is to increase the aspect-ratios of cracks perpendicular to the minimum compressional stress (Crampin & Zatsepin 1995, 1997), confirming the empirical hypotheses of Peacock et al. (1988). The good match of APE-modelling to observations of a variety of phenomena (Crampin 1999) is strong confirmation that the major cause of shear-wave splitting is microcracks rather than macrocracks, which would be much less compliant to small perturbations.

Previous observations of temporal changes

Changes in shear-wave polarisations, which would indicate changes in stress direction, have not been reliably observed. Changes in time-delays vary with azimuth and incidence angle within the shear-wave window. Consequently, the shear-wave window has been divided into two segments (Fig. 1) sensitive to crack aspect-ratios and crack densities, respectively. Changes in crack aspect-ratio of near-vertical parallel cracks cause changes in time-delays along ray paths within Band-1 of the shear-wave window (Crampin 1999). Band-1 is the double-leafed solid angle of ray paths with angles 15° to 45° either side of the average plane of the crack distributions (between 45° and 75° to the crack normal). Time-delays between split shear-waves in Band-1 are sensitive to crack aspect-ratio and hence sensitive to increasing stress, where increases of aspect-ratio increase the average time-delays in Band-1 (Crampin & Zatsepin 1997; Crampin 1999). Arrivals in Band-2 (ray paths ±15° to the crack plane) are sensitive principally to crack density.

The effects of increases of stress on crack aspect-ratios can continue until the cracking is so pronounced that shear strength is lost and fractures and earthquakes can occur. This happens at a level of cracking known as fracture criticality. First identified in observations by Crampin (1994), Crampin & Zatsepin (1997) show that fracture criticality is equivalent to the percolation threshold of stress-aligned fluid-saturated cracks (at about 5.5% shear-wave velocity anisotropy). At fracture-criticality, the cracking is so extensive that through-going cracks necessarily exist, and there is fracturing, faulting, earthquakes, and in some cases volcanic eruptions. Following fracturing and faulting, pore fluids are released, the
stress is relaxed and, without fluids to support aspect-ratios, cracks close, and the level of cracking relaxes to sub-critical.

Note that the observed levels of fracture-criticality, when earthquakes occur at time-delays of 11 to 14ms/km, is higher in Iceland (Paper 2) than typically found elsewhere (Crampin 1999). Levels of observed fracture-criticality vary strongly (Crampin 1993b) with properties of the matrix rock such as Poisson's Ratio, and with properties of the pore-fluid (pressure, acoustic velocity, viscosity, etc.). The expected cause is the high heat-flow in Iceland.

Variations in the time-delays between split shear-waves in Band-1 have been observed before several earthquakes worldwide. These are: the $\text{Ms} \, 6$, North Palm Springs Earthquake of 8th July 1986, in Southern California (Peacock et al. 1988; Crampin et al. 1990, 1991); an $\text{Ms} \, 4$, Parkfield earthquake (Liu et al. 1997); and before larger earthquakes in two isolated swarms (Crampin 1993a) at Enola, Arkansas (Booth et al. 1990), and Hainan Island, South China Sea (Gao et al. 1998). Time-delays (normalised by path length to ms/km) indicate aspect-ratios increasing until the (normalised) level of fracture criticality is reached when the earthquake or eruption occurs. Both rates of increase of time-delays, implying increasing aspect-ratios, and levels of fracture criticality, just before larger earthquakes occur, can be estimated from the variation in time-delays before larger earthquakes.

Continuing studies at Parkfield following Liu et al. (1997) would be possible. However, despite extensive investigations, such as the series of five papers concluding with Nadeau & McEvilly (1997), to our knowledge, no further studies of temporal variations of shear-wave splitting have been made at Parkfield.

The high seismicity, the extensive seismic network, and the rapid analysis and location procedures developed by the Icelandic Meteorological Office during the PRENLAB Project provided both the restrictive recording geometry and excellent facilities for monitoring changes in shear-wave splitting before earthquakes. This paper reports and interprets the changes in shear-wave splitting seen before earthquakes and volcanic eruptions, and discusses their implications for monitoring pre-fracturing deformation in crustal rocks and stress-forecasting earthquakes elsewhere, away from the high seismicity of SW Iceland.

It is worth noting that one of the aims of the SIL network was to record and locate small magnitude earthquakes (Stefánsson et al. 1993). The earthquakes in Table 1, below, show that this study would not have been possible without data from small magnitude earthquakes.
Preliminary notes

1) Observations of temporal changes in shear-wave splitting above small earthquakes due to the build up of stress before larger earthquakes or volcanic eruptions require a restrictive source-receiver geometry with seismic activity adequately distributed, both temporally and spatially, within the shear-wave window at each station. Earthquakes are typically clustered in space and time and the need for well-distributed activity within about 8km (the average focal depth) of the station for adequate data imposes severe constraints, even in areas of comparatively high seismicity such as SW Iceland.

