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Shear-wave splitting measured for permanent reservoir monitoring systems: an example from the Snorre field

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Friday 5th September, 2025

Abstract

Microseismic monitoring of offshore CO_2 storage projects is likely to include some deployment of offshore sensors. To improve the value proposition of this monitoring infrastructure, it is important to consider what other information can be gained about the CO_2 storage complex and the surrounding region. Shear-wave splitting is one potential source of added value to microseismic monitoring of CO_2 storage operations at minimal additional cost, if factored in during network design. Shear-wave splitting provides a means to passively monitor the *in situ* horizontal maximum stress azimuth and potentially the magnitude of differential horizontal stresses. We demonstrate this for offshore monitoring of reservoirs using data recorded by the permanent reservoir monitoring (PRM) network at the Snorre field. We measure shear-wave splitting for the M_W 5.1 Tampen Spur earthquake and subsequent microseismic aftershocks. Our results show that high-quality shear-wave splitting measurements can be made for microseismicity, with $M_L \geq 0.7$, recorded by seafloor instruments. At Snorre, the average shear-wave splitting fast polarisation direction $\phi_f = 92 \pm 15^\circ$ and percentage anisotropy $\xi = 2.68 \pm 0.26$. This is consistent with microcracks preferentially aligned with the maximum horizontal stress azimuth. At Snorre we estimate this as $108 \pm 4^\circ$ using data from the World Stress Map. The shear-wave splitting results contain two groups of fast polarisation directions. The four westernmost stations cluster around $\phi_f = 68 \pm 13^\circ$ with the remaining clustering around $\phi_f = 113 \pm 4^\circ$. This variation may be due to the depletion history of the reservoir. Incorporating shear-wave splitting into microseismic monitoring plans potentially allows for semi-continuous measurements of the changes to the stress field in the storage complex and surrounding region, provided there is sufficient microseismicity. This demonstrates that shear-wave splitting is a valuable dataset for monitoring the offshore subsurface stress field, which should be considered when planning offshore passive seismic monitoring.

1 Introduction

Geological CO_2 storage is an essential part of global net zero strategies. Many countries including the UK, Norway, Denmark, and the Netherlands are developing offshore geological storage projects in the North Sea, as the geology is favourable for geological storage and there is existing infrastructure and technical expertise which can be redeployed (Furre et al., 2019; Skurtveit et al., 2022). With the growth in project development there is an increasing demand for new approaches to characterise the *in situ* stress state of prospective sites, and to monitor the geomechanical response of the storage complex to CO_2 injection. This is needed to ensure the safety and operability of CO_2 storage sites (Skurtveit et al., 2022).

One important component of monitoring geological carbon storage projects is seismicity. Fluid injection has been associated with seismicity in a wide range of geological settings (e.g., Keranen and Weingarten, 2018) including CO_2 injection (e.g., Stork et al., 2015; Harvey et al., 2021; Bauer et al., 2022) and offshore gas storage (Cesca et al., 2014). For the North Sea, a high quality seismicity catalogue has been compiled (Kettlety et al., 2024), enabling improved assessment of faulting, the background stress field, leakage risk, and seismic hazard for projects (Kühn et al., 2025). One recommendation from this body of work is that offshore seismic monitoring infrastructure is needed if seismicity near CO_2 storage projects is to be monitored with sufficient accuracy. Offshore (i.e., near-source) observations are vital in calculating accurate earthquake locations (particularly depths), focal mechanisms, and magnitudes.

Achieving a high quality catalogue of seismicity is the primary product for any passive seismic monitoring programme, to better understand how the reservoir is responding to injection. However, deploying networks of seismometers on the seafloor has the potential to generate highly valuable secondary datasets, further aiding that understanding, particularly if acquiring these secondary data products is incorporated into the network design. One such dataset is shear-wave splitting, which is an indicator of seismic velocity anisotropy – the variation in seismic velocity with propagation direction – measured using microseismicity. This can provide a measure of the *in situ* stress field. Fracture-induced seismic anisotropy has previously been observed using 4 component sea bottom cable systems at the Valhall field using P wave amplitude variation with offset and azimuth (AVOA, Hall and Kendall, 2003) and seismic interferometry (Mordret et al., 2013). Such seafloor instrumentation has been previously used to monitor induced seismicity (Chambers et al., 2010)

1.1 Shear-wave splitting

Shear-wave splitting, or seismic birefringence, occurs when a shear-wave propagates through an anisotropic medium. The incident shear-wave is split into two sub-perpendicular shear-waves which propagate through the medium at different velocities (Figure 1). The polarisation of the fast shear-wave (ϕ) and the delay time between the split shear waves (δt) is measured where ϕ is related to the orientation of the symmetry axes of the anisotropic medium and δt to the strength of the anisotropy. Shear-wave splitting is typically measured using passive seismic data and has been measured in many industrial settings using microseismic data (e.g., Al-Harrasi et al., 2011b; Stork et al., 2015; Hudson et al., 2024, etc.,) including for offshore fields using data from borehole geophones (Valhall; Teanby et al., 2004a) and ocean bottom instruments (Ekofisk; Jones et al., 2014).

In the Earth there are many potential mechanisms for seismic anisotropy. At the reservoir scale and depth the predominant mechanism is preferential alignment of near-vertical micro-scale fractures with the azimuth of maximum horizontal stress (S_{Hmax}; Nur and Simmons, 1969; Crampin, 1987; Zatsepin and Crampin, 1997). This mechanism produces a hexagonal anisotropy with a horizontal symmetry axis, known as horizontal transverse isotropy (HTI). Sedimentary strata can also develop anisotropy either through periodic layering of different units (e.g., Backus, 1962), or preferred alignment of anisotropic minerals such as phylosillicates (Kendall et al., 2007). These mechanisms, however, produce a hexagonal anisotropy with a vertical symmetry axis or vertical transverse isotropy (VTI). In passive seismic studies where receivers are at the surface, the incident shear-waves are near vertical and, therefore, are not sensitive to VTI. If data from borehole geophones are used, where ray-paths are propagating horizontal from sources to receivers, then this contribution from sedimentary fabrics, and the fact that the combination of VTI and HTI mechanisms produces an anisotropy with an orthorhombic symmetry, must be considered but can provide additional reservoir information (Baird et al., 2013). Where the observed seismic anisotropy is due to the stress-induced alignment of micro-scale fractures, this can be used to gain information on the *in-situ* stress field, particularly the orientation of maximum horizontal stress (S_{Hmax}). S_{Hmax} is generally parallel to fracture strike and, therefore, to the shear-wave splitting fast polarisation direction. This allows for the orientation of S_{Hmax} to be interpreted from passive seismic shear-wave splitting datasets (e.g., Savage et al., 2010; Igonin et al., 2022; Guzman et al., 2022; Hudson et al., 2024). In some cases, it has been possible to infer changes in the stress state at a reservoir using temporal variations in shear-wave splitting (e.g., Teanby et al., 2004a; Stork et al., 2015) and in tectonic (e.g., Pastori et al., 2019) and volcanic settings (e.g., Gerst and Savage, 2004; Kendall et al., 2025). The Ekofisk Microseismic experiment, where microseismic data was acquired over an 18-day period in April 1997 at the Ekofisk oil field in the North Sea, showed that shear-wave splitting could be used to illuminate spatial variations in aseismic fracture sets (Jones et al., 2014).

