Mauerberger, A., Maupin, V., Gudmundsson, Ó.,
Tilmann, F. (2021): Anomalous azimuthal variations with
360° periodicity of Rayleigh phase velocities observed in
Scandinavia. - Geophysical Journal International, 224, 3,
1684-1704.

<https://doi.org/10.1093/gji/ggaa553>

Anomalous azimuthal variations with 360° periodicity of Rayleigh phase velocities observed in Scandinavia

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Accepted 2020 November 13. Received 2020 November 6; in original form 2020 February 27

SUMMARY

We use the recently deployed ScanArray network of broad-band stations covering most of Norway and Sweden as well as parts of Finland to analyse the propagation of Rayleigh waves in Scandinavia. Applying an array beamforming technique to teleseismic records from ScanArray and permanent stations in the study region, in total 159 stations with a typical station distance of about 70 km, we obtain phase velocities for three subregions, which collectively cover most of Scandinavia (excluding southern Norway). The average phase dispersion curves are similar for all three subregions. They resemble the dispersion previously observed for the South Baltic craton and are about 1 per cent slower than the North Baltic shield phase velocities for periods between 40 and 80 s. However, a remarkable $\sin(\theta)$ phase velocity variation with azimuth is observed for periods >35 s with a 5 per cent deviation between the maximum and minimum velocities, more than the overall lateral variation in average velocity. Such a variation, which is incompatible with seismic anisotropy, occurs in northern Scandinavia and southern Norway/Sweden but not in the central study area. The maximum and minimum velocities were measured for backazimuths of 120° and 300°, respectively. These directions are perpendicular to a step in the lithosphere–asthenosphere boundary (LAB) inferred by previous studies in southern Norway/Sweden, suggesting a relation to large lithospheric heterogeneity. In order to test this hypothesis, we carried out 2-D full-waveform modeling of Rayleigh wave propagation in synthetic models which incorporate a steep gradient in the LAB in combination with a pronounced reduction in the shear velocity below the LAB. This setup reproduces the observations qualitatively, and results in higher phase velocities for propagation in the direction of shallowing LAB, and lower ones for propagation in the direction of deepening LAB, probably due to the interference of forward scattered and reflected surface wave energy with the fundamental mode. Therefore, the reduction in lithospheric thickness towards southern Norway in the south, and towards the Atlantic ocean in the north provide a plausible explanation for the observed azimuthal variations.

Key words: Europe; Fourier analysis; Time-series analysis; Surface waves and free oscillations; Wave propagation; Wave scattering and diffraction.

1 INTRODUCTION

It has long been known that intrinsic azimuthal anisotropy affects the phase velocities of propagating surface waves primarily due to the alignment of olivine crystals in the lithosphere. Smith & Dahlen (1973) have shown theoretically that the azimuthal variation of the Rayleigh phase velocities has a predominantly 180° periodicity (2θ) in the presence of a weakly anisotropic medium. An early observation of such azimuthal variation was made by Forsyth (1975) with a

maximum anisotropy of 2 per cent at 70 s period. Montagner & Nataf (1986) demonstrated that very shallow anisotropy may even affect the azimuthal variation of long-period waves. The 2θ periodicity is a vital assumption for inversions based on dispersion curve measurements as depth-dependent azimuthal anisotropy (e.g. Montagner & Nataf 1986) and phase velocity tomography. Eikonal tomography studies (e.g. Lin *et al.* 2009) rely on the azimuthal variation as well as two-station methods (e.g. Gombert & Masters 1988; Prindle & Tanimoto 2006) and array based approaches as the two plane wave

method (Forsyth & Li 2005). Although rarely reported in the past, the expected azimuthal variation of phase velocities can be biased in the presence of local strong lateral heterogeneities (Menke & Levin 2002; Lin & Ritzwoller 2011a, b). In these studies, a 360° periodicity has been observed for periods >33 s at different locations in the United States, partly overwhelming the present 180° periodicity Lin & Ritzwoller (2011a). While surface waves are propagating through lateral heterogeneities, the wave train becomes perturbed by finite frequency effects as backward and forward scattering (e.g. Snieder 1986, 2002) where the latter gets masked by wavefront healing (e.g. Nolet & Dahlen 2000). Such wavefield perturbations are strongest when the incident wavelength is similar to or larger than the dimension of the anomaly, yielding to the failure of the ray theory (e.g. Wielandt 1993). Significant phase shifts and amplitude perturbations can occur that imply arrival time deviations of the wavefield (e.g. Snieder & Lomax 1996; Friederich *et al.* 2000; Spetzler *et al.* 2002). As a consequence, the phase velocities are locally perturbed which might lead to relatively strong measurement variations within small scales as shown for 3-D models by Bodin & Maupin (2008). Accordingly, Lin & Ritzwoller (2011b) have shown that also surface wave tomography can be affected by local phase velocity variations leading to over- or underestimated structures. Such a bias becomes substantial when events with uneven azimuthal distributions are used (e.g. Ruan *et al.* 2019; Zhao 2019).

During the past 25 yr southern Norway and Finland, and to a lesser extent central Norway and Sweden, has been explored with 2-D temporary networks as SVEKALAPKO (Bock *et al.* 2001), LAPNET (Kozlovskaya & Poutanen 2006) and MAGNUS (Weidle *et al.* 2010, see Fig. 1) and several passive profiles (e.g. Svenningsen *et al.* 2007), SCANLIPS (England & Ebbing 2012) and TOR (Cotte *et al.* 2002). Some temporary networks (Mansour *et al.* 2018) and active offshore profiles (e.g. Mjelde *et al.* 1995; Breivik *et al.* 2017) covered the Lofoten peninsula and surrounding areas. The permanent Swedish National Seismic Network (SNSN) allows studies of entire Sweden (e.g. Eken *et al.* 2007). In previous studies (see Maupin *et al.* 2013 and Ebbing *et al.* 2012, for reviews), an unusually thin crust has been found beneath the high-topography of southern Norway, that is the southern part of the Scandinavian mountain range seems to lack a crustal root. In fact, the lower topography regions of southern Sweden have a thicker crust than southern Norway, in contrast to the principles of Airy isostasy. A thin, high density layer in the lower crust of Sweden (Ebbing *et al.* 2012) is insufficient to account for this discrepancy. Instead, a pronounced low-velocity zone (LVZ) in the upper mantle beneath southern Norway at depths larger than 100 km has been observed with Rayleigh wave phase dispersion analysis (Maupin 2011; Schaeffer & Lebedev 2013, Fig. 2). A sharp transition in the lithospheric thickness approximately across the Oslo Graben has been confirmed with body wave tomography (e.g. Medhus *et al.* 2012; Wawerzinek *et al.* 2013; Kolstrup *et al.* 2015; Hejrani *et al.* 2017). Although, the estimates of LAB depths below Sweden differ between the different studies, a depth of >200 km is generally agreed upon. The topographic differences therefore seem to be primarily supported in the mantle, although there might be a contribution from Pratt isotropy, that is density variations within the crust (Ebbing *et al.* 2012).

In comparison, northern Norway is only poorly imaged but existing studies, based on sparse data (e.g. Hejrani *et al.* 2017), indicate a transition to low velocities near the coast.

Rayleigh phase velocities in southern and central Finland are similar to those in southern Norway for periods mostly sensitive to crustal structure (<34 s) but on average velocities increase much faster for longer periods, however, with small-scale lateral

variations (Bruneton *et al.* 2004a, b). Further array-based studies using both Rayleigh and Love waves have been conducted in Finland by Pedersen *et al.* (2006) and Pedersen *et al.* (2015). They find that radial anisotropy is dominant within the lithosphere whereas azimuthal anisotropy is only significant in the asthenosphere at depths exceeding 200 km.

The new ScanArray temporary array data set (2012–2017) connects the existing temporary networks and covers mainly Central and Northern Sweden and Norway (Fig. 1), with a few stations along the west coast of Finland. Here, we apply the beamforming technique from Maupin (2011) to this new data set for various subregions of Scandinavia in order to investigate regional-scale Rayleigh phase velocity and its azimuthal variations. We demonstrate, that the observed unusual azimuthal variations with 360° periodicity can be related, to a first order, to the strong local lateral variation of the lithosphere structure.

