Shallow anisotropy in the Mediterranean mantle from surface waves

F. Marone,[1,2] S. van der Lee,[1,3] and D. Giardini[1]

Received 29 October 2003; revised 11 February 2004; accepted 23 February 2004; published 30 March 2004.

[1] We present new evidence for the existence of the Love-Rayleigh discrepancy in the Mediterranean region and constrain the average polarization anisotropic structure of the Mediterranean mantle. We analysed regional Rayleigh and Love waveforms recorded at 3-component broadband seismic stations. None of the 3-component seismograms could be fit with a single 1D isotropic and smooth velocity model. However, satisfactory fits for individual Rayleigh and Love waveforms could often be obtained using realistic 1D velocity models. We used these 1D path-average velocity structures to derive 3D *S*-velocity models for the Mediterranean region. Our results show that Love waves require higher velocities (about 200 m/s) compared to Rayleigh waves between 30 and 120 km depth. We relate the observed anisotropy to lattice-preferred orientation of crystallographic axes of elastically anisotropic minerals such as olivine. *INDEX TERMS:* 7218 Seismology: Lithosphere and upper mantle; 7255 Seismology: Surface waves and free oscillations; 8150 Tectonophysics: Plate boundary—general (3040); 8180 Tectonophysics: Tomography. **Citation:** Marone, F., S. van der Lee, and D. Giardini (2004), Shallow anisotropy in the Mediterranean mantle from surface waves, *Geophys. Res. Lett.*, *31*, L06624, doi:10.1029/2003GL018948.

1. Introduction

[2] Early evidence for the anisotropic structure of the crust and upper mantle was provided by the incompatibility of dispersion curves measured for Love and Rayleigh waves [e.g., *Anderson*, 1961]. This evidence led to the compilation of a global 1D reference Earth model with 2–4% polarization anisotropy in the mantle down to 220 km [*Dziewonski and Anderson*, 1981]. The widespread character of polarization anisotropy was later confirmed for both the continental [e.g., *Debayle and Kennett*, 2000; *Gung et al.*, 2003] and oceanic upper mantle [e.g., *Nishimura and Forsyth*, 1989; *Ekström and Dziewonski*, 1998].

[3] Knowledge of the anisotropic structure of the Mediterranean upper mantle is scarce and limited to azimuthal anisotropy investigated with *SKS* splitting measurements [e.g., *Margheriti et al.*, 2003; Schmid et al., Delay times and shear-wave splitting in the Mediterranean region, submitted to Geophys. J. Int.] and inversion of *Pn* travel times [e.g., *Hearn*, 1999]. Information on polarization anisotropy is available only for the easternmost Mediterranean basin [*Muyzert et al.*, 1999], where up to 7% anisotropy is observed in the lithosphere, and for the Iberian Peninsula [*Maupin and Cara*, 1992], where anisotropy is required only below 100 km.

[4] Here we have analysed over 700 regional Love waveforms and through comparison with Rayleigh waveforms [*Marone et al.*, 2004], we present new evidence for a pronounced Love-Rayleigh discrepancy in the Mediterranean region. The 3D linear inversion of constraints derived from Rayleigh data [*Marone et al.*, 2004] gave information on the *SV*-velocity structure of the Mediterranean mantle. Applying a similar procedure to Love data provides information on the average polarization anisotropic structure. We discuss depth, magnitude and origin of the required anisotropy.

2. Modeling 3-Component Seismograms

[5] We considered Rayleigh and Love wave trains for shallow and intermediate depth events ($4.1 \leq M_w \leq 7.6$) that occurred between 1985 and 2002 in the Eurasia-Africa plate boundary region recorded at broadband seismic stations of the temporary MIDSEA array [*van der Lee et al.*, 2001] and of European permanent networks. We analysed surface waves with azimuths away from the fundamental mode wave source radiation nodes and good signal-to-noise ratio.