Whilst shear-wave splitting has been measured for microseismicity in offshore settings, such as at Valhall (Teanby et al., 2004a) and Ekofisk (Jones et al., 2014), this has relied on geophones installed in monitoring boreholes. Using sea floor instrumentation, such as ocean bottom seismometers or permanent reservoir monitoring deployments, to measure shear-wave splitting has proved challenging given the increased noise levels in the marine environment and uncertainty on sensor component orientations. Where shear-wave splitting has been measured using seafloor instruments it has often been for teleseismic shear-wave in deeper oceanic environments (e.g, Harmon et al., 2004; Collins et al., 2012; Scholz et al., 2018). Shear-wave splitting has been successfully measured for earthquakes with $M \geq 2.5$ using a deployment of 150 ocean bottom seismometers over a subsea area of 300×1000 km off the coast of northeastern Japan (S-net; Uchida et al., 2020).

The potential application to CO_2 storage is particularly exciting, since ensuring safe and reliable geological CO_2 storage requires new methods to monitor the geomechanical response of reservoirs to injection (Skurtveit et al., 2022). This is because *in situ* stress state can naturally have a significant impact on the operation and containment risk assessment of storage projects. Many potential CO_2 storage sites require drilling and operating injection wells in regions or depths that may not have had previous hydrocarbon exploration and, therefore, there may be fewer data to conduct leakage risk assessments. Additional means of constraining stress or fracturing are valuable, particularly when they are derived from independent geophysical methods, to image the reservoir, seal, and overburden units. Fracture and fault trends in particular are important inputs in containment risk assessment, as their orientations with respect to *in situ* stresses significantly affect their potential behaviour when stress changes occur as a result of injection. The likelihood of fault failure, fracture development, and other deformation are affected by stress, and thus it is a critical variable to constrain when assessing a field for CO_2 injection and monitoring operations.

Understanding the *in situ* stress state can have a significant impact on the operation and containment risk assessments of storage projects. Many potential CO_2 storage sites require drilling and operating injection wells in regions or depths that may not have had previous hydrocarbon exploration and, therefore, there may be fewer data to conduct leakage risk assessments. Additional means of constraining stress or fracturing are valuable, particularly independent geophysical methods to image the reservoir, seal, and overburden units. Fracture and fault trends in particular are important inputs in containment risk assessment, as their orientations with respect to *in situ* stresses significantly affect their potential behaviour when stress changes occur as a result of injection. The likelihood of fault failure, fracture development, and other deformation is affected by stress, and thus it is a critical variable to constrain when assessing a field for CO_2 injection and monitoring operations.

2 Data

2.1 Permanent reservoir monitoring systems

Permanent reservoir monitoring (PRM) systems, consisting of three-component geophones and hydrophones, have been deployed to monitor oil and gas fields in the northern North Sea (Thompson et al., 2015). Similar PRM systems could be an option for monitoring of offshore CO_2 storage fields, but shear-wave splitting is not routinely measured for data recorded by these systems. PRM data for three fields in the northern North Sea — Snorre, Grane and Oseberg — are good sites to test the potential for PRM systems to measure shear-wave splitting. Data from select PRM stations are shared with the Norwegian National Seismic Network (NNSN; Figure 3 Ottemöller et al., 2021). The PRM systems installed at the Snorre field is the only one found to have suitable seismicity, using the unified North Sea earthquake catalogue produced by the SHARP project (Kettlety et al., 2024; Kettlety et al., 2025). The PRM system installed at Snorre is one of the largest in the world (Thompson et al., 2015; Jerkins et al., 2024), consisting of a seismic cable containing 10,708 four component sensors, which have an eigenfrequency of 15 Hz.

Whether shear-wave splitting can be measured for an earthquake is limited by the 'shear-wave window'. Interactions with the free surface affect the particle motion of shallow incident angle shear-waves (Nuttli, 1961). To avoid these effects, shear-wave splitting is only measured where the incidence angle is less than the critical angle

(Booth and Crampin, 1985). We use a critical angle, or shear-wave window, of 45° for straight line ray paths between the source and receiver. This assumes that low velocity layers near the surface will turn incident shear-wave ray paths such that they are near-vertical at the free surface.

These events include the 21st March 2022 M_W 5.1 Tampen Spur earthquake and 26 subsequent aftershocks. The mainshock and five subsequent aftershocks are taken from the unified North Sea earthquake bulletin (Kettlety et al., 2024; Kettlety et al., 2025). This initial dataset is supplemented by additional aftershocks detected using the Snorre PRM system (Jerkins et al., 2024). The aftershocks have local magnitudes in the range $-0.6 < M_L < 2.6$. Waveform data for all earthquakes were obtained for the 10 PRM nodes shared with the NNSN and for the Tampen Spur mainshock data from an additional 50 PRM stations were provided by Equinor. An example of the data used, which includes horizontal component seismograms recorded by PRM stations within the shear-wave window of a M_L 0.7 aftershock, is shown in Figure 2.

2.2 Stress Data

Data from the 2025 release of the World Stress Map (Heidbach et al., 2025) are used to characterise the regional S_{Hmax} azimuth near the Snorre field (Figure 3). This dataset comprises 129 data points across the northern North Sea compiled from a variety of measurement types including: earthquake focal mechanisms (24 data points); borehole breakouts (58 data points); overcoring (5 data points); and drilling induced tensile fractures (24 data points). Data in the World Stress Map are assigned a data quality code based on their reliability to assess regional stress field orientation (Heidbach et al., 2016). Only data which are rated as A, B, or C on the World Stress Map data quality scheme are used, which gives a stress dataset comprising 50 measurements. The quality codes A, B, and C indicates the data has an uncertainty in S_{Hmax} orientation of ±15°, ±20° or ±25° respectively.

3 Method

If a shear-wave has only propagated through isotropic media, and has a sufficiently vertical incidence angle such that phase shifts from interactions with the free surface can be neglected (Nuttli, 1961), the displacement recorded by a single seismometer can be written as:

$$\mathbf{u}(\omega) = u(\omega)\hat{\mathbf{p}} \tag{1}$$

in the frequency domain, where $u(\omega)$ is the source wavelet in the frequency domain and $\hat{\mathbf{p}}$ is the source polarisation. In this case the expected particle motion in the horizontal components is linear. If, however, the shear-wave propagates through anisotropic media, the delay time added by shear-wave splitting results in a phase shift which produces a characteristic elliptical particle motion. Therefore, shear-wave splitting can be effectively measured by searching for a set of shear-wave splitting parameters $(\phi, \delta t)$ which restore a linear particle motion. Here a method known as eigenvalue minimisation is used to characterise particle motion linearity (Silver and Chan, 1991; Walsh et al., 2013). For a shear-wave isolated in a time window of interest, where the optimum time window is found using cluster analysis (Teanby et al., 2004b), the horizontal component seismograms are rotated from the geographic to the radial-transverse reference frame, and the covariance matrix is computed following Walsh et al. (2013). The first and second eigenvalues, λ_1 and λ_2, correspond to the energy of the radial and transverse components respectively. Using the implementation of Wuestefeld et al. (2010) we grid search over the range of plausible shear-wave splitting parameters, $-90° \leq \phi_f \leq 90°$ and $0\,\mathrm{s} \leq \delta t \leq 0.1\,\mathrm{s}$, seeking to minimise $\frac{\lambda_2}{\lambda_1}$. Error estimates in $\phi_f, \delta t$ are made using a 95% confidence region defined by

$$\lambda_2(\phi_f, \delta_t) \leq \lambda_{2_{min}} \{1 + [k/(v-k)] F^{0.05}_{k,v-k}\}, \tag{2}$$

where k is the number of parameters, in this case 2, v is the estimated degrees of freedom of the data, and $F_{k,v-k}$ is an F-distribution (Silver and Chan, 1991; Walsh et al., 2013). Standard errors in $\phi_f, \delta t$ are then estimated by taking the quarter of the length and width of the 95% uncertainty ellipsoid. Figure 4 shows an example shear-wave splitting measurement made for a $\mathrm{M_L}$ 0.7 aftershock of the Tampen Spur event. This example has a signal-to-noise ratio of 5.