2 TECTONIC SETTING

The Scandinavian mountains (Scandes) were formed by the Caledonian orogeny in the Palaeozoic age as a result of the continent-continent collision between the Laurentian and Baltica plates. During the Caledonian orogeny mountains have been created also in Scotland, Greenland (Krawczyk *et al.* 2008) and northeastern United States which are known as the Appalachians (Thomas 2006; Hatcher 2010). The Scandes runs approximately parallel to the western coast of Norway. The highest topography is found in southern Norway (southern Scandes) and northern Sweden (northern Scandes) with peak elevations up to 2500 and 2100 m, respectively (Fig. 1). Towards the east, Scandinavia consists of the Precambrian Baltic Shield (e.g. Gorbatshev & Bogdanova 1993) with very low topographies (Fig. 2). Geological reconstructions revealed repeating rifting episodes in the Paleozoic and Mesozoic when the Caledonian mountains have collapsed (e.g. Andersen 1998; Braathen *et al.* 2002). However, it is still under debate to what extent the mountain range destruction took place (e.g. Anell *et al.* 2009; Nielsen *et al.* 2009) as the present-day topography is still unusually high for a passive margin. Consequently, the question arises how the topography could be sustained without recent active tectonic forces. The most accepted hypothesis is a tectonic uplift in the Neogene about 30 Ma (Hay *et al.* 2002), as postglacial isostatic rebound can only be a second-order effect (Gabrielsen *et al.* 2005).

3 DATA

The ScanArray network designates a virtual temporary network of more than 220 broad-band stations consisting of the following contributing networks: 72 stations from ScanArray Core (1G network), which were deployed in Norway, Sweden and Finland by the ScanArray consortium (Thybo *et al.* 2012) specifically for imaging the crust and lithospheric structure below central and northern Scandinavia, 28 stations from NEONOR2 deployed in the Lofoten region (2-D network), 20 stations from the SCANLIPS3D experiment in central Norway and Sweden (ZR network), 72 stations from the permanent Swedish National Seismic Network (SNSN; UP network) and 35 permanent stations from the Finnish (FN, HE), Danish (DK) and Norwegian (NO, NS) permanent networks (Fig. 1, Table 1). The average inter-station distance of the ScanArray Core network is about 70 km. 15 out of the 72 seismic stations have corner periods of 240 s; all others are equipped with 120 s instruments. The SNSN permanent stations cover much of Sweden, but some sites

Figure 1. Map of study area with the temporary and permanent stations contributing to the virtual ScanArray network. The grey boxes encircle previous temporary networks which are referred to in the text. The Scandinavian mountain range dominates the western coast with the highest topography in southern Norway.

are installed with 30 or 60 s instruments. Here, we do not include stations equipped with 30 s instruments into the analysis. The exact operation times of the temporary networks differ slightly, but all the stations overlap from spring 2014 to fall 2016, with the exception of the SCANLIPS3D data collected during 2013–2014. Therefore, we only included teleseismic earthquakes from 2014 to 2016 in the analysis. In total, we considered 257 events with surface wave magnitudes $M_S \geq 5.6$, source depths less than 150 km and epicentral distances up to 130° . The instrument response was deconvolved to obtain velocity seismograms, which were then bandpass-filtered between 0.5 and 200 s and downsampled to 5 Hz. Some stations were significantly misoriented, which we corrected (Grund et al. 2017).

4 BEAMFORMING PROCESSING

4.1 Method

We apply the beamforming technique after Maupin (2011) to investigate Rayleigh phase velocities in several subregions of our study area (Fig. 2). This method has been previously applied to the MAGNUS network (Weidle et al. 2010) in southern Norway, which makes

possible a consistent and direct comparison with our data (Maupin 2011). Surface wave beamforming on a regional scale yields two pieces of information: (1) phase velocity for the region of the array and for each event and (2) the direction of propagation, or equivalently the measured backazimuth (baz), which when compared to the great-circle baz can indicate a systematic deviation. Cross-correlations are calculated in the frequency domain for all possible station pairs within the defined subarray and summed to obtain the final cross-power spectra. We use an event-dependent coordinate system as developed by Forsyth & Li (2005) to account for the curvature of the incoming wave front which cannot be neglected for our regional-scale arrays. As a result of this, and because the defined subarrays are not perfectly isotropic, the array response functions (ARF) vary slightly with azimuth from one event to another.

However, we forgo the additional deconvolution of ARF as done in Maupin (2011). Preliminary tests have shown no significant improvement in the final dispersion curves. The main purpose of ARF is to decrease the beam width to unveil potential higher modes and interfering wave trains, but actually no example has been found where a beam shift improved the detection of higher modes. The slowness is calculated in a local event-dependent coordinate system with a reference point chosen at the centroid of each subarray (Fig. 2). Tests show negligible influence of the position of the

Figure 2. Top: Definition of the subarrays in the context of the global surface waveform tomography shear velocity model by Schaeffer & Lebedev (2013) (version SL2013sv, updated April 2018). Note that the model uncertainty is high for central and northern Norway and Sweden due to the lack of data coverage for their model. The reference station of each array is highlighted with its name. The black line marks the Caledonian front and separates the thrust sheets of the Scandes. OG = Oslo Rift Graben. ISZ is the potential Iapetus-Suture-Zone; ThorS = Thor suture; CM = Continental Margin. The Sorgenfrei-Tornquist Zone (STZ) is the northern extension of the Tornquist-Teisseyre-Zone (TTZ). The gray shaded zone outlines the sharp transition in LAB thickness derived from Kolstrup *et al.* (2015); Hejrani *et al.* (2017). Bottom: Schematic west-east cross-section through southern Norway and Sweden, marked by the dashed black line on the map. LCL is a high-density lower crustal layer after Ebbing *et al.* (2012). The transition in LAB thickness is constrained to be $< \sim 100$ km wide by the body wave tomography of Kolstrup *et al.* (2015).

Table 1. Overview of the networks used for this study. UiB, University of Bergen; GEUS, Geological Survey of Denmark and Greenland.

Network code	Project/network name	Operation time	Stations	Reference
1G	ScanArray Core	2012–2017	72	Thybo <i>et al.</i> (2012)
2D	NEONOR2	2013–2016	28	UiB, NORSAR
ZR	SCANLIPS3D	2013–2014	20	England <i>et al.</i> (2015)
UP	Swedish National Seismic Network (SNSN)	Permanent	72	SNSN (1904)
HE	Finnish National Seismic Network	Permanent	15	Institute of Seismology & University of Helsinki (1980)
NS	Norwegian National Seismic Network	Permanent	8	Lieser <i>et al.</i> (2015)
FN	Northern Finland Seismological Network	Permanent	4	Lieser <i>et al.</i> (2015)
NO	Norwegian Seismic Array Network	Permanent	3	NORSAR
DK	Danish National Seismic Network	Permanent	3	GEUS
IU	Global Seismograph Network	Permanent	2	Albuquerque Seismological Laboratory (ASL 1988)

reference point on the beamforming results when the position slightly deviates from the centroid.

The phase velocity (c) and deviation from great-circle direction (δ_{BAZ}) are determined from the calculated slowness value pair (s_x, s_y) corresponding to the peak beam power according to

$$c = 1/\sqrt{s_x^2 + s_y^2} \quad (1)$$

and

$$\delta_{BAZ} = \tan^{-1}(s_y/s_x). \quad (2)$$

The slowness s_x points into the epicentral direction and s_y into the transverse direction where positive s_y values refer to a clockwise great circle path deviation.

The defined subarrays are displayed in Fig. 2. Scandinavia was separated into a northern (67 stations), central (49 stations) and southern (27 stations) subregion mainly covering Norway and Sweden. In defining the subarrays we aimed to make them comparable in size and for them to have an aspect ratio close to 1 in order to minimize the anisotropy of the ARF. In addition, the limits of the arrays were adjusted to correspond approximately to the changes in the Scandes topography along their western margin. Of course, the size and shape constraints needed to be relaxed a bit in order to cover the whole study area. We also processed the superset of these three subarrays in addition to stations in southern Finland as a single array across Scandinavia, referred to in the following as Complete array (143 stations). Not all stations available within these defined areas were used for the processing since evenly distributed receivers are beneficial for an even sampling of the studied area. In total, we used 159 out of the >220 available ScanArray stations. The three subarrays differ slightly in their aperture, but the regional arrays resolve waves of periods up to 90 s and the Complete array up to 120 s.

All seismograms were windowed between Rayleigh group velocities of 2.7 and 4.6 km s⁻¹. Outside this window a 500 s cosine taper is applied. We performed the beamforming in 10 frequency bands with 20 per cent overlap and center periods of 22, 25, 29, 33, 40, 50, 57, 66, 76 and 86 s. Three additional frequency bands, centered on 100, 110 and 120 s were measured with the Complete array. Events with low signal-to-noise ratio on too many stations of an array were discarded by a fully automatic selection code as described in Maupin (2011). In short, the threshold for discarding traces whose envelopes are too far from the average was set to 1.7 where the average envelope is calculated iteratively after the rejection of one trace. After the final iteration, a minimum number of 10 remaining traces is required for each event.