2) The shear-wave window is the vertical cone of directions bounded by $\sin^{-1}(Vs/Vp)$ which is equal to 35.26° for Poisson’s ratio of 0.25, where shear-wave seismograms are not disturbed by $S$-to-$P$ conversions at the free surface (Booth & Crampin 1985). However because of ray-curvature due to low-velocity near-surface layers, the effective straight-line cone of epicentral distance over focal depth can usually be extended to ±~45°. Note that because of ray curvature through the low velocity near-surface layers, wider-angle arrivals will also be in the shear-wave window (Menke et al. 1994). These will give consistent polarisation data, which are controlled primarily by near-recorder structure, but the more complicated ray path precludes interpretation of time-delays between split shear-waves. Since analysis of time-delays is the key to monitoring the effects of changing stress (Paper 2), we restrict observations to shear-wave arrivals within the straight-line cone ±~45°.

3) Observations of normalised time-delays between split shear-waves, the key parameter in this paper, show a very large scatter with the amplitude of the variations seen before earthquakes (a progressive increase ending in an abrupt decrease) less than the amplitude of the scatter. Unfortunately, schemes for statistical analysis for data including sporadic but meaningful abrupt changes in level do not yet exist, and would be difficult and time-consuming to develop. Consequently, without established techniques, we prefer to keep statistics to a minimum in order to present the data without possibly incorrect or incomplete statistical specifications.

4) Earthquake magnitudes referred to by “M” are magnitudes reported by the SIL seismic network in Iceland, which are approximately equivalent to the body-wave magnitude $mb$. 

Techniques for measuring shear-wave splitting
The most important parameters characterising shear-wave splitting are the polarisation direction of the first (faster) split shear-wave (projected onto the horizontal plane), and the time-delay between the two split shear-waves. In the Earth's crust, where azimuthally aligned shear-wave splitting is typically caused by nearly-vertical nearly-parallel EDA-cracks, the polarisations give some estimate of the strike of the cracks and hence the direction of maximum horizontal stress. The (normalised) time-delays give some measure of the magnitude of the stress, or rather the effects of the stress on the geometry of the crack distributions, along the specific ray-path directions.

These parameters can be measured visually. Attempts have been made to make measurements automatically by computer algorithms but these are successful only in particular conditions, which are not found in Iceland.

Visual (eye-ball) measurements
The procedures used for visual measuring of shear-wave splitting are described by Chen et al. (1987) and Liu et al. (1997), and see also Peacock et al. (1988). Fig. 2 shows the three plots (i), (ii), and (iii), required to accurately measure polarisations and time-delays. These are typical examples of shear-wave splitting from each of the four stations BJA, SAU, KRI, and GRI. Figs 1(i) show three-component seismograms: vertical, and two horizontal components, rotated into radial and transverse orientations. Within the shear-wave window, the horizontal component seismograms contain most of the shear-wave energy and most of the information (Crampin 1985). The vertical component seismograms are used to distinguish between the shear-wave motion and the possible $P$-to-$S$ or $S$-to-$P$ converted waves, which are polarised vertically ($SV$-waves), and nearly radial-horizontally, respectively (Crampin 1990a). Note that care must be taken to avoid misinterpreting phases converted at local topographic irregularities, which can produce anomalous signals and anomalous measurements (Crampin 1990a). Clear and impulsive onsets, typical of shear-waves recorded within the shear-wave window in Iceland, usually make the identification of the first shear-wave arrival relatively unambiguous. Occasionally (one case in ten, say), signal-to-noise ratios are low, so that no clear shear-wave onset can be identified. In such cases, the record is discarded.

The polarisation direction of the first arrival is determined from the corresponding polarisation diagram in Fig. 2(ii). The first motion of the shear-wave is usually sufficiently linear for the polarisation direction to be identified on the horizontal “LRTA” polarisation
diagram (Left, Right, Towards, and Away from the source), where the initial polarisation vector is marked by an arrow. When narrowly elliptical motion is observed, the average polarisation is chosen. The comparatively few cases without abrupt onsets, usually because of poor signal-to-noise ratios, are discarded. Weights or qualities (1, 2, 3) are assigned to the measurements according to their perceived reliability (1: good; 2: reasonable; 3: poor).

The time-delay is determined from the onset of the second arrival. This can be identified as an abrupt change of direction in the polarisation diagrams, and the time-delay measured from the samples (ticks) in the diagram. However, the preferred technique is the Ando rotation (Ando et al. 1980), where horizontal seismograms are rotated into components which are parallel and perpendicular to the direction of polarisation of the first (faster) shear-wave polarisation (Fig. 1(iii)). This allows time-delays to be measured directly from the difference in arrival times of the two orthogonal shear-wave traces. Identification of the second arrival is not always easy as second split shear-waves may be obscured by both signal-generated coda following the $P$-wave and by background noise. When in any doubt, a weight 3 is assigned to ambiguous seismograms and the measurement is discarded.