The measured shear-wave splitting delay time, δt, is integrated along the entire ray path. Therefore, δt may vary with earthquake depth depending on the thickness of the anisotropic medium. To correct for ray path length effects δt is converted to percent anisotropy,

$$\xi = 100(V_S * \frac{\delta t}{d}) \ , \tag{3}$$

where V_S is an assumed mean shear-wave velocity and d is the ray path length, assuming a straight ray from source to receiver.

4 Results

Shear-wave splitting measurements are made using waveform data for the Tampen Spur mainshock and subsequent aftershocks recorded by permanent reservoir monitoring stations at Snorre. After discounting stations that are outside of the shear-wave window for the earthquakes, we are able to make 124 shear-wave splitting measurements. After data quality control there are 25 good quality measurements of shear-wave splitting, with a further 28 clear null measurements where no shear-wave splitting is observed. Figure 5 shows the 25 good quality shear-wave splitting measurements at the recording station. The level of data attrition, with approximately 20% of measurements resulting in good quality splits is lower than other studies of microseismic shear-wave splitting (e.g., Teanby et al., 2004a; Pastori et al., 2019; Asplet et al., 2024; Asplet et al., 2025). One reason for the higher rate of data attrition is that few of the PRM sensors yield usable shear-wave splitting measurements for the Tampen Spur mainshock. This is due to the larger amplitude shear-waves either not being fully recorded due to data clipping, or the energetic wave trains cause significant ringing on the sensors. Both mean clear shear-wave splitting measurements cannot be made. Figure 6 shows the S-phase recorded by the closest line of 10 PRM sensors to the mainshock. No good quality shear-wave splitting measurements can be made for these data.

To estimate the regional S_{Hmax} azimuth at Snorre, we use data from the 2025 release of the World Stress Map (Heidbach et al., 2025). However, there are few datapoints close to Snorre, with the closest being 19 km from the centre of the PRM network. The regional S_{Hmax} azimuth is estimated by taking the circular mean of the 25 datapoints that are within 50 km of the centre of the PRM network. This gives a regional S_{Hmax} of $108 \pm 4°$. This estimate of S_{Hmax} rotates to favour a near east-west S_{Hmax} azimuth as the radius of the averaging area increases (Figure 7).

The good quality shear-wave splitting measurements are concentrated to the North of the Snorre field (Figure 5). Seven measurements of shear-wave splitting are made for the Tampen Spur mainshock, with the remaining 18 made for the aftershocks. Of those, 11 are made for aftershocks with $M_L \leq 1.0$. Averaging the ϕ_f results from all the PRM sensors gives a circular mean ϕ_f of $-85 \pm 13°$ (Figure 8). When aggregated the ϕ_f measurements show a slight bimodal distribution. This arises as the measured ϕ_f varies across the stations, with stations SNO01, SNO03, SNO05 and SNO08 measuring ϕ_f approximately oriented northeast-southwest (Figure 5,9a).

The 13 ϕ_f measurements made for data recorded by SNO01, SNO03, SNO05 and SNO08 (hereafter referred to as Group 1) have a circular mean of $\bar{\phi_f} = 68 \pm 13°$ and the circular mean for the 12 ϕ_f measurements made at the remaining stations (or Group 2) $-67 \pm 4°$ (Figure 9).

Shear-wave splitting delay times are converted to ξ following Equation 3. As all the earthquakes used here have focal depths in the range of $19.5 - 26.2$ km (Jerkins et al., 2024), the measured δt are small relative to the ray path lengths with a mean ξ of 1.09 ± 0.08, assuming a mean V_S of $5.7\,\mathrm{km\,s^{-1}}$, calculated from a 1-D velocity model for the region (Jerkins et al., 2024). As we expect the majority of the anisotropy to be concentrated in the upper 5 km we also calculate ξ using this assumption, fixing the ray path length to 5 km and using an average V_S of $4.2\,\mathrm{km\,s^{-1}}$. Under this assumption, the mean ξ is 2.68 ± 0.26. Figure 10 shows histograms of ξ for all shear-wave splitting measurements.

5 Discussion

The results show that shear-wave splitting can be measured for microseismicity recorded by seafloor permanent reservoir monitoring systems. This demonstrates projects with nearby seafloor passive seismic sensors could use shear-wave splitting as a valuable tool for monitoring *in situ* S_{Hmax} azimuth at a higher spatial resolution than can be practicably achieved with stress data derived from borehole data. The percentage anisotropy results suggest that, as expected, the shear-wave splitting is mainly sensitive to anisotropy due to aligned microcrack in the upper crust. When we assume this region is 5 km thick, $\bar{\xi} = 2.68 \pm 0.26$, which is in line with what would be expected for anisotropy due to aligned microcracks in the uppermost crust (e.g., Teanby et al., 2004a; Al-Harrasi et al., 2011a). The ϕ_f results show a clear bimodal pattern (Figure 8a). This is due to spatial variability in the data, with ϕ_f data for the Group 2 stations (circular mean $-67 \pm 4°$, Figure 9b) strongly agreeing with regional S_{Hmax} azimuth ($108 \pm 4°$; Figure 8b). The ϕ_f results for Group 1 form a second cluster rotated by 45° from the Group 1 results. This could represent a 45° rotation in the local S_{Hmax} azimuth at these southern and western stations. Similar scale local scale variations in ϕ_f have been observed by local studies of shear-wave splitting for microseismicity near a geothermal project and interpreted as local rotation of the S_{Hmax} azimuth (Hudson et al., 2024). Reservoir scale rotations of microscale fracture strike were observed for the Valhall field from amplitude variation with offset and azimuth (AVOA Hall and Kendall, 2003) and seismic interferometry (Mordret et al., 2013). Spatial and temporal variations in shear-wave splitting were also observed using downhole microseismic monitoring, with reservoir scale rotations in ϕ_f of up to 90° (Teanby et al., 2004a). At Valhall the elliptical pattern in seismic anisotropy is associated with a radial S_{Hmax} azimuth tangential to a production related subsidence bowl (Hatchell et al., 2009; Herwanger and Horne, 2009; Mordret et al., 2013). The temporal variations in seismic anisotropy have been since been explained a cyclic recharge and dissipation of cap-rock stresses in response to production-driven compaction of the underlying oil reservoir (De Meersman et al., 2009).