Finally, 168 events were accepted for the South subarray, 172 events for the Central subarray, 187 events for the North subarray and 188 events for the Complete subarray (see Fig. 3 for the analysed

events for the Complete subarray). The event coverage is acceptable in terms of azimuth and distance for all subarrays, with each 30° segment having at least five events. In contrast to Maupin (2011) we use the same parameters for the analysis of teleseismic and regional events. We incorporate also some very deep events since we find the fundamental mode very coherent across the array without overlapping with higher modes from visual inspection of the data. Further, the upper limit of 4.6 km s⁻¹ for the applied group velocity window reduces the interaction of higher modes.

4.2 Results

An example of the beamforming procedure from the North subarray is shown in Fig. 4. The error bars in Figs 5 and 6 are defined arbitrarily by the 98 per cent contour of the maximum beam strength in the s_x – s_y plane as shown in Fig. 4 and therefore primarily reflect the resolution capability of the array at the respective period. The 98 per cent contour was chosen to make the measure comparable with the results from Maupin (2011). Although the beam width will have some dependency on the coherency of the incoming wavefield, this measure cannot be interpreted as an estimate of the absolute error, but reflects the relative resolution of the array in δ_{BAZ} and absolute slowness direction. We note that it is therefore not meaningful to compare the size of the error bars across different periods because naturally longer wavelengths will result in a wider beam. However, it does not follow that the slowness is necessarily measured with less precision. As can be expected, measurements for the complete array result in smaller error bars.

The measured deviations from the great-circle paths are shown for all events in Fig. 5 at 22, 40 and 86 s. At short periods <29 s the deviations from the great circle path are very high, up to ±30° across all regions. This observation can be explained by waves that traverse heterogeneous lithosphere, including (multiple) plate boundary zones. Short wavelengths are much more affected, which results in complex wave trains due to scattering and multipathing as observed in many studies before (e.g. Cotte *et al.* 2000; Tanimoto & Prindle 2007; Foster *et al.* 2014; Chen *et al.* 2018). No systematic variation with azimuth is observed. If there are any systematic trends here at all, these at most affect events from a particular source region and a very narrow azimuth band. For longer periods with propagation paths in the uppermost mantle the δ_{BAZ} deviation reduces to about ±10°.

Owing to the resolution decrease at very long periods the 86 s δ_{BAZ} deviation measurements result in larger error bars. Interestingly, very distant events from Papua New Guinea/Solomon Islands at 50° δ_{BAZ} are less scattered at 86 s than at 40 s, whereas the far distant events with 260° δ_{BAZ} show more scatter again. All of these characteristics are independent from the source depth.

Figure 3. Distribution of the events used for beamforming for the complete array. The black triangle displays the center of the study area. Grey circles denote 20° epicentral intervals. Grey solid lines mark the major plate boundaries.

Pedersen *et al.* (2015) carried out a Rayleigh waves array analysis for northern Finland (LAPNET network) with focus on great-circle deviations. They found a significant variation with period with a maximum of 9° deviation from the mean over all events used at 20 s. Deviation decreased to about 3° at 100 s. The maximum deviation for individual events at a period of 20 s was 22° in agreement with our measurements shown in Fig. 5.

Fig. 6 shows the phase velocities as a function of baz and epicentral distance for the four arrays at 22 and 40 s where the measured (rather than the great-circle) baz is used. The variation of phase velocity with baz does not appear to be correlated with the baz deviation.

At a period of 22 s we do not observe any clear differences between the arrays. Neither can we identify any pronounced azimuth related variation or event clusters with characteristic anomalies. Only the intermediate-distance events between 0° and 50° baz measured with the North subarray and events from 250° (South America) to 300° (Central America) are slightly more scattered than events from other azimuths. As mentioned above, the surface waves from these directions propagate across multiple subduction zones and both oceanic and continental crust.

At a period of 40 s the three subregions yield remarkable differences. Whereas the central area lacks systematic azimuthal deviations, the northern and southern regions reveal a significant 360° periodicity, that is $\sin(1\theta)$ variation of the phase velocity with baz (Fig. 6b). The 1θ feature is most pronounced in northern Scandinavia where the maximum and minimum velocities vary by ± 2.5 per cent (i.e. $\pm 0.1 \text{ km s}^{-1}$) around the average velocity of 4.05 km s^{-1}

at 40 s, measured from opposite directions at 120° and 300° baz, respectively. From the South subarray we determine a maximum velocity deviation of around ± 2 per cent. We exclude the possibility that our observed 1θ feature is caused by anomalies far away from the array locations somewhere along the propagation path since we observe this pattern independently of epicentral distance. Neither is it affected by event depth or magnitude. The phase velocities calculated from the Complete array also show this azimuthal effect as an average over all subregions.

The azimuthal variation of surface wave velocities in weakly azimuthally anisotropic media has the following form (Smith & Dahlen 1973):

$$c(\omega, \theta) = C_0(\omega) + A_2(\omega) \cos(2\theta) + B_2(\omega) \sin(2\theta) + A_4(\omega) \cos(4\theta) + B_4(\omega) \sin(4\theta). \quad (3)$$

The equation implies that waves traveling along a direction θ should have the same speed as waves traveling in the opposite direction. Here, $c(\omega, \theta)$ is the measured phase velocity as a function of the frequency ω and the propagation azimuth θ . C_0 refers to the isotropic phase velocity. The 2θ term corresponds to a 180° periodicity and is the dominant part of the azimuthal anisotropy for Rayleigh waves, whereas, the 4θ term (90° periodicity) is less pronounced for Rayleigh waves (Maupin 1985; Maupin & Park 2015) and will be neglected here. The coefficients A_n and B_n are also frequency dependent and depend on the variation of the anisotropic parameters of the medium with depth (Montagner & Nataf 1986). The azimuthal variation of observed phase velocities in

Figure 4. Beamforming example for one event from Nepal on 2015/05/12 07:05 UTC for the North subarray. Left-hand panel: record section bandpass filtered with a narrow bandpass around 50 s. Bad traces have been removed. The mean epicentral distance is 55.5° and the mean theoretical backazimuth is 101° . Right-hand panel: Corresponding beam in the slowness plane from which the phase velocities and backazimuth are measured. s_y indicates the deviation from great-circle path which is slightly resolved better than s_x , the slowness in the great-circle direction, due to the array and event configuration. White error bars mark the 98 per cent interval around the maximum beam.

the northern and southern region are clearly not well described by the 180° symmetry expected from azimuthal anisotropy. Therefore, we tentatively attribute the observed 1θ variation to structural heterogeneity.

In order to quantify the observed 360° periodicity we fit the following equation, which, compared to eq. (3), includes the 1θ harmonic contribution as an additional term and drops the 4θ dependence:

$$c(\omega, \theta) = C_0(\omega) + A_1(\omega) \cos(\theta) + B_1(\omega) \sin(\theta) + A_2(\omega) \cos(2\theta) + B_2(\omega) \sin(2\theta). \quad (4)$$

We inverted for the coefficients using a robust L1 norm. We present the fits for the 1θ term (orange lines) only, the 2θ (green lines) only and both terms (blue lines) in Fig. 6(c). To reduce the potential bias due to the uneven azimuthal distribution, we used median binned phase velocities to perform the model fitting. Each black scatter point in Fig. 6(c) represents a median phase velocity within a 10° wide bin; bins are spaced with 5° overlap. Error bars give the standard deviation of all single measurements within each bin.

Outliers have been defined by a threshold of 1.25 times the standard deviation of the residuals relative to a first fit with all data and then removed before the final model fitting. Only a few outliers were identified by this threshold (red crosses in Fig. 6c) but due to the robustness of the L1 inversion, the detailed choice of threshold only has a minor influence on the resulting fits. Both the 1θ (orange lines) and $1\theta + 2\theta$ (blue lines) models fit the observed data within

the error estimates and differ little from each other, whereas neither the 2θ (green lines) assumption implied by seismic anisotropy nor an isotropic phase velocity assumption are able to explain the phase velocity variation from the north and south subarrays. Since the azimuthal variation of the phase velocity is < 1 per cent for the Central array, a single isotropic phase velocity value is sufficient to explain the measurements.