Shear-wave splitting is a robust and pervasive phenomenon but the detailed waveforms are subtle and complicated. The exact appearance of split shear-wave arrivals depends among others on frequency and amplitude of the arrival, source polarisation, angle of emergence, azimuth of arrival, attenuation, structure along the ray path, $P$-wave coda, proximity to shear-wave singularities, as well as the polarisations and time-delays of the split shear-waves themselves. As Fig. 2 indicates, it is not always simple to measure shear-wave splitting, and subjectivity is difficult to avoid.

Some 12 years ago, a test of repeatability of subjective visual techniques was made at the British Geological Survey. Five seismologists, some without familiarity with shear-wave splitting, independently measured azimuthal directions and time-delays on 20 plots of waveforms and polarisation diagrams from a variety of earthquakes, following the prescription of Chen et al. (1987). Some 80% of the readings were identical (within $\pm 10^\circ$ of azimuth, and within one or two samples equivalent to $\pm 0.02s$). The remaining 20% were ambiguous, and would have been discarded in monitoring exercises. The conclusion was that visual techniques did provide sufficient objectivity and sufficient repeatability.

Although the above guidelines for visual measurement confer consistency so that different observers obtain closely similar measurements, it would clearly be beneficial if automatic-reading techniques could be used. The main problem of automatic techniques is
that although they aim to be objective, none to date give reliable measurements for shear-wave splitting along crustal ray paths (even for the comparatively classic examples of shear-wave splitting in Iceland). In addition, when they fail for crustal shear-waves, they typically fail spectacularly and yield gross errors which are not always identified (see Crampin et al. 1991), whereas visual techniques are usually free of gross errors. Consequently, automatic techniques require stringent visual checking, to confirm they are a true image of the seismograms. Without reliable automatic techniques, careful visual checking is still required, and we consider it more direct to avoid having to assess the reliability of automatic techniques and only use visual techniques.

**Automatic measurements**

The waveforms of shear-wave splitting in the upper mantle are much simpler than those in the crust but because of longer wavelengths and lower-frequency signals, upper mantle shear-wave splitting frequently has elliptical polarisations which are difficult to measure visually. Consequently, over the years a number of automatic techniques have been developed for successfully measuring shear-waves in the mantle (Bowman & Ando 1987; Silver & Chan 1991; amongst many others). Splitting in the crust, however, has waveforms complicated by multiple arrivals and surface interactions and, although a number of automatic techniques have been suggested over the years (Shih et al. 1989; Aster et al. 1990, 1991), none have been wholly successful. Classic examples of shear-wave splitting with impulsive shear-wave arrivals, with good signal to noise ratios, and with significant time-delays are comparatively easy to measure both visually and automatically. Such classic arrivals are usually rare but are comparatively common in Iceland, which generally has good recordings of shear-wave splitting. The principal difficulty in both visual and automatic techniques is that the behaviour of shear-wave splitting in the crust varies widely from region to region, station to station, and between different ray paths within the shear-wave window at the same station (see causes of differences in previous section). (There may also be short-term temporal instabilities, discussed below, so that observations from identical source and receivers may show varying time-delays.) The visual observer can make appropriate adjustments but, to our knowledge, no automatic technique has sufficient flexibility to have wide application.

It is comparatively straightforward to design automatic techniques that will reliably read polarisations and time-delays of classic shear-wave splitting at a particular station with
a particular source-receiver geometry. The temptation is to apply these techniques outside their region of competence, when they may be seriously inadequate.

To our knowledge, the most successful automatic measurements have been made by the cross-correlation function technique (CCF) of Yuan Gao when applied to an isolated swarm of small earthquakes on Hainan Island, Southern China (Gao et al. 1998). (A reviewer points out that this technique is similar to that of Silver & Chan 1991.) Their CCF algorithm calculates the cross correlation function between two rotated horizontal shear-wave components. For each angle of rotation a cross-correlation coefficient (CCC) is evaluated, with values between 0 and 1. The polarisations of faster shear-wave and the time-delays between fast and slow waves can be estimated at the largest values of the CCC. The technique obtained reliable variations of time-delays with a much reduced scatter at two stations within the shear-wave window which showed the changes in behaviour expected in Band-I before a $M_L$ 3.7 earthquake (Gao et al. 1998).