[6] We determined the 1D path average *S*-velocity structure and Moho depth by non-linear waveform fitting of wave trains composed of fundamental and higher mode surface waves following the Partitioned Waveform Inversion (PWI) scheme [*Nolet*, 1990; *van der Lee and Nolet*, 1997; *Marone et al.*, 2004]. Since we mainly analysed waves with short paths, most fundamental and higher modes in the seismograms show coherence and could be fit with a 1D model up to 60 mHz, despite the strongly heterogeneous Mediterranean upper mantle. Some of the shortest paths allow upper frequency limits up to 100 mHz. Although satisfactory fits for 1136 individual Rayleigh and 719 Love waveforms (Figure 1) could be obtained using a realistic 1D velocity model (Figure 2), it was never possible to fit both Love and Rayleigh waves for the same path with a single 1D isotropic smooth velocity model. Love waves always required higher velocities than Rayleigh waves down to 200 km. This is a trend commonly observed throughout the world [*Dziewonski and Anderson*, 1981].

[7] We investigated whether the discrepancy in the 1D velocity models retrieved from Rayleigh and Love waves is determined by methodology choices and consider this unlikely [*Marone*, 2003]. In particular, we tested different misfit definitions and model parametrizations. Synthetic tests show that a horizontally laminated structure [*Mitchell*, 1984] alternating high and low rigidities in the lower crust [e.g., *Pohl et al.*, 1999] and/or lithosphere cannot explain our observations. We also investigated the influence of

[1]Institute of Geophysics, Eidgenössische Technische Hochschule-Hönggerberg, Zurich, Switzerland.
[2]Now at Berkeley Seismological Laboratory, University of California, Berkeley, California, USA.
[3]Now at Department of Geological Sciences, Northwestern University, Evanston, Illinois, USA.

0094-8276/04/2003GL018948$05.00

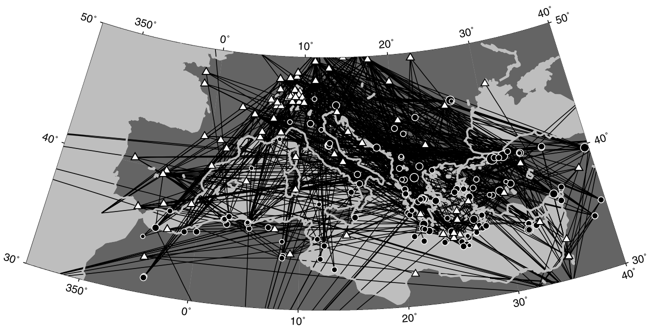

Figure 1. Great circle ray paths for the transverse component seismograms used in this study. White triangles and black dots represent used broad band seismic stations and events, respectively.

scattered energy and uncertainties in source parameters on the 1D models [*Marone*, 2003]. In particular the shallower sensitivity of Love waves compared to Rayleigh waves could make them more susceptible to scattering and refraction at, for example, ocean-continent boundaries. However, these effects cannot account for the observed discrepancy [*Marone*, 2003] and thus we conclude that it is most likely caused by radial anisotropy.

3. 3D *SH*- and *SV*-Velocity Models

[8] In the second step of PWI, linear constraints obtained from waveform fitting and independent constraints on Moho depth taken from the literature are jointly inverted for *S*-velocity and crustal thickness [*van der Lee and Nolet*, 1997; *Marone et al.*, 2004]. Owing to the incompatibility of the 1D velocity models derived from Love and Rayleigh waves and to different characteristics of these two data sets (e.g., path coverage), we inverted linear constraints retrieved from Love and Rayleigh data separately for an *SH*- and *SV*-velocity model, respectively. As starting model we used an average 1D Mediterranean velocity model [*Marone et al.*, 2004]. This procedure produces neither a pure *SV*- nor *SH*-velocity model [*Kirkwood*, 1978], but is suitable [*Wielandt et al.*, 1987] for a first order estimate of the required *SH*/*SV* velocity difference.

[9] The 3D inversion of the linear constraints derived from Rayleigh data and the resulting upper mantle 3D *SV*-velocity model (EAV03) are described in *Marone et al.* [2004]. We performed an equivalent inversion with the linear constraints provided by the Love waves [*Marone*, 2003]. For regions with a comparable path coverage we would expect EAV03 and the derived *SH*-velocity model to have similar characteristics (e.g., size of the anomalies), but this is not the case. Correlation of the uppermost mantle anomalies retrieved from Love waves with surface geology and tectonic processes is not straightforward: the model is dominated by broad high velocity anomalies uncorrelated with tectonic subdomains.