One important difference between the data used in this study and anisotropy studied at Valhall is that the seismicity have depths in the range of $19.5 - 26.2$ km (Jerkins et al., 2024), and that, therefore the shear-wave splitting and interpreted S_{Hmax} corresponds to the stress field averaged across the upper ca. 5 km of crust and not the state of stress within the Snorre field or overlying formations. Whilst the rotation in ϕ_f at Snorre may be associated with the depletion history of the reservoir, with the data available other explanations cannot be ruled out. One plausible alternate explanation is a more complicated anisotropic fabric, perhaps due to multiple cross cutting fracture sets (e.g., Al-Harrasi et al., 2011a; Verdon and Kendall, 2011). For this study it was only possible to use data from a handful of PRM stations at Snorre, and analysis data from all 10,708 PRM sensors would allow for more detailed analysis as to whether this rotation in ϕ_f represents a local rotation in S_{Hmax} across the area.

Shear-wave splitting has been previously measured using microseismicity at reservoir depths using both downhole (Teanby et al., 2004a) and seafloor instruments (e.g., Jones et al., 2014). Our results demonstrate that shear-wave splitting can also be routinely measured using naturally occurring microseismicity using offshore seafloor monitoring systems. Shear-wave splitting represents a valuable secondary dataset for a monitoring program, measured after microseismicity has been detected and located, which can be used passively infer the *in situ* stress field. Whilst deploying a PRM system at the same scale as that at Snorre may be unfeasible for typical CO_2 storage projects, the results for the 10 sensors shared with the NNSN (magenta triangles, Figure 5) yield a significant (15 out of 25) proportion of the shear-wave splitting dataset. Removing the 10 measurements made using an additional 50 sensors for the Tampen Spur mainshock does not change the interpretation of the results. This shows that even a minimal seafloor deployment could be used to measure microseismic shear-wave splitting for stress field monitoring.

To maximise the value shear-wave splitting can add to a monitoring network it is important to consider the limitations of the technique, particularly the spatial limitations of the shear-wave window effect. For this study we are fortunate that the 2022 Tampen Spur mainshock and subsequent aftershocks were sufficiently deep, in the range of 19.5 – 26.2 km (Jerkins et al., 2024), and close to the Snorre field that, with the exception of stations SNO7, SNO8 and SNO10, the PRM system was within the shear-wave window of natural seismicity. This limits the impact on data availability due to the shear-wave window effect (Nuttli, 1961). To make use of data from potential induced microseismicity, which may have depths on the order of ca. 1 km depending on the depth of the reservoir, sensors would have to be more densely spaced or placed closer to the microseismic events. The amount of good quality shear-wave splitting measurements which could be made relative to the number of measurements attempted is lower than for land based studies of microseismic shear-wave splitting (e.g., Teanby et al., 2004b; Igonin et al., 2022; Asplet et al., 2024; Asplet et al., 2025). This was expected as offshore sensors typically yield nosier passive seismic data. Where measurements are made, the shear-wave waveforms are clear and measurements can be made for multiple earthquakes with $M_L < 1.0$.

Presently, shear-wave splitting can be used to monitor S_{Hmax} azimuth using microseismicity (e.g., Igonin et al., 2022; Hudson et al., 2024; Kühn et al., 2025; Asplet et al., 2025) and to infer temporal variations in stress (e.g., Teanby et al., 2004a; Stork et al., 2015). Further work should link models of stress-induced shear-wave splitting, such as the anisotropic poroelasticity (APE) model of Zatsepin and Crampin (1997), with geomechanical reservoir models. This would allow for the adaptation and extension of existing methods to invert shear-wave splitting for reservoir fractures (e.g., Verdon et al., 2009; Verdon and Kendall, 2011; Al-Harrasi et al., 2011b; Jones et al., 2014), or resolving the magnitude of the *in situ* differential horizontal stress.

6 Conclusion

If seafloor microseismic monitoring infrastructure is to be installed for offshore geological CO_2 storage projects, the information gained on the reservoir and surrounding formations should be maximised. We have shown, using analogous data from the permanent reservoir monitoring network at Snorre, that shear-wave splitting can be measured for microseismicity at the field scale using seafloor instrumentation. Shear-wave splitting has the potential to enable monitoring of the *in situ* azimuth of maximum horizontal stress at a higher spatial resolution that borehole measurements, and could be used for monitoring for fluctuations in maximum horizontal stress azimuth over time. At Snorre, we see that the measured shear-wave splitting fast polarisations are consistent with the regional maximum horizontal stress azimuth in the crust of $108 \pm 4°$ with a possible 45° local rotation in maximum horizontal stress azimuth towards the south of the field. This variation may be due to the depletion history of the reservoir, similar to results at Valhall.

This work demonstrates that shear-wave splitting is a valuable tool for monitoring spatiotemporal changes in maximum horizontal stress azimuth, providing additional reservoir information at minimal added cost. The primary requirement is that the measurement of shear-wave splitting and the likely sources of microseismicity should be considered in the design and installation of the monitoring network. With further research to link geomechanical models and the geophysical observations it may also be possible to use shear-wave splitting to constrain the magnitude of differential horizontal stress, further increasing the value of shear-wave splitting as a tool to monitor *in situ* stress for offshore CO_2 storage projects.

7 Acknowledgements

This work was conducted as part of the "Stress history and reservoir pressure for improved quantification of CO_2 storage containment risks" (SHARP Storage) project, a collaboration between 16 research institutions and commercial companies in Norway, UK, the Netherlands, Denmark, and India under the Accelerating CCS Technologies (ACT3) Programme. The SHARP project has been subsidized through ACT (EC Project no. 691712), by RCN and Gassnova (Norway), ROV (The Netherlands), DST (India), BEIS (UK) and EUDP (Denmark). We are grateful to Equinor ASA and license partners on Snorre (Harbour Energy Norge AS, INPEX Idemitsu Norge AS, Petoro AS, and Vår Energi ASA), for providing us with permanent reservoir monitoring system (PRM) data for the 21 March 2022 Tampen Spur event and its subsequent aftershocks. The views and options expressed in this research are those of the authors and are not necessarily shared by the license partners. TK was also supported by the University of Oxford's Strategic Research Fund, through the Oxford Net Zero programme.

The authors thank Eirik Tvedt, Zoya Zarifi, Brian Baptie, Dan Roberts and members of the SHARP project for discussions, suggestions and feedback which have improved and framed this work. We also thank the reviewers for their constructive feedback on this manuscript.

Part of this research was funded by Natural Environment Research Council (NERC) [NE/W004976/1] as a part of Agile initiative program at the Oxford Martin School.

8 Data Availability

The stress data used here is taken from the World Stress Map (Heidbach et al., 2025) and is publicly available. Earthquake detections and locations used here were first published by (Jerkins et al., 2024) with picks being taken from the Norwegian National Seismic Network bulletin (Norwegian National Seismic Network (NNSN), 2025). The offshore waveform data from was provided to us by Equinor, and the data are not openly accessible to the public. Figures were created using PyGMT (Uieda et al., 2021) and matplotlib (Hunter, 2007). Preprocessing of waveform data was done using routines in Obspy (Beyreuther et al., 2010).