Since it produced the most consistent results yet obtained with automatic techniques for shear-waves with ray paths in the crust, we applied the CCF technique of Gao et al. (1998) to a subset of the Icelandic data at Station BJA. Fig. 3 shows time-delays processed visually contrasted with those processed with the automatic CCF. Data are from 1st January to 10th November 1998 during which time there were two $M \sim 5$ earthquakes near the station. The CCF technique was applied to all earthquakes in this interval that had good visual estimates of shear-wave splitting. Only readings with values of CCC larger than (an arbitrary) 0.6 were retained. Only 21 of the 37 visually processed events had CCC greater than 0.6 (including two with zero time-delay erroneously indicating no splitting). Such automatic techniques are, not surprisingly, much less flexible and less adaptable than visual techniques. The time and magnitude of the November, 1998, event was successfully stress-forecast from visual data (Paper 2, and Crampin et al. 1999a). This forecast was made before the foreshock and aftershock data had been processed and made available over the Internet. Stress-forecasting would not have been possible with the omission of so many events by the automatic technique.

The success of the CCF technique applied to the Hainan Island data (Gao et al. 1998) in contrast to the much less successful application to Icelandic data in Fig. 2, is almost certainly because the Hainan Island data were from an isolated swarm (Crampin 1993a) with a small hypocentral diameter. This meant that there was only a limited range of ray paths to each of the two stations within the shear-wave window, so that the CCF technique was repeatedly accessing similar waveforms. This is close to the ideal time-lapse
configuration where comparatively similar source and receiver geometry (Crampin 1990b) produce comparatively consistent time-delays.

Clearly automatic techniques for estimating shear-wave splitting parameters from earthquake arrivals in the crust are desirable but difficult to design. Comparatively simple deterministic techniques are unlikely to be sufficient, except in exceptional circumstances, as in the isolated swarms on Hainan Island (Gao et al. 1998). Combinations of visual and automated techniques, such as those used by Menke et al. (1994), are still largely subjective. Widely applicable wholly automated programs would require considerable sophistication, and wholly successful techniques have not yet been developed. As a consequence, shear-wave splitting in Iceland was measured using the visual techniques of the previous section.

Probably the most effective automated analysis would use Artificial Neural Network (ANN) technology. Such techniques have been applied to exploration data (Dai and MacBeth 1994) with some success. However, the difficulty of selecting appropriate ANN training sets would still make application to shear-waves from earthquake signals complicated and would preserve whatever subjectivity was in the training sets.

Shear-wave splitting in Iceland (1996 - 1999)
The persistent seismicity of much of Iceland, and the seismic network and processing system developed for the SIL seismic network in Iceland, particularly in SW Iceland, satisfy the restrictive geometrical conditions required for monitoring shear-wave splitting before earthquakes (Crampin 1991a). The northern leg of the transform zone in SW Iceland, the Tjörnes fracture zone, is mostly offshore but is also of interest, as a number of destructive earthquakes \( (Ms > 6.5) \) have occurred there in the past. Most of the seismic stations are located in the vicinity of these two transform zones, with the major activity in SW Iceland. The map in Fig. 4 shows the SIL network during 1996 to 1999 with rose diagrams of shear-wave polarisations, read by visual techniques, superimposed on roundels of equal-area polar maps of the polarisations of the faster split shear-waves.

Many of the rose diagrams in Fig. 4, particularly BJA, HAU, SAN, SOL, and SAU, are exceptionally linear, reflecting the classic examples of shear-wave splitting common in Iceland (Fig. 2). Other rose diagrams show some scatter (discussed in the next sections) but there is a predominantly NE-SW alignment in SW Iceland and a NNE-SSW alignment in North-Central Iceland, indicating marginally different directions of the dominant compressional stress in southern and northern transform zones of the Mid-Atlantic Ridge in
Iceland. The NE-SW alignment in SW Iceland was previously identified in SW Iceland from observations of shear-wave splitting by Menke et al. (1994), who showed that, as is typically the case, the polarisations of the faster shear-waves were parallel to numerous dykes, fissures, tensional cracks, and micro-earthquake focal mechanisms.

Note that the polarisation directions of stress-aligned shear-wave splitting are exceptionally sensitive to the shear-wave window (Booth & Crampin 1985). Generally the stations showing largest scatter of polarisations in Fig. 4 are those with severe local surface- or subsurface-topography where the azimuth and angle of incidence of arrivals distorts the effective shear-wave window.

**Examples of shear-wave splitting**

In order to demonstrate some of the range of variations in shear-wave splitting time-delays, we plot in Fig. 5 horizontal seismograms rotated into the preferred fast and slow polarisation directions for two suites of earthquakes. We select two sequences of typical seismograms illustrating particular phenomena.

Fig. 5a shows the sequence of increasing (normalised) time-delays from 12th July, 1998, to 18th October, 1998, in Band-1 of the shear-wave window at Station BJA. These are plotted in Fig. 1a of Paper 2, and were the sequence which allowed the M5 earthquake to be successfully stress-forecast (Crampin et al. 1999a). The time-delays for this increase shows less scatter about the mean value than is typical for other increases.