The phase dispersion curves for various subsets of events are shown in Fig. 7 for the north array. The events from the E and SE (blue curve) are up to 0.15 km s^{-1} faster than the median over all events (black line). Events from the W and NW are up to 0.2 km s^{-1} slower than the median curve, that is the total azimuthal variation is > 5 per cent for periods $> 40 \text{ s}$, as seen also in the harmonic function fits. Note that the azimuthal variation is much larger than the lateral variation expected in Scandinavia according to previous studies (Maupin 2011; Pedersen et al. 2013). No clear systematic variation with azimuth is resolved at periods less than 33 s. This is also true for the southern and complete region. In the central area this azimuthal fluctuation is absent at all periods.

As expected, the isotropic phase velocities C_0 calculated by fitting eq. (4) (black dashed line) are very close to the median curve over all individual and binned events. The isotropic and the individual median velocities for the other subarrays are in good agreement as well. Averaging over all azimuths yields a sufficiently good evaluation of the isotropic phase velocity at each period (see Fig. 7). No bias is observable between the isotropic fit of C_0 and the average of the azimuthally binned measurements. The regional median dispersion curves are obtained from all single measurements at each period and compared in Fig. 8. Despite the substantial difference of the phase

Figure 5. Deviation of the predicted backazimuth as a function of the theoretical value for the subarrays at 22, 40 and 86 s. Short periods are more strongly affected by scattering effects. Error bars are derived from the width of the beam, see text for details.

velocity variation with azimuth in the different regions, no significant difference is apparent between the median dispersion curves, except that the Complete array shows slightly faster velocities at long periods. This is reasonable since the Complete array includes some additional stations in southern Finland, which is known to have a high-velocity lithospheric keel (Sandoval *et al.* 2004). The median velocities are similar to the results from the Palaeoproterozoic domain in South Finland (SVEKALAPKO network) (Bruneton *et al.* 2004a) for >45 s. The dispersion curve in southern Sweden (South subarray) indicates higher velocities (~ 2.2 per cent at 70 s) than for neighboring southern Norway (solid grey line in Fig. 8). Our measured phase velocities from all subarrays are ~ 1 per cent lower than the velocities obtained by Pedersen *et al.* (2013) from the Archean domain of the Baltic Shield in North Finland (LAPNET network). Although the North subarray overlaps the LAPNET area significantly, the difference is not surprising, as it extends westward

to the Norwegian coast, where lower mantle velocities appear to be present (Hejrani *et al.* 2017; Mauerberger *et al.* 2020).

To cross-validate our observations of the remarkable azimuthal variation of phase velocity, we applied a different beamforming method based on slant-stacking in the frequency–wavenumber domain (following Park *et al.* (1998), modified by Rindharisaona *et al.* (2020, submitted). In this implementation the reference station is the farthest station for each event and the beam signal with respect to the reference station is summed over all stations, not over all cross-correlated station pairs as in the method of Maupin (2011). We applied this technique to our North and Central subarrays using the same teleseismic events as before. Both subarray measurements are in good agreement with the previous beamforming results and confirm the exceptional azimuthal behavior in northern and southern Scandinavia. Again, the central part lacks any significant phase velocity variation with azimuth.

Figure 6. Phase velocities as function of measured backazimuth resulting from the beamforming processing for the north, central, south and complete arrays (from top to bottom). Panels (a) and (b) show the measurements with error bars for each event at periods of 22 and 40 s, respectively, colour-coded by the epicentral distance. The same colour scale as in Fig. 5 is used here. Note the different y-axis scales. (c) displays the median binned measurements at 40 s with 10° bin widths and 5° overlaps together with their standard deviations (grey error bars). The binned values were fitted to 1θ (orange line), 2θ (green line) and $1\theta+2\theta$ (blue line) harmonic functions. Outliers are marked as red crosses and have been discarded from the fitting (see text).

5 2-D SURFACE WAVE MODELLING

5.1 Method

As discussed in the Introduction, previous studies from southern Norway and Sweden (e.g. Medhus *et al.* 2012; Wawerzinek *et al.* 2013; Kolstrup *et al.* 2015; Hejrani *et al.* 2017) indicate a pronounced step of the LAB at around 13°E and 61°N (cf. Fig. 2). Below this LAB the shear wave velocities decrease significantly by about 5.5 percent (Maupin *et al.* 2013). Less is known about the upper mantle structure below the North subarray, but a shallowing LAB towards the west at the transition from the Baltic Shield to the Caledonian unit or from continental to oceanic lithosphere is also expected (Hejrani *et al.* 2017). As this known lithospheric step is at the western edge of our South subarray with the fast observed phase velocity perpendicular to the step, this lithospheric

scale heterogeneity is a prime candidate for causing the peculiar Rayleigh wave variations. Here, we aim to test with simplified synthetic models whether a structural anomaly on a regional scale could be responsible for the 360° periodicity of the Rayleigh wave velocity anomalies. Thereby, we try to understand the extremes of the variation for waves propagating in opposite directions perpendicular to the strike of the structure. We thus expect to be able to understand the basics of this effect based on 2-D simulations. We do not seek to invert our observed seismograms or dispersion curves exactly, as this is beyond the scope of this paper.

We make use of the *Salvus* software package (Afanasiev *et al.* 2019). The spatial discretization is implemented with the spectral-element method of continuous Galerkin style. A completely automatic association of the entire elements in arbitrary complex models is capable by the implementation to handle fully unstructured domains, instead of regular grids, which save computational time

Figure 7. Dispersion curves for the north array. The solid black curve shows the median over all individual events used for beamforming where the solid dark grey line gives the median over all binned measurements. The blue and the red curves are the median over the events from the fast and slow observed directions, respectively. For comparison, also the median dispersion curves for southern Norway from (Maupin 2011) and for northern Finland (LAPNET network) (Pedersen *et al.* 2013) are shown. The isotropic phase velocities resultant from fitting the coefficients in eq. (4) are shown as dashed line.

Figure 8. Median regional dispersion curves for the arrays defined in Fig. 2. For periods < 32 s the median phase velocities are similar to southern Norway. For higher periods the phase velocities are comparable with the average values from the South Baltic craton.

where the medium is rather homogeneous. This approach holds true for any dimension and any scale. To avoid artificial reflections from the model edges first order approximations of absorbing boundary conditions are implemented in *Salvus* adopting the approach by Clayton & Engquist (1977). Accordingly, energy is absorbed efficiently for nearly all incidence angles.

To simulate teleseismic surface waves which are well separated from the body waves and also not interacting with the model edges the lateral dimension of the mesh has to be sufficiently large. We choose a lateral extent of 10 000 km and a total depth of 400 km for our initial model which is a satisfactory trade-off between the

prerequisites mentioned before and computational time. The mesh has been created to be suitable down to a period of 10 s. Together with a resolution criterion of two mesh elements per wavelength a horizontal grid spacing of 8 km is obtained where the mesh is a simple rectangular grid. The vertical grid size is about 15 km on average. Across boundary layers and edges the mesh size is automatically refined depending on the physical properties of the interface. We built the lithospheric topography starting at a depth of 105 km down to 350 km where the step has a horizontal dimension of 100 km according to previous studies (e.g. Wawerzinek *et al.* 2013; Kolstrup *et al.* 2015). Shear-wave velocities are also adopted from

these previous studies: The crust has three layers and the shallow lithosphere velocity is lower on the thin than on the thick side.

The minimum shear-wave velocity within the low-velocity zone (LVZ) is 4.4 km s^{-1} immediately to the West of the step. Towards the western margin the velocities increase smoothly in order to have a localized LVZ and to avoid an additional step in the model. It is implausible that such a strong heterogeneity exist along the entire path. The maximum vertical V_s contrast across the defined LAB is about 5 per cent and the lateral V_s contrast is 6.4 per cent (*cf.* Figs 2 and 9). The V_p structure is set by using a V_p/V_s ratio of 1.78 and the density (in kg m^{-3}) is approximated by the relation $V_s(\text{m s}^{-1})-1000$. We construct two different models representing a source located either to the west or east of the lithospheric step (Figs 9a and b). These two model set-ups are labelled 270° BAZ and 90° BAZ, which represents the observed slow and fast phase velocity directions (see Figs 6 and 7), respectively. Absorbing boundaries have been set at the left, right and bottom margins of the model.

A moment tensor source is applied with a dominant frequency of 0.03 Hz. This source has been placed 4500 km away from the model boundaries at a depth of 30 km. 13 receivers are located symmetrically above the lithospheric step with a spacing of 50 km corresponding approximately to our real average station spacing. We model the ground velocity and use the vertical component to measure the dispersion curves.

5.2 Results

Since we are dealing with synthetic data where our receivers are placed along a straight line (Figs 9a and b) we have chosen to apply the beamforming technique by Rindraharisaona *et al.* (2020) (submitted) to measure phase velocities from the synthetic waveforms. As mentioned above, we have verified that both beamforming techniques yield consistent results by inspection.