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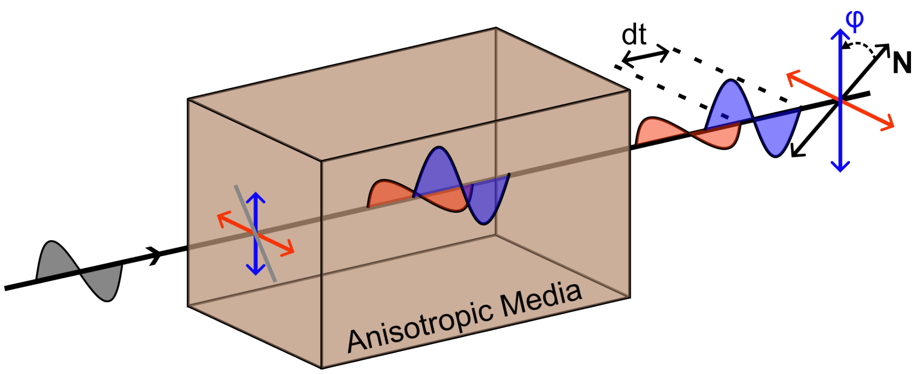

Figure 1: Schematic cartoon illustrating shear-wave splitting.

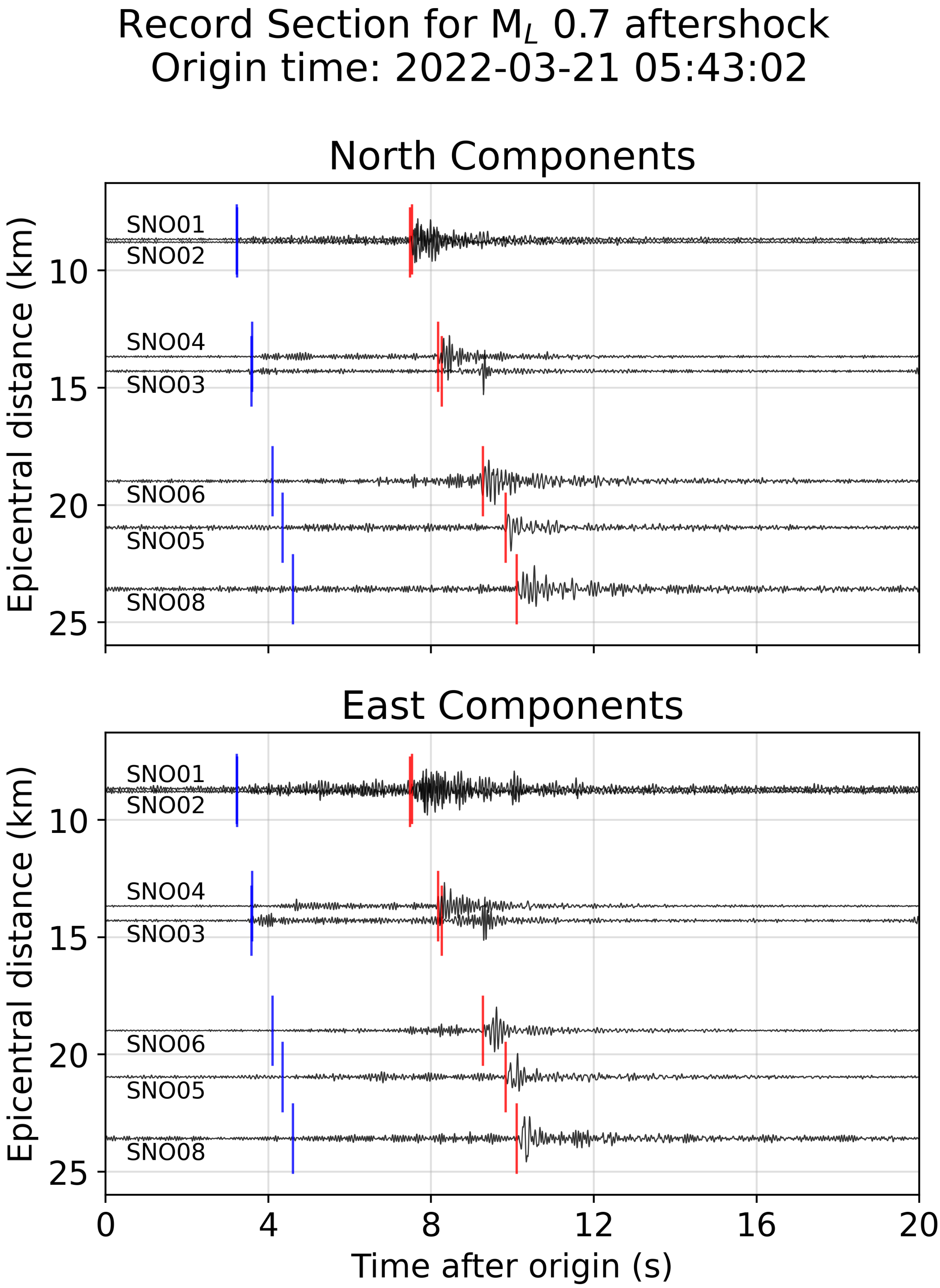

Figure 2: Horizontal component seismograms recorded by permanent reservoir monitoring stations for a M_L 0.7 aftershock of the M_W 21st March 2022 Tampen Spur earthquake. Blue and red bars show the P and S arrival times reported in the NNSN bulletin (Norwegian National Seismic Network (NNSN), 2025). Only data for PRM sensors within the shear-wave window of the event are shown.