Vertical “stripes” of time-delays, where a wide range of time-delays occurs in a very short space of time, are common in plots of time-delay variations (see Fig. 1 of Paper 2). They are particularly common in aftershock sequences. Typically such arrivals are from earthquakes in tight source volumes of limited dimensions. Fig. 5b shows time-delays from such a sequence of earthquakes from 15th to 17th March, 1999, observed at Station SAU are plotted in Band-2 of the shear-wave window in Fig. 1b of Paper 2. Only the first earthquake in the stripe, (a) in Table 1c, is from outside a 1km diameter source volume.

The time-delays in Fig. 5b (both non-normalised in seismograms and normalised in equal-area plots) show an exceptionally large scatter from 2 to 16ms/km (±80%). It is difficult to explain this by any conventional geophysical mechanism (see section on scatter, below). Note that some seismograms from the localised source volume in Fig. 5b show strong similarities but there are no exactly similar seismograms. In general, pairs of similar seismograms are rare in seismograms from Iceland.
Shear-waves are very sensitive to surface interactions, and stations that show the most irregular scatter in polarisations in Fig. 4 are typically locations on steep slopes and/or close to irregular surface or sub-surface topography (Crampin 1990a; Liu et al. 1997). Some sources of scatter in polarisations can be identified: GRI is on a steep sided island; KRO is in the onshore rift of the Mid-Atlantic Ridge; and KRI has an irregular cliff-face less than a wavelength from the station. Small earthquakes are typically the result of some stress or tectonic anomaly which is likely to disturb measurements of shear-wave splitting, so the remarkable linearity of polarisations implying comparatively uniform microcrack- and stress-alignments at many of the stations is surprising.

The polarisations in Fig. 4 are from arrivals within the straight-line shear-wave window defined by $\tan^{-1}(\text{epicentral-distance/focal-depth}) = 45^\circ$. The $45^\circ$ limit is to allow for ray curvature due to low-velocity near-surface layers bending arrivals within the theoretical $35.26^\circ$ shear-wave window for a Poisson's ratio of 0.25. The occasional orthogonal polarisations are expected when the source mechanism radiates shear-waves in the direction of the station which happen to be polarised parallel to the slower split shear-wave so the faster wave is not excited (Crampin et al. 1986). Rose diagrams of shear-wave polarisations elsewhere typically display a scatter of at least $\pm 20^\circ$ regardless of whether the shear-waves are read automatically or visually.

Note that no corrections are made for the free-surface interactions. All observations are within the effective shear-wave window, where the effects are minimal, and in any case free-surface interactions would only marginally affect the rose diagrams in the roundels. Free-surface corrections do not affect time-delays.

Note that Menke et al. (1994), who previously examined shear-wave splitting in SW Iceland, generally found a larger scatter of polarisations at the stations common to Fig. 4: HEI and SOL; and especially BJA, HAU and SAU which are particularly linear in Fig. 4. This is probably because Menke et al. used arrivals from earthquakes with epicentral distances up to 60km from stations, which the assumed crustal structure suggested would still arrive within the shear-wave window. The scatter in Menke et al. (1994) suggests that the arrivals are, either outside the effective shear-wave window, or that complications along the long ray paths disturbed the wave trains. As such long ray paths would greatly complicate time-delays between the split shear-waves, which we interpret as indicating
stress changes in Paper 2, we restrict measurements to arrivals within the straight-line shear-wave windows of Fig. 4.

The scatter of time-delays
Since one of the most puzzling aspects of shear-wave splitting is the scatter of time-delays, we discuss this scatter in some detail. Measured time-delays in Figs 3b and 5b, and in Fig. 1 of Paper 2 show a large scatter about the nine-point moving averages and about the least-squares lines. The degree of scatter, sometimes as large as ±80%, is similar to that observed whenever time-delays above earthquakes are measured (Crampin 1999). Thus a large scatter appears to be an inherent feature of measurements of time-delays above small earthquakes. The only exceptions, where measurements of shear-wave splitting show considerably less scatter, are observations above isolated swarms of small earthquakes as in Arkansas (Booth et al. 1990) and Hainan Island (Gao et al. 1998.), where the shear-waves travel along a restricted range of ray paths from very tight clusters of foci.

Possible sources of scatter in a conventional (non-critical) crust
There are at least five possible sources of scatter of time-delays in a conventional (non-critical) crust where the usual deterministic geophysics applies.

1) Possible source: Variations in time-delays inherent to geometry of time-delays along ray paths in Band-1 assuming distributions of nearly-vertical nearly-parallel cracks.
Line-singularities (Crampin & Yedlin 1981) cross Band-1 directions in shear-wave windows in parallel vertical cracks. Consequently, time-delays in the solid angle of directions in Band-1 may be expected to vary from zero to nearly the maximum value (for example, see Crampin et al. 1990; Crampin 1993b). This means that time-delays along any arbitrary choice of directions in Band-1 will naturally have a wide variation. (We have found that attempting to correct for variations in time-delays for different directions within Band-1, assuming parallel vertical cracks, does not significantly reduce the scatter. Since this makes estimated values dependent on structure and earthquake locations without improving scatter, we have not made such corrections.)