[10] Potential bias from azimuthal anisotropy would differ for the *SV*- and *SH*-velocity models. *Lloyd* [2003] has shown that such bias is negligible for EAV03 [*Marone et al.*, 2004]. To average out azimuthal variations, Love wave path sampling needs to be homogeneous only within half of the azimuthal range required by the Rayleigh waves, due to the 4-lobes dependence of the *SH*-velocity compared to the 2-lobes pattern for the *SV*-velocity [*Lévêque et al.*, 1998]. Thus azimuthal anisotropy is unlikely to significantly affect the observed Love-Rayleigh discrepancy.

[11] Differences are not limited to the retrieved models, but are also observed for the convergence of the inversion and the remaining residual. Although the system of linear equations provided by Love waves is smaller, the inversion needs more iterations to converge and the remaining residual for each constraint is larger, suggesting that the constraints derived from the Love waves might be more internally inconsistent compared to those in the Rayleigh data set. Inspection of surface wave sensitivity to Earth structure indicates that such inconsistencies could arise from trade-offs between crustal and upper mantle velocities, and Moho depth and uppermost mantle velocities, which are larger for Love than Rayleigh waves [*Marone*, 2003]. At long periods, horizontal component seismograms are generally noisier than vertical ones. This could contribute to the larger uncertainties of Love wave data. We thus estimated the uncertainties in the Love data to be twice those in the Rayleigh data.

[12] Due to the instability of the *SH*-velocity model, a direct comparison with *SV*-model EAV03 would not give meaningful information on the anisotropic structure of the Mediterranean upper mantle. We note, however, that similar residuals as observed for this *SH*-velocity model (1.63σ) are achieved with models with a more physical relation to EAV03. We constructed three such *SH*-models to test different hypotheses on the depth and lateral distribution of anisotropy. The first model was obtained in a strongly damped inversion of the linear constraints derived from Love waves using EAV03 as starting model. The highest possible value was assigned to the damping parameter λ that still allowed the Love data to be fitted within two a priori standard deviations (remaining residual = 1.92σ). The strong damping assures that the obtained model shows the

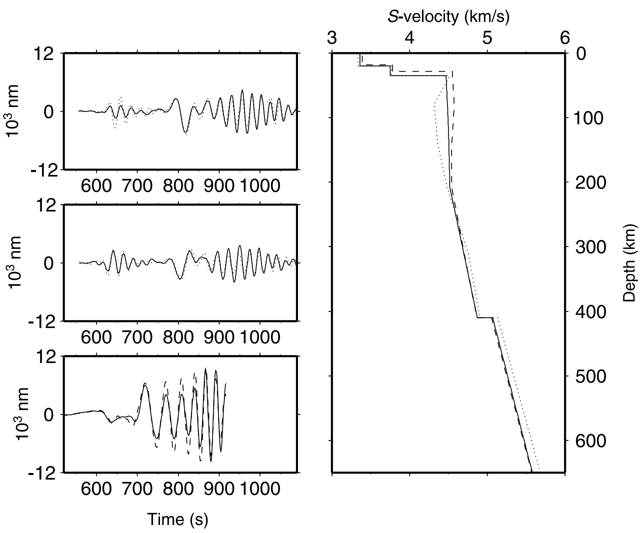

Figure 2. Waveform fitting results for the event in Egypt on November 23, 1995 recorded at the permanent station TAM in Algeria. On the left observed waveforms (solid lines, vertical, radial and transverse component from top to bottom) together with synthetic seismograms (dotted for Rayleigh, dashed for Love) computed with the corresponding 1D path average velocity models (on the right with *iasp*91).