The dispersion curves derived from all synthetic waveforms (receivers R1–R13) are shown in Fig. 9(e). Two reference dispersion curves are also shown based on 1-D laterally homogeneous models which are equal to the left (thin lithosphere) and right (thick lithosphere) step side as marked in Fig. 9(c). The corresponding V_s -depth profiles are shown in Fig. 9(d).

For propagation in the direction of decreasing lithospheric thickness (90° baz source model) in Fig. 9(a) the phase velocities are clearly increased at the longer periods $>45 \text{ s}$ (blue line) compared to both 1-D reference models. The error bars were estimated from the beam in the wavenumber domain for selected frequencies assuming that the individual beams correspond to a normal Gaussian density. A 68 per cent confidence interval around each individual maximum beam was taken. We obtain the final phase velocity mean and variance by sampling from the wavenumber beam distribution.

For propagation in the opposite direction (270° baz case), we observed significantly lower phase velocities for the longer periods. These results are in qualitative agreement with our observed velocities and demonstrate that a strong vertical and horizontal velocity gradient in the lithosphere can produce the 1θ variation of intra-array phase velocities. Note that despite the absence of added noise, the error bars in the heterogeneous models are larger than in the 1-D models, attesting to the more complex nature of the wavefield in these models.

To investigate the effects on phase velocities in more detail, we considered the left (R1–R6, on thinner lithosphere) and right (R7–R13, on thicker lithosphere) subsets of the simulated array separately, where the latter is likely to simulate our real recording

configuration both in the north and in the south, with stations located eastwards of the potential LAB step (Fig. 2). For each source direction, we compare the results at the subsets of proximal (closest to the source) and distant (further away) stations where we find four noticeable characteristics: (I) We note first that the variation depends more on the propagation direction than on the subset of stations used. (II) By comparing the azimuth-dependent variations for the left and right subarrays, we observe that the velocities measured at the proximal subarrays (R1–R6 for 270° baz source and R7–R13 for 90° baz source) are generally closer to the 1-D case than at the distant subarrays. (III) At subarray R1–R6, the velocities are roughly equal to those measured with the full array for both source directions, and therefore display the same variation with propagation direction as the full array. (IV) The R7–R13 measurements also show considerable variation with propagation direction but the velocities are lower than in the R1–R6 subset cases, although that the stations are located on the thicker and therefore faster lithosphere.

We note that the wavelength at 60 s is about 250 km which is the distance between the first and last stations and the lithosphere step. As it may take some distance for the velocity information at depth to reach back to the surface, it is reasonable that it is not only the structure beneath the array but also the structure somewhat shifted towards the source that determines the measured velocity at a given array. Such a lateral shift is observed in Bodin & Maupin (2008) and also imaged in the phase velocity maps by Lin & Ritzwoller (2011a). This might explain to some degree why the velocities with the 90° baz source are higher than with the 270° baz source for which the wavetrain has propagated through the entire LVZ. However, that does not explain the fact that the direction-dependent velocity variation significantly exceeds the difference expected from the difference between the thick and thin lithosphere 1-D models.

As we observe the distant subarrays to be more affected than the proximal subarrays, we may attribute the larger phase velocity deviations primarily to forward scattering and a related mode-coupling mechanism. On the other hand, the proximal subarray velocities are also affected, so there appears to be some influence of reflected (backscattered) waves (Lin & Ritzwoller 2011a), but weaker than the forward scattering effect.

Some snapshots from the surface wave train propagation are shown at different time steps in Figs 10(a)–(d) for the 90° baz model to investigate the wave field perturbation and scattering around the lithosphere step. The leading shorter wavelength wave train seen at, for example 2000 km distance in Fig. 10(a) is likely a higher mode with amplitudes lowered by about 50 compared to the main wave train. They are also well separated in time on the seismograms from which the phase velocities have been measured. As we observe this feature also in the 1-D reference models with their simple dispersion curves, we rule out any influence on the origin of the 1θ azimuthal variation.

The dispersive effect of the Rayleigh waves can nicely be seen with the longer periods sampling deeper structures (Fig. 10a). When the fundamental mode traverses the lithospheric step and the LVZ strong perturbations of the wave field occur where the long-period waves are predominantly affected as seen from the dispersion curves in Fig. 9. Ultimately, the waves are backward scattered and interfere with the forward propagating wave train (Figs 10b and c). The more advanced the wave propagation, the more clearly the backward scattered waves appear (Fig. 10d). Similar scattering effects can be seen for the 270° baz model.

An interference of the incident main wave train and backscattered waves produces phase variations with a wavelength equal to half of the incident wavelength (Maupin 2001). Measured at an array

Figure 9. (a) and (b) Synthetic 2-D models used for full waveform forward modelling. Yellow stars mark the source, yellow triangles across the lithospheric step are the receivers. Source–receiver distances are equal for both models. (c) Zoom of model (a) which is equal to model (b) around the lithospheric step. (d) Shear wave velocity with depth for the thin lithosphere at 2404 km distance (dashed line, left-hand step side) and thick lithosphere at 2700 km distance (solid line, right-hand step side). Both profiles are marked in (c). (e) Phase velocity dispersion curves retrieved from the models shown in (a) and (b). The thin and thick lithosphere dispersion curves result from 1-D reference models with velocity models shown in (d). For the model with the source to the right (90°) much higher velocities are obtained for periods >45 s. The left source (270°) model results in significantly lowered velocities which is in agreement with our data observations shown in Fig. 7. See text for a description of the interpretation of error bars.

with a dimension of at least one wavelength, the backscattered waves are therefore expected to produce noise in the measured phase velocities but not a systematic shift. Considered with our synthetic dispersion curves and the presumed forward scattering, backward scattering seems to have only a secondary impact on the phase velocity variation.

In search of the origin of the 1θ variation we have tested additional models with various shallow lithosphere structure and LAB configurations. In a first test, we altered only the shallow lithosphere above the LAB that we made faster on the thin side and slower on the thick side (i.e. vice versa to the reference model in Fig. 9) as seen in other continental margins (e.g. van der Lee 2002; Fishwick

Figure 10. (a)–(d) Time step snapshots of the surface wave train propagation across the lithospheric step at 830, 918, 1010, 1095 s after origin to demonstrate the wave perturbation and scattering effects around the LAB step. The vertical component of the velocity wave field is shown with a logarithmic colour scale, negative values are masked to highlight the wave perturbations. The source is located to the right of the step on the thick lithosphere side (90° baz source). See Fig. 9 for more details. Grey line outlines the LAB. The inset in (c) refers to the sketch in (f). (e) Amplitude spectrum of $V(z)$ for the wave train shown (d). (f) Summary sketch of the proposed scattering mechanism and main results. The forward propagating fundamental mode from 90° baz (shown as purple lines) is simplified from the inset in (c). The fundamental mode is scattered at the LAB step and within the LVZ both in forward and backward direction illustrated by the black arrows. An interference occurs of the scattered wave with the main wave train. Grey and black triangles indicate the subarrays R1–R6 and R7–R13, respectively. The size of the blue and red arrows above the receivers corresponds to the order of phase velocity variation in Fig. 9. Forward scattering (FS) leads to higher velocity variations than backward scattering (BS).

et al. 2008) and possibly below the north subarray (Fig. S2). We obtained a qualitatively similar pattern of phase velocity variations as before (*cf.* Fig. S2 bottom right and Fig. 9) which is maybe not surprising considering that vertical and lateral Vs contrasts are very similar to the reference model. In another test, we incorporated a high-density lower crustal layer (LCL) on the thick lithosphere side in the model of Fig. 9 according to the study by Ebbing *et al.* (2012, Fig. 2). Again, the magnitude of phase velocity perturbations is comparable to the reference model. Therefore, the lowermost crust and shallow lithosphere structure seems not to be a decisive factor for the occurrence of the phase velocity perturbations, and maybe

surprisingly, does not introduce propagation direction dependent variation at smaller periods, at least in the modeled 2-D case. The discrepancies between the observed and synthetic data with respect to the onset of the 1θ variations (>35 s for observations, >50 s for synthetics) must therefore be explained in a different way and might be related to shallow 3-D scattering in the lithosphere.