Arguments against: In contrast, under the same assumptions, the theoretical time-delays in Band-2 should vary very little. Since in general, the scatter of time-delays in Band-2 is observed to be very similar to those in Band-1 (see Fig. 5b), a similar source of scatter may be expected. Consequently, although variations in Band-1 are expected,
the wide variation of theoretical values is unlikely to be the major source of the scatter. The ±80% range in Fig. 5b is in Band-2 directions where there are no shear-wave singularities.

2) Possible errors in location of earthquake foci provide two sources of scatter:

2a) Possible source: Ray path directions in Band-1 of the shear-wave window through distributions of nearly-vertical nearly-parallel cracks are expected to include directions of shear-wave line-singularities where shear-wave splitting is irregular and time-delays may be zero. This means that time-delays in Band-1 may be expected to vary widely for small changes in ray path directions (Crampin 1991b; Crampin et al. 1990).

Arguments against: Although variations within the shear-wave window can certainly cause scatter, there are several reasons why this is thought not to be the major source of scattering:

i) Time-delays in Band-2 are not expected to include singularities from geometrical considerations, and yet the scatter in Band-2 as in Fig. 5b is similar to that in Band-1, where there is a line-singularity.

ii) All four stations, but particularly KRI and GRI, include local activity where clustering is common: many events within the shear-wave window within a short period of time. Such repeated seismicity is likely to have similar foci, similar ray paths, and consequently similar time-delays. However, these "stripes" of time-delays (Fig. 5b, and Fig. 1 of Paper 2) typically show a scatter as large as, and in some cases larger than, at other times for the particular station (Fig. 5b).

iii) Reynir Böðvarsson relocated a selection of events within the shear-wave window with a multi-event relative-location technique (Slunga et al. 1995). This improved locations, as was shown by increased spatial clustering of foci, more planar locations, and reduction in location errors, but the scatter in time-delays was not significantly reduced.

2b) Possible source: Time-delays are normalised by hypocentre path length to ms/km so that time-delays along different paths can be compared, and if the measured path lengths are in error, the normalised time-delays will also be in error.

Arguments against: Possible errors in location are usually much too small (see error bars in time-delay plots in Fig. 3 and in Fig. 1 of Paper 2, and the tight focal zone in Fig. 5b) to cause the ±80% scatter in time-delays.

3) Possible source: Errors in picking and measuring time-delays.
Arguments against: Most measurements are checked by more than one analyst and any shear-wave measurements that appear in any way doubtful are rejected. Consequently, it is unlikely that errors large enough to cause the ±80% scatter are common.

4) Possible source: Complicated rock structure beneath the station causing variations in velocity structure, so that small differences in source location result in different ray paths with different degrees of anisotropy.

Arguments against: Although clearly a possible source of scatter, it would not be expected to cause such significant scattering on almost all occasions at all stations, particularly along ray paths from the comparatively tight clusters of foci in the stripes of KRI and GRI in Fig. 1 of Paper 2, and at SAU in Fig. 5b.

5) Possible source: Complicated stress-aligned EDA-crack structure beneath the station.

Arguments against: There are now a large number of measurements of shear-wave splitting in controlled source exploration experiments in mostly sedimentary rocks (Crampin 1994, 1996; Winterstein 1996). Controlled-source experiments typically allow detailed high-precision measurements, particularly in VSP experiments (Li & Crampin 1991a, 1991b; Yardley & Crampin 1993). These measurements typically show regular well-ordered splitting with no sign of the ±80% scatter in time-delays above small earthquakes. It should be noted that exploration experiments are typically in sedimentary strata, whereas ray paths above small earthquakes are usually, but not always, in igneous and metamorphic rocks. However, in view of the observed similarities in shear-wave splitting in all types of rock (Crampin 1994, 1996), it seems unlikely that the scattering can be wholly attributed to igneous and metamorphic rocks.

The five possible sources of the scatter, listed above, have each been rejected as the primary source. It is possible that each source contributes so that the total scatter is a combination of the effects of several sources of scatter, but in view of the consistency of exploration experiments in Item 5, above, this does not seem likely. Remarkably, the variations before earthquakes and volcanic disturbances appear reasonably well established despite this large scatter. This suggests that the scatter is in some sense a random fluctuation about the nine-point moving average and is independent of the stress-induced mean-value modifications that we report in Paper 2. This is confirmed by the seismograms in Fig. 5b where an ±80% scatter of time-delays is observed along similar ray paths in a short period of time when there is no known major change of conditions.
Possible source of scatter in a critical crust

The anisotropic poro-elasticity (APE) model, which successfully matches so much of the behaviour of shear-waves and cracks in the crust (Crampin 1999, 2000), is a mean field theory (in the sense of Jensen 1998). APE assumes that the cracks in the crust are a critical system held close to (fracture) criticality (Crampin 1998; Crampin & Chastin 2001). Mean field theories average over the clustering of heterogeneities at all scale lengths and all dimensions inherent in the self-similar, fractal, or power-law distributions of many phenomena associated with cracks and earthquakes (Crampin 1999), including the well-known Gutenberg-Richter relationship between numbers and magnitudes of earthquakes.