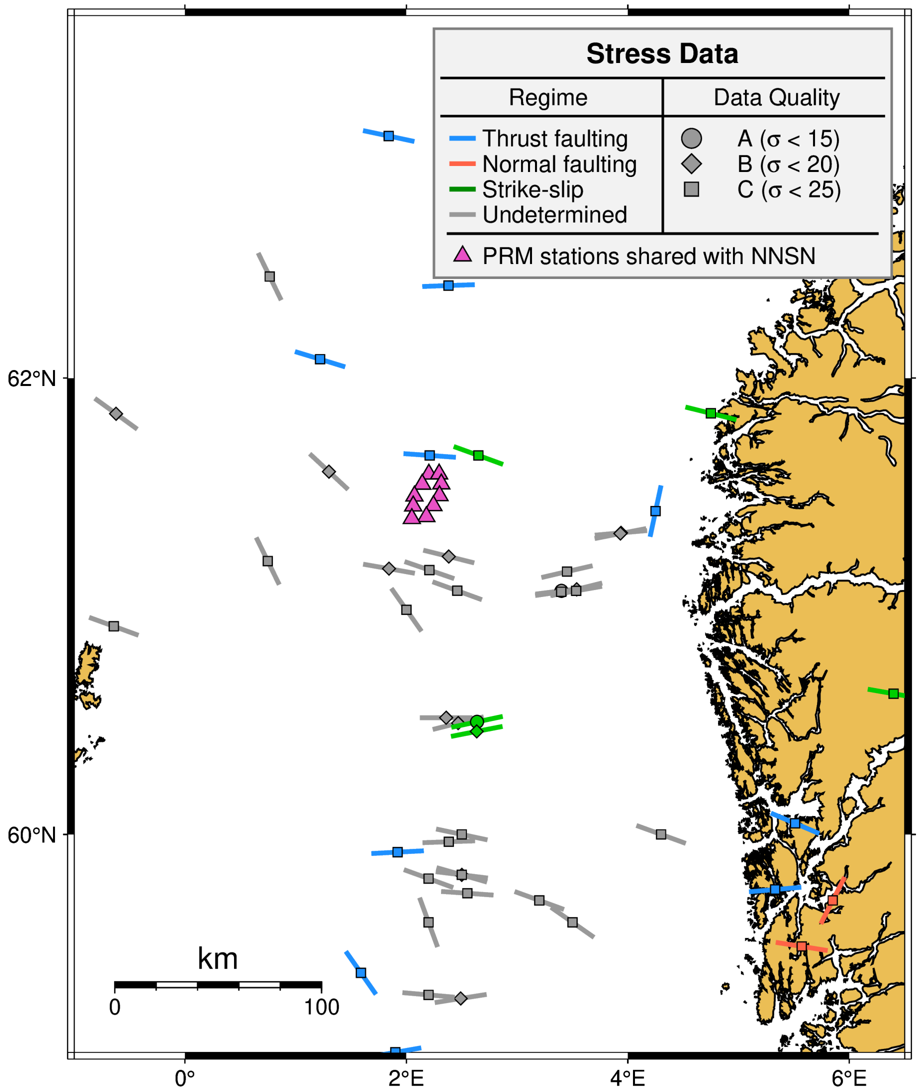

Figure 3: Map showing borehole stress data from the World Stress Map database (Heidbach et al., 2025). Bars show the interpreted S_{Hmax} orientation and symbols correspond to data quality where A (circle) has an uncertainty in S_{Hmax} orientation of $< 15°$, B (diamond) has an uncertainty of $< 20°$ and C has an uncertainty $< 25°$. Grey triangles show the location of Snorre PRM stations that share data with the NNSN.

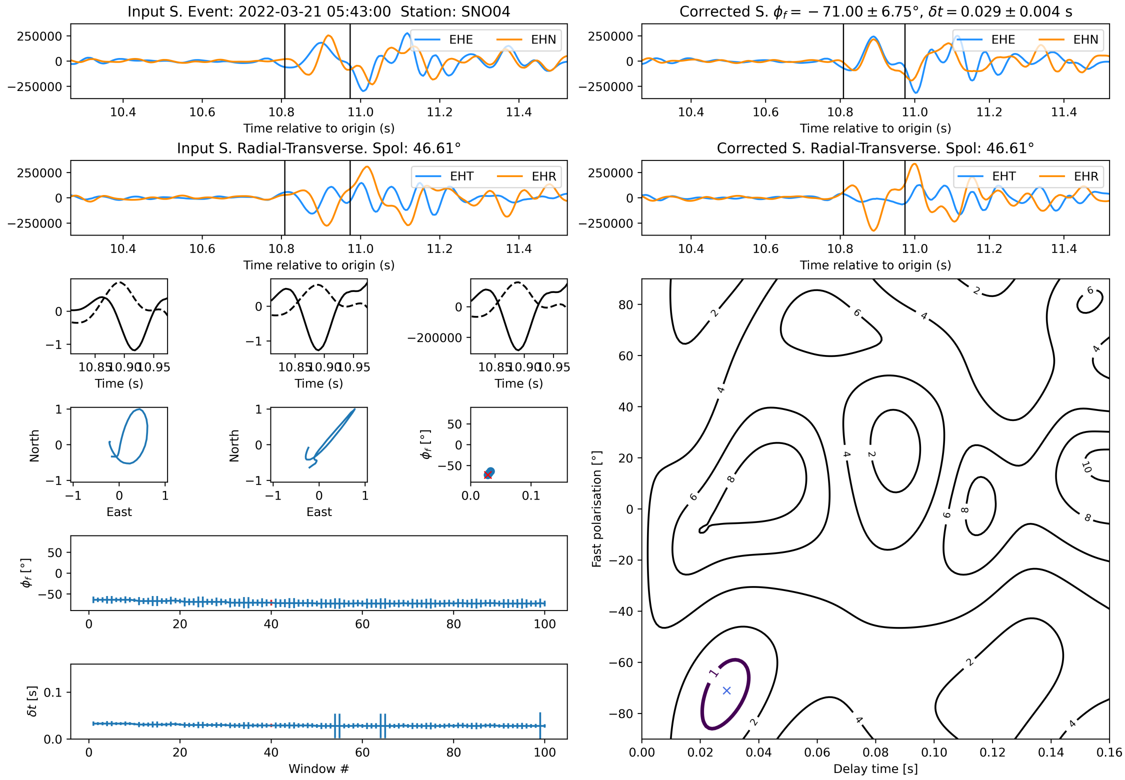

Figure 4: Example shear-wave splitting measurement made at for a M_L 0.7 earthquake which occurred at 2022-03-21 05:41:43 UTC and was recorded by PRM sensor SNO04 at the Snorre field. After manual inspection this measurement is categorised as an 'A' or high quality measurement of shear-wave splitting. Top panels show the input (top left) and corrected (top right) shear-wave phase, where the vertical black bars show the optimum analysis window. The second row shows in the input and corrected waveforms rotated to the measured source polarisation direction. The third row shows the normalised input and corrected waveforms, along with the unnormalised, corrected waveforms. The fourth row shows particle motion plots, which shows the North and East component waveforms plotted against each other, for the input and corrected waveforms, along with the measured fast polarisation (ϕ_f) and delay time (δt) for each window used in the cluster analysis of (Teanby et al., 2004b). Lower panels show ϕ_f and δt plotted against window number. The contour plot on the lower right shows the summarised result of the grid search over $\phi_f, \delta t$ for the optimum analysis window, with contours showing the objective function $\frac{\lambda_2}{\lambda_1}$. Here $\frac{\lambda_2}{\lambda_1}$ has been normalised by the estimated 95% confidence value (Equation 2) with the bold contour enclosing the 95% confidence region. Blue cross shows the best fitting shear-wave splitting parameters.