Models where we have modified the properties of the LAB transition are illustrated in Fig. 11. A simple model containing three crustal layers as in Fig. 9 but only a homogeneous LVZ with a uniform velocity and enclosed by a homogeneous fast medium is illustrated in Fig. 11(a). This model leads to nearly equal phase

Figure 11. Examples of models with their corresponding dispersion curves. Only the region around the lithospheric step is shown similar to Fig. 9(c). The label above the receivers denotes the model setting. The black dispersion curve is the same reference curve as in Fig. 9 as the velocity structure on the thick lithosphere side does not vary between the models (b–d). See text for details.

velocities for both source directions, although the vertical and lateral V_s contrast is almost 5 per cent as in the reference model. We conclude that the presence of a lithospheric step alone is not sufficient to create a 1θ variation, rather, a pronounced heterogeneous velocity structure is necessary.

By comparing the phase velocity differences between the 90° and 270° baz source models in Figs 11(b)–(d), we note a significant effect of both the lithospheric step width (in km) and the vertical or lateral V_s contrast (i.e. the strength of the LVZ anomaly, where the vertical and lateral V_s contrasts are similar, cf. Fig. 9d). In case of a 100 km step width and a very strong 11 per cent vertical V_s contrast (Fig. 11b) we measure distinct fast and slow phase velocities for the two source directions. However, the velocity discrepancy for the 270° baz source is slightly higher compared to a 5 per cent vertical V_s contrast as shown in Fig. 9, whereas the 90° baz source gives rather similar velocities. Hence, the dispersion curves are not simply proportional to the vertical V_s contrast. For a much broader step width of 400 km, even with a strong vertical V_s contrast (Fig. 11c), or a sharp transition (100 km step width) but a weak vertical V_s contrast (Fig. 11d), the phase velocity variations further reduce and become increasingly less affected by the propagation direction.

Fig. 12 summarizes the phase velocity measurements made in some additional models with various LAB configurations for a period of 80 s. The relative phase velocities are determined with respect to the 1-D thick lithosphere reference model as shown in Fig. 9(d) since the structure to the right of the step is the same for all models. Generally, the phase velocity discrepancy drops down approximately proportionally to the vertical V_s contrast. Similar azimuthal velocity differences are obtained for both 100 and 200 km wide steps. However, for 400 km wide lithospheric step models the phase velocity discrepancy reduces significantly, independently of the strength of the LVZ. In conclusion, the azimuthal bias becomes negligible for smooth vertical and lateral gradients.

We remark that we did not reproduce propagation direction dependent variations for moderate periods at around 40 s as seen in the observed data (Fig. 7). Although the modelled medium is similar to the observed V_s and depth structure as we know so far from southern Norway, we unambiguously reproduce the directional velocity effect only for periods > 50 s.

6 DISCUSSION

We observed a remarkable Rayleigh phase velocity variation with backazimuth in the northern and southeastern part of Scandinavia, which cannot be explained by seismic azimuthal anisotropy. We will discuss this surprising result in terms of its theoretical interpretation, tectonic implications and further seismic observations.

6.1 Theoretical considerations

Bodin & Maupin (2008) analysed the influence of isotropic anomalies on synthetic phase velocities where the array dimensions are about of the size of the heterogeneity (20–140 km), and both are smaller than the wavelength ($\lambda \sim 80$ –400 km). They also evaluated an azimuthal bias of the anomalies on the phase velocities. Generally, they find coupling to higher modes and Love waves negligible compared to the self-coupling of the fundamental Rayleigh mode. Arrays which are placed very close to such anomalies can generate a combined 1θ and 2θ phase velocity variation with azimuth. This 1θ component, however, is weaker compared to our observations and diminishes very rapidly with increasing distance of the array

from the anomaly. At 100 km distance from the anomaly, which is further away than our left and right subarrays (section 5.1), they did not observe significant phase velocity variations anymore. This could be due to the 3-D nature of their heterogeneity and related wavefront healing which is insignificant in a 2-D setting (e.g. Nolet & Dahlen 2000). The array modelled by Bodin & Maupin (2008) is only 60 km wide and is further away from the heterogeneity than in our case, where the array starts right of the edge of the heterogeneous region. At 20 km distance, that is in proximity to the heterogeneity, the synthetic data by Bodin & Maupin (2008) show a significant 1θ variation of apparent phase velocity.

Heterogeneities which are smaller than half of the wavelength may perturb the phase velocities (Lin & Ritzwoller 2011a). When regarding single phase measurements from only one azimuth waveforms are distorted by reflections and diffractions due to the anomaly. However, these artefacts can be related to the small array dimensions compared to the wavelengths considered and larger aperture arrays such as ScanArray are usually expected to give more robust measurements.

The concept of scattered waves after Aki & Richards (1980, chapter 13, fig 13.11) depends on the ratio between the dominant wavelength of the seismic wave, λ , and the scale of the anomaly, a , expressed as the correlation length ratio $ka = \frac{2\pi a}{\lambda}$ on a logarithmic scale. Strong scattering occurs if the dimension of the structural perturbation and the dominant wavelength are of similar size, that is ka is around 1. This is the case for the lithospheric step in southern Scandinavia with the observed 1θ effect for periods longer than around 40 s ($\lambda > \sim 200$ km) and consistent with the findings of Bodin & Maupin (2008). From our modeling results, summarized in Fig. 12, the threshold for a measurable 1θ variation is a ~ 400 km wide horizontal velocity gradient and a 5 per cent V_s contrast in the lithosphere. In terms of the scattering classification after Aki & Richards (1980), we can therefore evaluate $ka = 0.3 - 1.0$. Hence, the 1θ effect could be placed in the low scattering regime in the Aki & Richards (1980) scheme. Numerical methods are needed to quantify the scattering effects affecting regional arrays.

Datta et al. (2017) conducted a theoretical study of surface wave mode coupling in the presence of various lateral heterogeneities but with a larger dimension than in our case. A waveform amplitude gain of converted modes due to mode conversion appears mainly for higher modes at periods < 30 s which is below our observed 1θ variation period range. Generally, the effect influences the fundamental mode less than the overtones and affects Love more than Rayleigh waves. Taking additionally into account our observed waveforms with a dominant fundamental mode (Fig. 4), we consider any bias by an anomaly caused by higher mode conversion as unlikely but we cannot rule out this completely as coupled higher modes would not appear as separated wave trains in the seismogram. Neither do our modeling results from Fig. 9 conform to local mode expectations (Maupin 2007) as the subarray velocity measurements do not represent the average structure below.

Since there is no evidence for higher mode coupling or conversion, we consider the occurrence of self-mode coupling at the lithosphere step, that is the coupling of the fundamental mode to itself as the likely dominant process. The lithospheric step with its lateral and vertical abrupt variations acts hereby as a scatterer, as illustrated in Fig. 10(f). A mode coupling originates at the point of scattering where a phase shift is directly induced from the coupling (cf. eq. 26 in Snieder 1986; Maupin 2007). In an equivalent formulation, the coupling implies a phase velocity variation of the mode. Note that the sketch in Fig. 10(f) idealizes the problem with a point scatterer. However, due to the long wavelengths $\lambda > \sim 200$ km of the

Figure 12. Overview of the phase velocity results from various 2-D models for 80 s with 400 km depth, respectively. The vertical V_s contrast refers to the largest velocity difference across the LVZ (*cf.* Fig. 9d). The fast and slow phase velocities are determined with respect to the homogeneous reference model as shown in Fig. 9. Blue colors indicate a modeled source to the right of the lithospheric step (90° baz), that is propagation towards shallower lithosphere and red colours show the results from a source to the left, that is propagation towards deeper lithosphere. See Fig. 11 for details.

incident wavefield, the scattered wave should be rather considered as an integration over the lateral extent of the whole scatterer (Snieder 1986) as seen in Fig. 10. According to our modeling results, forward scattering caused by an incident wave from 270° baz would result in a large negative phase shift to produce the observed lowered phase velocities east of the step. By contrast, scattering in backward direction would cause a lower positive phase shift to match with the increased velocities at these arrays. This behavior is vice versa for a propagating wave train from 90° baz. However, that hypothesis cannot be confirmed at the current stage.

The effect of longer-wavelength heterogeneities on surface wave velocities is described by the path-average approximation (PAVA, Woodhouse & Dziewonski 1984). The PAVA, which is generally valid when the lateral scale of the heterogeneous structures is significantly larger than the wavelength, is one of the most vital assumptions in ray theory based tomography, or in general plane wave approximations, since Woodhouse & Dziewonski (1984) formulated their waveform inversion scheme. It states that the effective phase velocity perturbation for a path corresponds to an average of the phase velocity perturbations predicted for the velocity–depth structure below each point of the path. This implies that the direction in which the path is traversed has no influence on the inferred perturbations. Our results indicate that PAVA is locally violated (Fig. 9). This is maybe not surprising as at a period of 40 s the wavelength is about 160 km whereas the lateral inhomogeneous structure is only 100 km wide. We emphasize that we describe a local PAVA violation only as a result of strong heterogeneities immediately below a regional array. It does not necessarily imply that the phase velocity perturbations along the whole propagation path between source and receiver, as used for example in global group velocity tomography,

are not properly described by PAVA even when traversing a strong heterogeneity.