minimum deviation from EAV03 and thus the simplest anisotropic structure required by the data. Although the data fit for this model is not as good as for the *SH*-velocity model obtained in the unconstrained inversion, the data are significantly better matched than by an average 1D Mediterranean model [*Marone et al.*, 2004] (remaining residual = 7.11σ). Two other models were obtained with the same procedure, but damping towards structures simulating radial anisotropy either in the lithosphere or asthenosphere. In both cases, we used EAV03 as starting model, but with increased velocities (by 200 m/s) between 30 and 300 km depth in regions characterized by positive velocity anomalies, to mimic anisotropy in the lithosphere, and between 60 and 300 km depth in regions characterized by negative velocity anomalies, to mimic anisotropy in the asthenosphere. Due to the strong laterally varying Mediterranean lithosphere-asthenosphere boundary, these starting models represent a more accurate simulation of radial anisotropy in either the lithosphere or asthenosphere, than models obtained by adding a constant velocity layer throughout the model at a particular depth and thus allow a more straightforward conclusion on the tested hypothesis. Since the residuals in these two cases are not significantly different (2.32σ and 2.09σ for anisotropy in the lithosphere and asthenosphere, respectively), it is not possible to discriminate between the two hosts for anisotropy. Comparison of the residuals of these two models with the residual of the model obtained by damping towards EAV03 shows that a better data fit is not achieved by forcing anisotropy in the lithosphere or asthenosphere and that anisotropy spread more homogeneously is favored by the data. Thus we prefer the *SH*-velocity model obtained in the inversion damped towards EAV03 (Figure 3). To test its robustness, we performed the same inversion with different starting models (EAV03 with additional constant high velocity layers at different depths, with different velocities and thicknesses). In all obtained solutions, anisotropy is required between 30 and 120 km depth. Reducing the linear constraints derived from waveform fitting to those based on the low-passed (corner frequency = 30 mHz) version of the waveforms increases the relative effect of damping but does not significantly affect the pattern of velocity anomalies.[1]

[13] Our results show that Love waves require higher velocities than Rayleigh waves (about 200 m/s) between 30 and 120 km depth (Figure 3). Deeper anisotropy can exist, but is not required. Below 150 km, the resolving power of the Love data set is reduced, but resolution tests show that velocities significantly higher than of EAV03 would be detected down to 200 km. The retrieved high velocity anomaly relative to EAV03 extends throughout the region covered by the data. Lateral variations of the anomaly amplitudes are not well resolved and resolution tests show their correlation to the inhomogeneous path coverage.

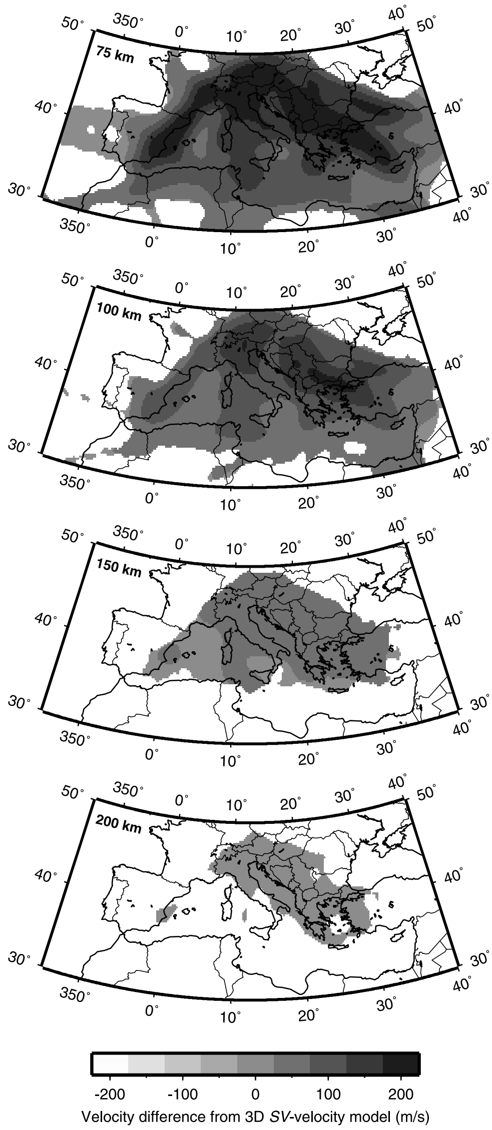

Figure 3. Horizontal slices through our preferred *SH*-velocity model. The velocity anomalies are relative to the 3D *SV*-model EAV03.