APE-modelling suggests and both field- (Crampin 1999) and laboratory-observations (Crampin et al. 1999b) confirm that stress-aligned fluid-saturated cracks are sensitive to even minor modifications of stress, pressure, and other conditions. The driving mechanism for deformation is fluid migration by flow or diffusion along pressure-gradients between neighbouring grain-boundary cracks and low aspect-ratio pores at different orientations to the stress-field (Zatsepin & Crampin 1997; Crampin & Zatsepin 1997; Crampin 1999). These distributions necessarily contain clusters of cracks with different numbers of cracks, different aspect-ratios, and different crack-densities. This means that although the averages may be stable (witness the wide range of phenomena, where APE-modelling appears to be a good, or at least satisfactory, model of observations, Crampin 1999), any individual observation may deviate substantially from the mean.

Note that the very wide scatter often occurs in comparatively short periods of time, in foreshock and aftershock sequences and short-lived series of small earthquakes, at all stations (Fig. 5b, and Fig. 1 of Paper 2). This suggest that the mechanism that drives the phenomena responds very quickly to changes, with time constants of perhaps at most a few hours. The time constants cannot be too small, much less than an hour, say, or the effects would disturb the smooth variations in measurements of shear-wave splitting in reflection profiles and VSPs as in Li and Crampin (1991a, 1991b) and Yardley & Crampin (1993), and would have been recognised previously.

Thus, although in principle, some scatter is expected in a critical crust, and better understanding of the mechanism that modifies the distributions is required. One of the first aims of the proposed controlled-source stress-monitoring site, discussed in Paper 2, will be to examine the stability of shear-wave splitting and investigate the cause of the scatter.

We can speculate that the driving mechanism for this scattering is the stress fluctuations introduced by Earth and Ocean Tides. Tatham et al. (1993) found approximately 0.5%
variations in both horizontally and vertically propagating $P$-wave velocities correlating with, particularly the diurnal, Earth Tides in transmission and reflection experiments in Illinois, USA. Tatham et al. attributed these variations to "the opening and closing of cracks, pores and/or pore throats". Since the sensitivity of shear-wave splitting to comparatively minor modification of crack distributions is much greater than that of $P$-waves, we can expect these effects to cause significant variations in time-delays in shear-wave splitting. The sensitivity of fluids in the crust to Earth Tides is well-known, for example, Gupta et al. (2000) observe tidal signals in water-level fluctuations in wells around the Koyna Dam in India. Kümpel (1997) reviews water-level changes in wells, and reports distinctive fluctuations of water level correlating both with diurnal tides and with barometric pressures. Stresses from such phenomena produce changes in water levels by modifying the aspect-ratios of water-filled cracks in the surrounding rock, so that water is squeezed in and out of the rock mass. This is exactly the process that is monitored by shear-wave splitting and is expected to be one of the causes of the scatter in shear-wave time-delays.

Conclusions
Shear-wave splitting in Iceland, particularly above the persistent seismicity of the onshore transform zone of the Mid-Atlantic Ridge in SW Iceland has been monitored for the four years. The seismicity and seismic network of Iceland seems exceptionally favourable for measuring shear-wave splitting having high seismicity within the shear-wave windows of a number of SIL stations, and good recording and location techniques (particularly of small earthquakes), as well as seismic records being available on the Internet. All shear-wave arrivals within the shear-wave window, with very few exceptions, display shear-wave splitting. The shear-wave splitting is consistent with propagation through distributions of nearly-vertical nearly-parallel fluid-saturated cracks striking approximately parallel to the direction of maximum horizontal stress. The shear-wave polarisations, and inferred directions of maximum compressional stress, in the southern half of Iceland average about NE-SW, whereas the polarisations and stress directions in northern Iceland average about NNE-SSW.

The suggested importance of these studies of microcrack deformation monitored by shear-wave splitting is that in principle any modification in crack geometry can be directly interpreted in terms of changes in direction and magnitude of stress. We suggest this opens a new window for examining and interpreting stress in \textit{in situ} rock.
As suggested in the Forewarning at the beginning of this paper one of the major problems raised by this paper and Paper 2, and by almost all measurements of shear-wave splitting elsewhere (reviewed by Crampin 1999), is the cause of the large scatter in measured time-delays between split shear-waves. They appear to be too large to be caused by conventional (non-critical) geophysics. The dense distributions of cracks suggested by the levels of recorded shear-wave splitting strongly suggest that cracks are a critical system verging on (fracture) criticality. This suggests that the large scatter is due to wave propagation through the heterogeneities and clustering in time and space inherent in critical systems (Bruce & Wallace 1989) of cracks on the verge of criticality. One of the major aims of the controlled-source stress-monitoring site currently being developed in northern Iceland will be to investigate the source of the scatter.