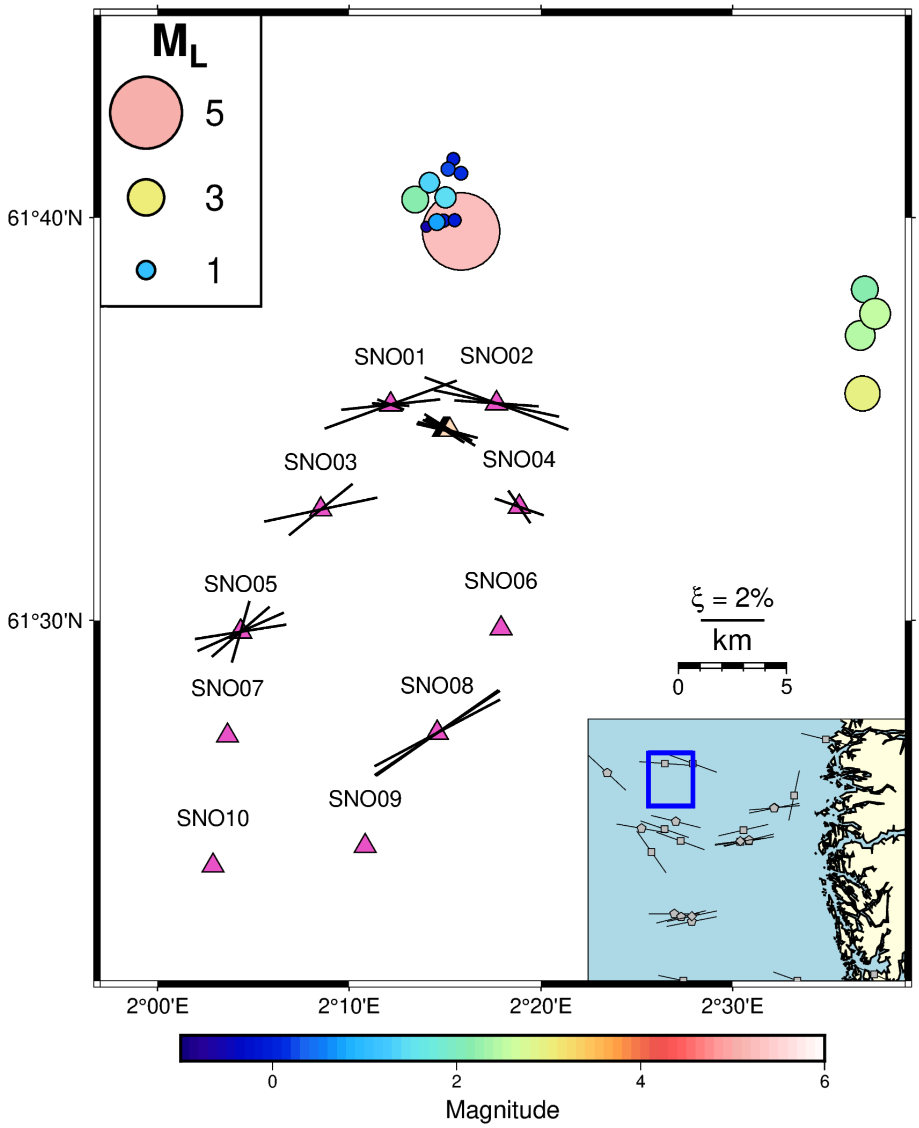

Figure 5: Shear-wave splitting measurements for data from permanent reservoir monitoring (PRM) stations (triangles) at the Snorre field. Shear-wave splitting measurements are shown by the black bars whose orientation corresponds to ϕ and length is proportional to ξ, which is calculated from δt following equation 3 assuming a V_S of $2.8\,\mathrm{km\,s^{-1}}$ and a 5 km thick layer of anisotropy. Earthquakes used (blue circles), the 21_{st} March 2022 M_W 5.1 Tampen Spur earthquake and subsequent aftershocks, are plotted at the locations of Jerkins et al. (2024). Data from 10 PRM stations, which is shared with the Norwegian National Seismic Network (Ottemöller et al., 2021), is used for all earthquakes. For the Tampen Spur mainshock, waveform data from an additional 50 PRM stations was provided by Equinor. Inset map shows borehole stress data taken from the World Stress Map data base (Heidbach et al., 2025), plotted as in Figure 3.

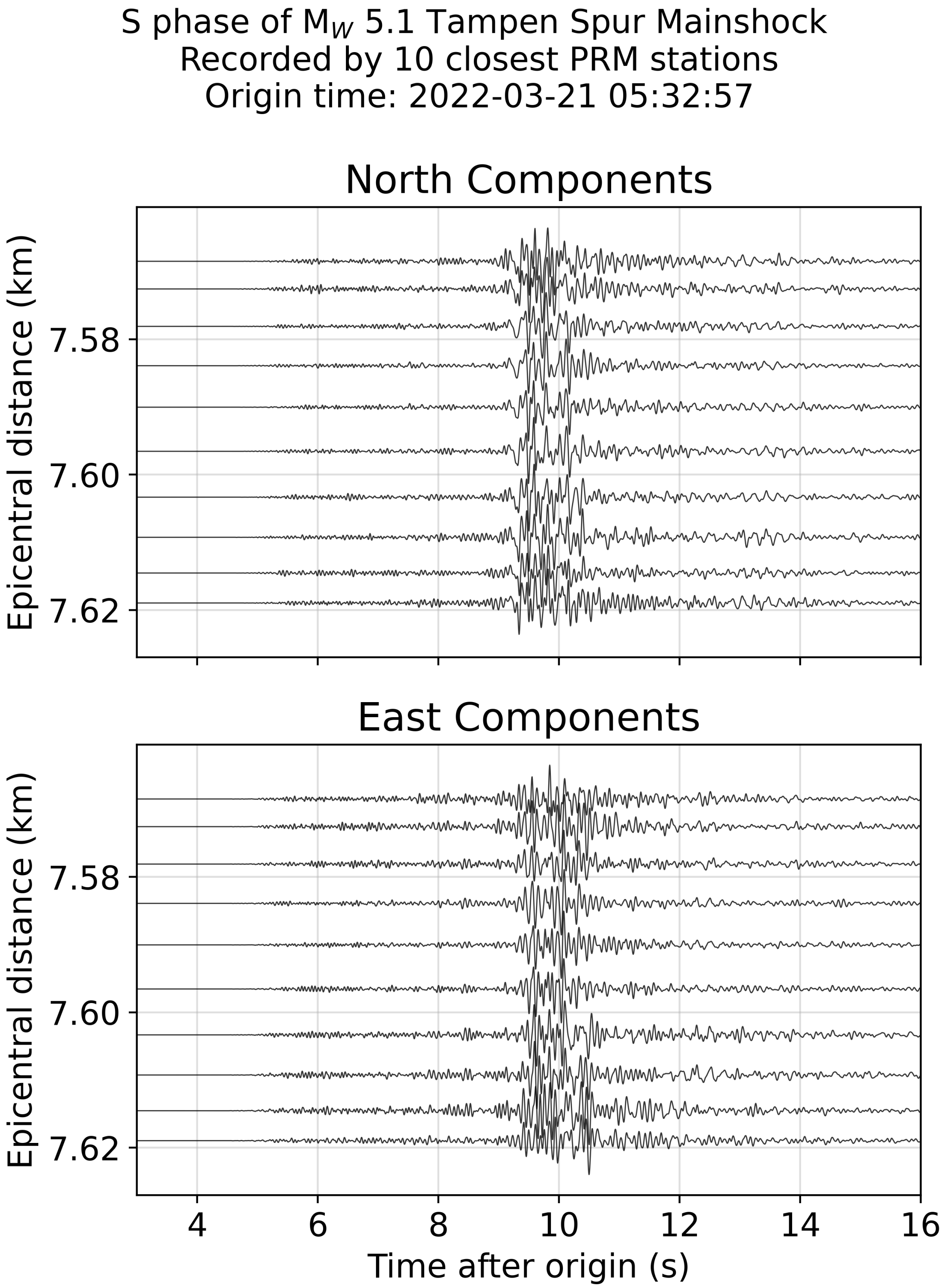

Figure 6: Horizontal component seismograms showing the shear-wave arrivals from the M_W 21st March 2022 Tampen Spur earthquake recorded by the 10 PRM sensors closest to the hypocenter that were available to this study.