4. Discussion

[14] Two mechanisms could explain the anisotropic structure required by our data: preferential orientation of intrinsically anisotropic minerals such as olivine or a horizontally laminated structure alternating high- and low-velocity layers. Azimuthal anisotropy suggested throughout the Mediterranean by *SKS* splitting measurements [*Margheriti et al.*, 2003; Schmid et al., Delay times and shear-wave splitting in the Mediterranean region, submitted to Geophys. J. Int.] and *Pn* studies [*Hearn*, 1999] supports the presence of intrinsic anisotropy. However, the coexistence of preferred mineral orientation and a laminated structure cannot be excluded, though this latter mechanism is unlikely, since 1D velocity models with thin layering of the lower crust and/or lithosphere could not explain the discrepancy of the 1D models derived from Love and Rayleigh data.

[15] Large travel time differences between fast and slow shear wave components (mean value of 1–1.6 s) have been observed in the Mediterranean region [Schmid et al., Delay times and shear-wave splitting in the Mediterranean region, submitted to Geophys. J. Int.; *Margheriti et al.*, 2003]. Considering the average anisotropy in this study (∼4.5%), a delay time of 1–1.6 s would correspond to a ∼100–170 km thick anisotropic layer. A 100 km thick anisotropic zone is

[1] Auxiliary material is available at ftp://ftp.agu.org/apend/gl/2003GL018948.

consistent with our results, a thicker layer is not contradictory, though not required.

[16] Anisotropy depth estimates beneath Italian stations according to *SKS* splitting measurements [*Margheriti et al.*, 2003] suggest larger values (100–250 km) than our study (30–120 km). However, anisotropy in the uppermost mantle (<50 km), consistent with our results, was found in the central and southeastern Mediterranean by *Pn* studies [e.g., *Hearn*, 1999].

[17] From our analysis of regional Love and Rayleigh waveforms we conclude that the Mediterranean on average is characterized by ∼4.5% radial anisotropy in the uppermost mantle down to at least 120 km, probably related to lattice-preferred orientation of crystallographic axes of elastically anisotropic minerals such as olivine. Due to the strongly heterogeneous upper mantle and to the complex lateral variation of the lithosphere-asthenosphere boundary in this region [*Marone et al.*, 2004], it is not possible to relate the observed anisotropy either to frozen-in preferred orientation of olivine in the lithosphere or to alignment of olivine crystals in a present-day asthenospheric flow. Discrimination between these two anisotropy hosts can also not be achieved by the analysis of the residuals for velocity models obtained by inversions damped towards structures forcing anisotropy either in the lithosphere or asthenosphere. However, the shallow depth of the anisotropy required by our data suggests that at least in some regions the lithosphere could be anisotropic. Grain mobility and development of seismic anisotropy throughout the Mediterranean might have been enhanced by the presence of water in the upper mantle possibly released from past and presently subducting oceanic lithosphere [*Karato*, 1995; *Van der Meijde et al.*, 2003].

[18] **Acknowledgments.** The Swiss National Science Foundation (SNF) supported this study. The MIDSEA seismic array was funded by the SNF, with additional support from the Carnegie Institution of Washington, the French National Scientific Research Center and the University of Nice at Sophia-Antipolis, the Italian National Institute of Geophysics and Volcanology and local organizations [see *van der Lee et al.*, 2001]. We thank the individuals associated with MIDSEA [see *van der Lee et al.*, 2001]; their support has been invaluable. An anonymous reviewer provided constructive comments on the manuscript. We benefitted from discussions with W. Friederich. Contribution number 1330 of the Institute of Geophysics, ETH Zurich.

References

Anderson, D. (1961), Elastic wave propagation in layered anisotropic media, *J. Geophys. Res.*, *66*, 2953–2963.

Debayle, E., and B. Kennett (2000), Anisotropy in the Australasian upper mantle from Love and Rayleigh waveform inversion, *Earth Planet. Sci. Lett.*, *184*, 339–351.

Dziewonski, A., and D. Anderson (1981), Preliminary reference Earth model, *Phys. Earth Planet. Inter.*, *25*, 297–356.

Ekström, G., and A. Dziewonski (1998), The unique anisotropy of the Pacific upper mantle, *Nature*, *394*, 168–172.