We note that PAVA is often used together with the great circle approximation (GCA) where the wave is assumed to propagate nearly along the great circle path. In this study, however, we clearly distinguish between these two approximations. In contrast to PAVA, any GCA violations appear to only add to the scatter of individual event measurements but not cause a bias as we did not observe systematic variations from the theoretical azimuth (Fig. 5). Neither is the great-circle path deviation larger than in other phase velocity studies that showed either no significant azimuthal variation or the 2θ pattern expected for anisotropic structure. Furthermore, the azimuth of propagation is treated as an unknown in the beamforming, implying that any deviation is effectively taken into account in the determination of the phase velocity. In our measurements, phase velocities for observed opposite propagation directions result in highly different velocities, without any correlations with back-azimuthal deviations (Fig. 5). Therefore, the 1θ effect appears to arise from intra-array wavefield interference, or interactions in the immediate neighborhood of the array, as also demonstrated by the synthetic tests (Section 5.1).

Whereas we noticed and studied the 1θ effect based on beamforming array analysis, it would affect all approaches reliant on ray theory to interpret measurements of phase differences between stations of a regional array, such as the two-station method (e.g. Gombert & Masters 1988; Soomro *et al.* 2016), tomographic inversions reliant on two-station measurements (Prindle & Tanimoto 2006) or also the two-wave method of Forsyth & Li (2005).

In the traditional PAVA approximation only the propagation path is accounted for the phase perturbation while contributions for

source and receiver are neglected (Dahlen & Tromp 1998). However, the newly observed local violation is induced from strong perturbations right below the receivers which might give rise to the discrepancy.

Conceptually, the simplest way to overcome these pitfalls would be the abandonment of the PAVA in preference to more accurate finite frequency schemes (Spetzler *et al.* 2002) or full-waveform inversion. Another helpful methodology applies the concept of coupled modes (e.g. Maupin 1988, 2001; Li & Romanowicz 1995).

The 1θ variation we observe with beamforming is expected to occur also, and possibly more severely, for studies based on a detailed local tracking of the phase of the wavefront within the network of stations, such as Eikonal tomography (Friederich 1998; Lin *et al.* 2009). As shown by Wielandt (1993), the gradient of the phase is equal to the structural phase velocity only for wavefields with a constant amplitude (equivalent to plane wave assumption), and all methods relying on the phase only to measure the velocity will to some degree suffer from deviations from the plane wave model, even within small arrays (Friederich 1998). Two (or several) plane wave models can be used to account for heterogeneities located outside the network (Forsyth & Li 2005), but near-by scattering is more difficult to account for. If coupling to overtones and Love waves is negligible, as discussed before, the solution is to introduce a correction based on the lateral first and second order derivatives of the amplitudes (Wielandt 1993), known as Helmholtz equation. Although this method has been shown to be very effective on synthetic data (Friederich *et al.* 2000; Bodin & Maupin 2008), and used on real data (Lin & Ritzwoller 2011b; Jin & Gaherty 2015), the requirements in terms of precision of the measurement of the amplitude are not met in many studies, including ours. The necessity to mitigate biases due to non-plane wave geometry by a good but also rather uniform azimuthal distribution of paths should be emphasized, also for Eikonal tomography.

We recommend to balance the azimuthal event distribution used for tomographic studies either by subsampling events from well-covered azimuths or by applying a weighting scheme that compensates for unequal backazimuthal coverage. Whereas this does not guarantee to completely remove the bias, our synthetic tests suggest that the azimuthally averaged value is close to the expected value (Figs 7 and S1). We point out that for phase velocity analysis with dense regional arrays, which aims to resolve and interpret detailed structures, considerable phase velocity artefacts may occur and thus bias the inferred shear velocity structures. Body wave tomographic artefacts due to non-uniform source distributions have been discussed previously (e.g. Zhao 2019). The impact of such azimuthal bias on tomographic inversion results for surface waves is demonstrated in Mauerberger *et al.* (2020). Whereas for global imaging with large-scale station distances the velocity or structural bias might be neglected.

The observation of the 1θ effect might also have an implication on anisotropy studies. In the presence of pronounced heterogeneities below the receivers, the structural azimuthal anisotropy might be masked by 1θ variations. In addition, the 1θ effect of the heterogeneities is just the lowest order in azimuth of a more complex azimuthal variation (e.g. Bodin & Maupin 2008). The fact that so strong 1θ variations occur suggests that significant 2θ anisotropy due to heterogeneities may also occur and be mistaken for seismic anisotropy. Lin & Ritzwoller (2011a) have discussed biased 2θ anisotropy and proposed this can be mitigated when finite frequency effects are considered for the inversion.

In conclusion, we deduce from our modeling that self-mode coupling of the fundamental mode as a result of forward scattering,

caused by strong vertical and lateral velocity contrasts at the LAB as well as sharp lithosphere step widths, generates larger phase velocity variations than backward scattering. However, it is important to note that the prevailing scattering mechanism and therewith a potential phase shift, depends on the propagation direction and the location of the regional (sub)array (Fig. 10f). A predominantly self-mode coupling hypothesis is also supported by the synthetic tests of Bodin & Maupin (2008). Nevertheless, we cannot determine with certainty the origin of the 1θ effect from a wave propagation point of view.

In this paper, we considered only wave propagation in 2-D models with opposite propagation directions, simulating propagation perpendicular to the strikes of geological structures. For a further examination of the 1θ effect, modeling of oblique propagation in 2.5-D models (e.g. Maupin 1992; Takenaka *et al.* 2003) would be interesting and could be done in a future study.

6.2 Observations in other regions and tectonic implications

1θ variations of Rayleigh phase velocities have been reported previously by Menke & Levin (2002) and Lin & Ritzwoller (2011a). Menke & Levin (2002) observed this effect at the eastern rim of the Appalachian mountain range in the northeastern United States — which are the orogenic pendant to the Scandinavian Caledonides (Thomas 2006), that is both represent conjugate structures. The lithospheric structure beneath the Appalachians is therefore mirrored compared to Scandinavia. Here, an unusual thin lithosphere (about 100 km thickness) with very low shear velocities (about 10 per cent V_s contrast) in the asthenosphere (e.g. North Appalachian Anomaly) associated with a sharp thickening towards the western Proterozoic craton has been imaged by many studies (e.g. Li *et al.* 2002; van der Lee 2002; Rychert & Fischer 2007). This strong 10 per cent vertical V_s contrast along with the observed ± 5 per cent variation around the average phase velocity at 33–50 s period (Menke & Levin 2002) corresponds well to our modeling results (Fig. 12). Moreover, the fast and the slow observed phase velocity directions by Menke & Levin (2002) are also mirror-inverted where the fast direction is facing the LAB step. This means they made the same observation with regard to the velocity and structural contrast as in Scandinavia. In both cases the lithospheric edge is mainly perpendicular to the continental margin and to the mountain front, that is the transition to the craton.

The lithosphere beneath the Appalachians was subjected to various tectonic processes through time as subduction by the Avalonia Plate, active volcanism, hot spot tracks and the opening of the Central Atlantic Ocean about 200 Ma (Menke *et al.* 2016; Thomas 2006). Nevertheless, the Scandinavian and Appalachian lithosphere are very similar, resulting in the 1θ observations. The last known tectonic process which is supposed to have affected both conjugate structures is an uplift in Cenozoic times about 30 Ma (Hay *et al.* 2002). From these indicators, we can speculate that the imaged lithospheric structures have been created in recent times overprinting Mesozoic and Palaeozoic processes.

Lin & Ritzwoller (2011a) detected a 1θ apparent anisotropy in northwestern United States across the Snake River Plain (Idaho) which is also related to local strong velocity contrasts (Schmandt & Lin 2014). They find increasing 1θ variations for periods > 50 s which is consistent with our observations. Although the geological regime is very different due to the Yellowstone hot spot track, the common feature is a sharp transition in uppermost mantle structure

with a pronounced velocity contrast. However, unlike the Scandinavia and Appalachians settings, the fast directions of the 1θ variation point toward the location of the faster structure. By simulations with finite frequency kernels, Lin & Ritzwoller (2011a) propose prevailing backward scattering near the receiver site to be responsible for the 1θ variation. A dominating backscattering effect on the phase velocity would be strongest when the receivers are very close to a sharp lateral gradient, that is within a distance of $\sim 3/16$ of the wavelength or ~ 50 km at 60 s. This corresponds well to our modeling results and can be nicely seen in the wave propagation snapshots in Fig. 10. Otherwise, no impact of backward scattering can be expected for short periods with wavelengths < 200 km which nearly corresponds to the observed 1θ emergence at 35 s.