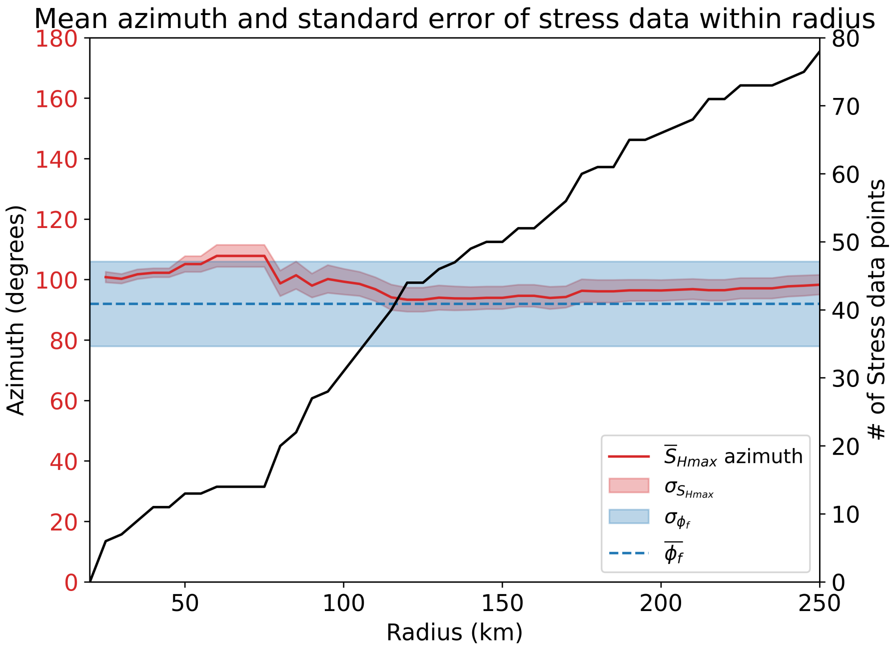

Figure 7: Regional S_{Hmax} azimuth for Snorre (blue line), estimated by taking the circular mean of S_{Hmax} azimuth data within a given radius of the centre of the PRM network, as a function of averaging radius. Shaded region shows the circular standard error. Red line shows the number of datapoints included in the averaging. S_{Hmax} azimuth data is taken from 2025 version of the World Stress Map (Heidbach et al., 2025)

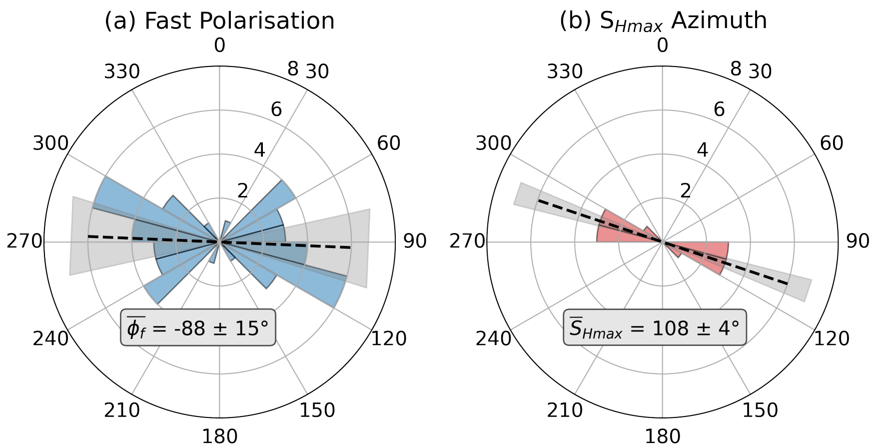

Figure 8: Rose histograms showing the shear-wave splitting fast polarisation data at Snorre (a) and S_{Hmax} azimuth data within 50 km of Snorre (b). The black dashed line indicates the circular mean of each dataset, with the grey shaded region representing the circular standard error. Individual shear-wave splitting measurements can be seen in Figure 5.

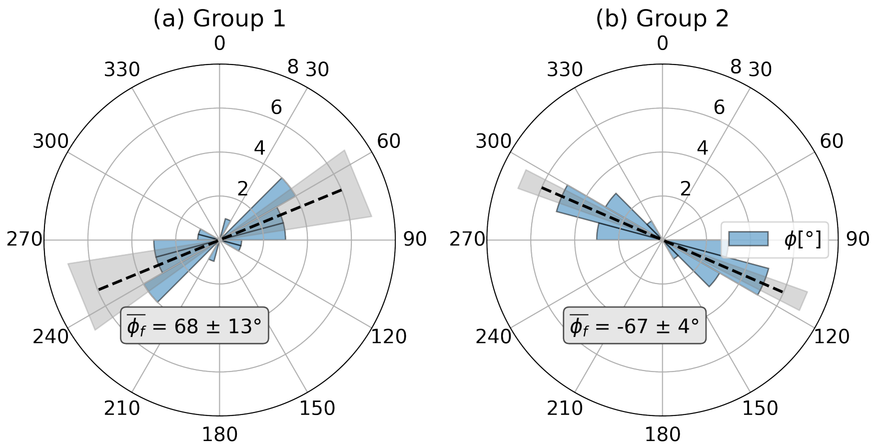

Figure 9: Rose histograms showing the shear-wave splitting fast polarisation data at Snorre recorded by the PRM stations SNO01, SNO03, SNO05 and SNO08 (a) and the remaining eastern stations (b). The black dashed line indicates the circular mean, $\bar{\phi_f}$ of each dataset, with the grey shaded region representing the circular standard error.

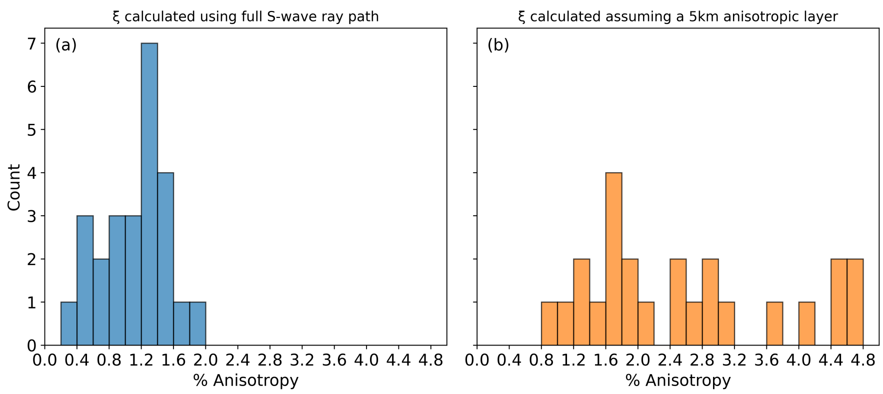

Figure 10: Histograms showing the estimated percentage anisotropy, calculated following equation 3, assuming that anisotropy accumulates along the entire shear-wave ray-path (a) and anisotropy only in the uppermost 5 km of the crust. In (a) we assume a mean V_S of $5.7\,\mathrm{km\,s^{-1}}$ and in (b) we assume a mean V_S of $2.8\,\mathrm{km\,s^{-1}}$.