Gung, Y., B. Romanowicz, and M. Panning (2003), Global anisotropy and the thickness of continents, *Nature*, *422*, 707–711.

Hearn, T. (1999), Uppermost mantle velocities and anisotropy beneath Europe, *J. Geophys. Res.*, *104*, 15,123–15,139.

Karato, S.-I. (1995), Effects of water on seismic wave velocities in the upper mantle, *Proc. Jpn. Acad., Ser. B*, *71*, 61–66.

Kirkwood, S. (1978), The significance of isotropic inversion of anisotropic surface-wave dispersion, *Geophys. J. R. Astron. Soc.*, *55*, 131–142.

Lévêque, J., E. Debayle, and V. Maupin (1998), Anisotropy in the Indian Ocean upper mantle from Rayleigh- and Love-waveform inversion, *Geophys. J. Int.*, *133*, 529–540.

Lloyd, S. (2003), Influence of anisotropy on 3D *S*-velocity models in the Mediterranean region, junior thesis, Eidg. Tech. Hochsch., Zurich, Switzerland.

Margheriti, L., F. Lucente, and S. Pondrelli (2003), *SKS* splitting measurements in the Apenninic-Tyrrhenian domain (Italy) and their relation with lithospheric subduction and mantle convection, *J. Geophys. Res.*, *108*(B4), 2218, doi:10.1029/2002JB001793.

Marone, F. (2003), Inversion of regional seismograms for upper mantle *S*-velocities and crustal thickness in the Eurasia-Africa plate boundary region, Ph.D. thesis, Eidg. Tech. Hochsch., Zurich, Switzerland.

Marone, F., S. van der Lee, and D. Giardini (2004), Upper mantle shear-wave velocity structure in the Eurasia-Africa plate boundary region, *Geophys. J. Int.*, in press.

Maupin, V., and M. Cara (1992), Love-Rayleigh wave incompatibility and possible deep upper mantle anisotropy in the Iberian peninsula, *Pure Appl. Geophys.*, *138*, 429–444.

Mitchell, B. (1984), On the inversion of Love- and Rayleigh-wave dispersion and implications for Earth structure and anisotropy, *Geophys. J. R. Astron. Soc.*, *76*, 233–241.

Muyzert, E., H. Paulssen, and R. Snieder (1999), A seismic cross-section through the east European continent, *Geophys. J. Int.*, *136*, 695–704.

Nishimura, C., and D. Forsyth (1989), The anisotropic structure of the upper mantle in the Pacific, *Geophys. J. R. Astron. Soc.*, *96*, 203–229.

Nolet, G. (1990), Partitioned waveform inversion and 2-dimensional structure under the network of autonomously recording seismographs, *J. Geophys. Res.*, *95*, 8499–8512.

Pohl, M., F. Wenzel, T. Weiss, S. Siegesmund, T. Bohlen, and W. Rabbel (1999), Realistic models of anisotropic laminated lower crust, *Pure Appl. Geophys.*, *156*, 139–155.

van der Lee, S., and G. Nolet (1997), Upper mantle *S*-velocity structure of North America, *J. Geophys. Res.*, *102*, 22,815–22,838.

van der Lee, S., et al. (2001), Eurasia-Africa plate boundary region yields new seismographic data, *Eos Trans. AGU*, *82*(51), 637, 645–646.

Van der Meijde, M., F. Marone, D. Giardini, and S. van der Lee (2003), Seismic evidence for water deep in Earth's upper mantle, *Science*, *300*, 1556–1558.

Wielandt, E., A. Plesinger, A. Sigg, and J. Horalek (1987), Deep structure of the Bohemian Massif from phase velocities of Rayleigh and Love waves, *Stud. Geophys. Geod.*, *31*, 1–7.

D. Giardini, Institute of Geophysics, ETH-Hönggerberg, CH-8093 Zurich, Switzerland. (giardini@seismo.ifg.ethz.ch)

F. Marone, Berkeley Seismological Laboratory, University of California, 215 McCone Hall, Berkeley, CA 94720, USA. (federica@seismo.berkeley.edu)

S. van der Lee, Department of Geological Sciences, Northwestern University, 1850 Campus Dr., Evanston, IL 60208, USA. (suzan@earth.northwestern.edu)