South of Scandinavia, at the border to the European continent, a sharp transition in the lithosphere had been reported by Cotte *et al.* (2002) across the STZ. Here, the change from thin (120 km) to thick (200 km) lithosphere strikes roughly WE in contrast to the NS orientation across the Oslo Graben but phase velocity variations were not investigated in that study. Another example of shallow LAB with a sharp lateral step and strong velocity changes can be found in south-east Australia (e.g. Fishwick *et al.* 2008; Rawlinson *et al.* 2017). From these conclusions we predict a high potential of observing a 1θ variation across the Greenland Caledonians (Artemieva 2019), southeast Australia and the STZ.

Moreover, due to the 1θ variation seen from the North subregion in this study, we expect a sharp lithospheric step in the order of 100 km with a pronounced LVZ below the LAB having a V_s contrast of > 5 per cent. We estimate a higher V_s contrast in the north since the 1θ variation is more pronounced than in the south.

The corresponding lithospheric step in the north might result from the transition of the Precambrian craton (Baltic Shield) to the Paleozoic terrane across the Caledonian front as indicated by a recent body wave (Hejrani *et al.* 2017) and surface wave tomography (Mauerberger *et al.* 2020). Another candidate could be the ocean-continent transition because the continental margin is very close to the North subarray (Fig. 2). Also a combined influence of both strong gradients seems reasonable, at least in the north because in the south the continental margin is further away from the coast and the onshore lithosphere gradient.

In contrast, a smoothly varying lithosphere with weak lateral heterogeneities might be responsible for the absence of a 1θ variation in the central area of Scandinavia. By comparison of the reported 1θ effects and in the context to the tectonic features described above, we assume significant lateral heterogeneities on small scales in terms of velocity and LAB thickness lead to 1θ observations.

In southern Norway (Maupin 2011), there is only little evidence (at 50 s period) for a systematic phase velocity variations with azimuth in the MAGNUS data set. This might be due to the fact that she used stations which are mainly located above the flat part of the LAB beneath southern Norway and some stations which coincide with the central subarray in this study where the azimuthal variation is absent. Therefore, some of the stations of the array used by Maupin (2011) are presumably barely affected by scattering related perturbations which mitigate the 1θ variation.

Pedersen *et al.* (2006) conducted an array-based anisotropy survey for both Rayleigh and Love waves in central/southern Finland (SVEKALAPKO network) with moderate 2θ variation for all periods considered. No 360° periodicity was observed. Again, the lack of 1θ variation supports our assumption that for generating this phase velocity effect sharp steps in lithosphere thickness along with strong vertical and lateral velocity gradients are necessary. The lithosphere below central/southern Finland seems to be very

complex on small scales, however, there is no indication for any sharp step with large velocity perturbations as seen across the Oslo Graben (Bruneton *et al.* 2004a).

7 CONCLUSION

From the recent ScanArray project and taking advantage of many permanent stations, we were able to analyse the average phase velocity structure with a beamforming technique. By forming subarrays covering the north, central and south regions of Scandinavia, we were able to unveil exceptional phase velocity variations. In the north and the south we observe a 360° periodicity (1θ variation) of the phase velocity with propagating azimuth for periods > 35 s. Fast velocities were obtained from eastern directions and very low velocities from western azimuths with a maximum deviation of 5 per cent. In the central area the 1θ variation is absent for all periods. By averaging over all propagation azimuths we can nevertheless obtain a stable phase velocity estimate for each subarray, finding little difference between the regions. At mantle depth, the average phase velocities are slightly lower compared to previous results from Finland. The 1θ variation implies also the importance and necessity of an even azimuthal event distribution to avoid biased phase velocity measurements.

Since we suspected a relationship between the lithospheric structure and the observed 1θ variation, we conducted a 2-D waveform modeling for the Rayleigh surface waves. By simulating various structures, we found increased phase velocities for waves propagating from thicker to thinner lithosphere whereas for waves propagating from thinner to thicker lithosphere the phase velocities are clearly decreased. From analysis of the variation of apparent phase velocities for subarrays located to the left and right of the main lithospheric heterogeneity in synthetic tests, we identify forward scattering as the main contributor to the 1θ phase velocity variation. Investigation of the wave train propagation indicates the presence of backward scattering near the lithospheric step, which, however, is likely to be less important than forward scattering as cause for the azimuthal variations in phase velocity measurements.

A 1θ anisotropy has been observed before (Menke & Levin 2002) across the northeastern Appalachian Mountains which have also been created during the Caledonian orogeny. Here, the LAB step as well as the fast and slow phase velocity directions are mirror-inverted compared to Scandinavia. We conclude that sharp horizontal structural gradients in combination with strong vertical and lateral velocity contrasts in the lithosphere lead to 1θ variations. Since we observed the 1θ effect also in northern Scandinavia, we assume a distinct east-west orientated lithospheric structure with pronounced velocity contrasts in that region. Strong horizontally lithospheric gradients could exist due to the Palaeozoic–Precambrian transition across the Caledonian front and/or the ocean–continent transition. Finally, we see high potential to observe 1θ variations across the Greenland Caledonians and southeast Australia as those regions also exhibit strong changes in lithosphere thickness associated with large velocity contrasts.

ACKNOWLEDGEMENTS

This study was funded by the German Research Foundation (DFG) where the LITHOS-CAPP project (grants LITHOS-CAPP DFG Gz TI 316/3-1 and -2) is part of the international ScanArray project. Valérie Maupin acknowledges support from the Research Council of Norway through its Centers of Excellence funding scheme,

Project Number 223272. Many thanks to the GFZ Geophysical Instrument Pool Potsdam (GIPP) and other national and institutional instrument pools for providing seismic instruments. We thank the GFZ GEOFON team for archiving the ScanArray project data. Additional seismic data used in this study were retrieved from the data centers GEOFON, ORFEUS, SEIS-UK and University of Bergen. We are grateful for the engineering and logistical support by Werner Scherer (KIT Karlsruhe), Ben Heit and Thomas Zieke (both GFZ Potsdam) as well as University Uppsala and University Oulu. Seismic data have been pre-processed using ObsPy (Beyreuther *et al.* 2010). Figures were generated with GMT5 (Wessel *et al.* 2013) and Matplotlib2.2 (Hunter 2007). We thank Christian Weidle and an anonymous reviewer for their constructive comments which helped us to improve the manuscript.

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SUPPORTING INFORMATION

Supplementary data are available at [GJI](https://doi.org/10.1016/j.gji.2021.106314) online.

Figure S1. Dispersion curves for the central, south and complete subarrays.

Figure S2. 2-D model similar to Fig. 9 but with different shallow lithosphere structure.

Figure S3. This is the corresponding movie to Fig. 10. File name: movie_model.10.90baz.685.1200sec.avi

Figure S4. Dispersion curves for the other subarrays as shown in Fig. 7. From top to bottom: central, south and complete subarray. The solid black curves show the median over all individual events used for beamforming where the solid dark grey lines give the median over all binned measurements. The blue and the red curves are the median over the events from the fast (between 80° and 160°) and slow (between 250° and 330°) observed directions, respectively. For comparison, also the median dispersion curves for southern Norway from (Maupin 2011) and for northern Finland (LAPNET network, Pedersen et al. 2013) are shown. The isotropic phase velocities resultant from fitting the coefficients in eq. (4) are shown as dashed line.

Figure S5. Top panel: synthetic 2-D model 90° baz source, zoomed around the lithosphere step and similar to Fig. 9(c) but with different shallow lithosphere structure. Yellow triangles across the lithospheric step are the receivers. The 270° baz source model is similar. Lower left-hand panel: shear wave velocity with depth for the thin lithosphere at 2404 km distance (dashed line, left step side) and thick lithosphere at 2700 km distance (solid line, right step side). Both profiles are marked in the top model. Lower right-hand panel: phase velocity dispersion curves retrieved from the models shown above. The thin and thick lithosphere dispersion curves result from 1-D reference models with velocity models shown in the left-hand figure. For the model with the source to the right (90°) much higher velocities are obtained for periods >45 s. The left source (270°) model results in significantly lowered velocities which is in agreement with our data observations shown in Fig. 7.

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