This work was partially supported by European Commission PRENLAB Projects, Contracts ENV4-CT96-0252 and ENV4-CT97-0536, and SMSITES Project, Contract EVR1-CT1999-40002. We thank Ragnar Stefánsson, of the Iceland Meteorological Office, without whose collaboration this investigation would not have been possible. We also thank: Reynir Böðvarsson, of Uppsala University, who recalculated earthquake locations; Yuan Gao, of the Chinese Seismological Bureau, for the use of his CCF technique for automatic measurement of shear-wave splitting; and many colleagues at home and abroad, too numerous to mention individually, who have each contributed to the arguments in one way or another.

References


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——— 1996. Anisotropists Digest 149 and 150. anisotropists@sep.stanford.edu.


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Table 1: Earthquakes in Figures

1.1 Earthquakes for seismograms in Fig. 2

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<tr>
<th>Station</th>
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<th>Time (hr min sec)</th>
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<th>Mag. (M)</th>
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1.2 Earthquakes for seismograms in Fig. 5a

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1.3 Earthquakes for seismograms in Fig. 5b

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FIGURE CAPTIONS

**Fig. 1.** Geometry of Band-1 and Band-2 in the shear-wave window in a medium with distributions of parallel vertical cracks. The circle is a horizontal slice of the vertical cone within which source events are in the effective shear-wave window of a seismic station on a horizontal free-surface, where there are typical low-velocity surface layers.

**Fig. 2.** Typical examples of seismograms of local earthquakes recorded within the shear-wave window for the earthquakes listed in Table 1.1. The sampling rate is 100 samples per second. At each station: (i) seismograms are rotated into vertical, horizontal radial, and horizontal transverse directions, with numbered 0.1s-intervals for polarisation diagrams of the shear-wave motion. (ii) Shows mutually-orthogonal polarisation diagrams, where: U, D, L, R, T, and A refer to Up, Down, and Left, Right, Towards, and Away directions from the source. The number top left is the polarisation interval in (i), above; top right is the amplitude factor (number of relative multiplications of the traces), and the arrowhead marks the arrival, and the arrow marks the horizontal vector polarisation, of the leading split shear-wave. (iii) Shows horizontal seismograms rotated into polarisation directions from (ii) of the faster and slower split shear-waves, where the time-delay is the difference in arrival times in ms.

**Fig. 3.** A comparison of visual and automatic techniques for measuring time-delays. Time-delays, normalised by path length, from 1st January to 13th November, 1998, in Band-1 of the shear-wave window at station BJA. Time-delays are obtained by, at top, the automatic CCF technique for cross-correlation coefficients (CCC) greater than 0.6 and, at bottom, the visual techniques outlined in the text. Vertical lines through the normalised time-delay points are error bars (observed time-delays in ms divided by the ends of the range of location errors). Dashed lines are nine-point moving averages. Straight lines through the visual measurements are least-square estimates, beginning just before a minimum of nine point average and ending at a large earthquake. Arrows with magnitudes and epicentral distances from BJA indicated mark the times of $M=5$ earthquakes.

(cont.)
Fig. 4. Map of Iceland, showing the seismic network and analyses of shear-wave splitting. Triangles are SIL seismic stations, several of which were installed during the four years 1996 to 1999. Shaded areas are ice fields of which the largest is Vatnajökull. Open circles mark the active volcanoes of, from north to south, Bárðarbunga, Grímsvötn, Hekla, and Katla. The roundels are equal-area polar plots of polarisations of the faster split shear-wave arrivals in the shear-wave window (out to 45°) during the four years, containing rose diagrams (with 10° petals) of these polarisations. Roundels are shown only for those named stations where there are more than 10 arrivals within the shear-wave window. The named stations without roundels (ASM, KRO, MID) have sufficient arrivals but the wave-forms and polarisations are severely disturbed by local rifting and/or local surface or sub-surface topography.

Fig. 5. Examples of shear-wave splitting in sequences of seismograms. Upper diagrams are horizontal seismograms rotated into fast and slow polarisation directions with arrivals marked with vertical bar. Roundels are equal-area polar plots out to incident angles of 45°: (left) bars of fast shear-wave polarisations; and (right) circles of shear-wave time-delays with diameters proportional to delay. (a) Data from nine earthquakes listed in Table 1.2 showing increasing time-delays. (b) Data from nine earthquakes listed in Table 1.3 showing variation in normalised time-delays for earthquakes in a "stripe" of time-delays, eight of which have nearly identical foci.
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Fig. 1
Figure 2 a,b
Figure 2 c, d
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Figure 3
Shear-wave splitting in Iceland: 1

Figure 5a
Figure 5 b