One-dimensional models of shear wave velocity for the eastern Mediterranean obtained from the inversion of Rayleigh wave phase velocities and tectonic implications

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SUMMARY
On a SW–NE profile from the Libyan coast towards central Turkey phase velocity curves of the fundamental Rayleigh mode were measured using a two-station method. The inversion of phase velocity curves yields 1-D models of shear wave velocity down to approximately 200 km depths that may be interpreted as estimates of average models between neighbouring stations on the profile. Strong lateral variations in the shear wave velocity structure are imaged along the profile.

The subducted oceanic African mantle lithosphere is indicated in 1-D models for the region around Crete by significantly enlarged shear wave velocities. It is also imaged by an average model of the structure between stations on Crete and Santorini. On a path crossing the Libyan Sea south of Crete the resulting model is slower than a model expected for 110 Myr old oceanic lithosphere. The passive African margin is thus assumed to extend northwards beneath the Libyan Sea. Anomalous low shear wave velocities are found for the uppermost mantle beneath central Turkey down to a depth of approximately 130 km.

Using two stations on Crete the average depth of the oceanic Moho within the subducting slab is estimated to be at approximately 50 km beneath Crete. For this arc-parallel path, an enlarged standard deviation of the measured phase velocities of approximately 0.2 km s\(^{-1}\) between 10 and 30 mHz is observed that is probably caused by strong lateral heterogeneity related to the subducting slab. In addition, in this frequency range an anomalous propagation of the fundamental Rayleigh mode is detected that is indicated by measured phase velocities that are approximately one standard deviation faster than phase velocities expected from a great-circle approximation. An average shear wave velocity of approximately 3.5 km s\(^{-1}\) is observed above the oceanic Moho.

In order to explain the recent lithospheric structure of the Hellenic subduction zone a tectonic model is assumed for the NE–SW striking profile considered. It is based on the calculated 1-D models, tectonic reconstructions and on a model derived from the metamorphic history of rocks exposed on Crete. The suggested model summarizes the tectonic development at a lithospheric scale starting in the Late Cretaceous. Accretion of crustal material of two microcontinents to Eurasia is assumed, while continuous subduction of the oceanic lithosphere of different ocean basins and possibly of the mantle lithosphere of the microcontinents resulted in a single slab. The length of the oceanic lithosphere that was subducted south of Crete is estimated to be not greater than approximately 550 km.

Key words: eastern Mediterranean, inversion, lithospheric structure, phase velocity, surface waves.

1 INTRODUCTION
The tectonics of the Hellenic subduction zone in the eastern Mediterranean have been studied intensively since the pioneering work by McKenzie (1970, 1972, 1978). The Arabian plate is moving northwards faster than the African plate. Its collision with Eurasia forces the westward extrusion of the Anatolian–Aegean plate. The movement of the Anatolian–Aegean plate may be described by a rigid rotation towards the southwest and an additional internal extension of the plate (e.g. LePichon et al. 1995; McClusky et al. 2000).
Subduction of African oceanic lithosphere along the Hellenic arc and the associated rollback is supposed to be the cause of the extension and might contribute to the rotation of the Anatolian–Aegean plate (LePichon et al. 1995). The seismicity of the region provides valuable information on the geometry of the subduction zone, the deformation and the stress field in the region. Fig. 1 shows relocated ISC-hypocentres of the seismic events between 1964 and 1998 (Engdahl et al. 1998). Pronounced seismic activity is visible along the Hellenic arc. The depths of the hypocentres increase towards the volcanic arc in the region of the Cyclades and the Dodecanese indicating the subducting African lithosphere. The Aegean Benioff zone was studied e.g. by Papazachos & Comninakis (1971), Hatzfeld & Martin (1992), Knapmeyer (1999) and Papazachos et al. (2000). The Benioff zone steepens from the western towards the eastern part of the Hellenic subduction zone. The hypocentres plotted in Fig. 1 show maximum depths in the eastern part of approximately 160 km. North of western Crete hypocentres deeper than 100 km are absent.

The subducting African lithosphere has been imaged by several studies using tomographic techniques. Hovland & Husebye (1982) inverted P-wave traveltime residuals of teleseismic events for the 3-D distribution of the P-wave velocity in the upper mantle beneath eastern Europe. Spakman et al. (1988) showed that the slab of the Hellenic subduction zone penetrates deeper than 200 km into the upper mantle. Granet & Trampert (1989) used a similar technique and presented results that indicate the presence of a NE-dipping slab in the upper mantle beneath the Aegean. Spakman et al. (1993), Bijwaard & Spakman (1998) and Karason & van der Hilst (2000) image the slab down to a depth of approximately 1200 km. Travel-times of local events were studied by Papazachos & Nolet (1997) and Alessandrini et al. (1997) in order to obtain regional models. The model by Papazachos & Nolet (1997) shows that the western part of the slab in the Hellenic subduction dips at a shallower angle than the eastern part. At approximately 100 km depth the two parts almost form a right angle. In the western part a change in the dip angle of the slab was observed at a depth of approximately 70 km. Beneath the Peloponese crustal velocities were detected down to approximately 40–45 km depth and beneath Rhodes down to approximately 40 km. In the same regions the model indicates a low-velocity zone at a depth of approximately 10–15 km. The slab has been shown to be associated with enlarged S-wave velocities.

The first studies of surface wave dispersion in the Aegean were presented by Papazachos et al. (1967), Payo (1967, 1969) and Papazachos (1969). Calcagnile et al. (1982), Kalogeras &
Burton (1996), Martinez et al. (2000) and Karagianni et al. (2002) investigated the structure of the crust and uppermost mantle using path-average group velocities of the fundamental Rayleigh mode for regional paths obtained by the one-station method. 3-D models of the S-wave velocity of the European upper mantle including the Aegean were determined by inversion of long-period waveforms using surface wave mode-summation algorithms for the calculation of synthetic waveforms (Zielhuis & Nolet 1994; Marquering & Snieder 1996).

For the Hellenic arc low Pn-velocities and arc-parallel Pn-anisotropy were observed by Hearn (1999). Low Pn-velocities are reported for western Turkey (Hearn 1999). Numerical modelling of the tectonic evolution of the Mediterranean were presented by de Jonge et al. (1994) and Hafkenscheid et al. (2002). The modelling is based on tectonic reconstructions. The calculated structure is compared with tomographic images. While de Jonge et al. (1994) assume a single ocean basin in the eastern Mediterranean Hafkenscheid et al. (2002) present a model with several ocean basins.

In this study upper mantle S-wave velocities down to approximately 200 km are investigated on a SW–NE profile that runs from northern Africa via the Libyan Sea, Crete and Santorini towards central Turkey. The aim of this study is to obtain information on the local structure of the lithosphere along the profile. For example, the depth of the African oceanic Moho beneath Crete is a matter of debate (e.g. Knapmeyer & Harjes 2000; Bohnhoff et al. 2001; Li et al. 2001). The knowledge of the depth of the oceanic Moho and the depth of the plate interface is essential for the interpretation of the seismicity and the understanding of the lithospheric structure in this region. The profile is shown in Fig. 2. Phase velocity curves of fundamental Rayleigh modes are obtained by a two-station method for neighbouring stations on the profile. They are inverted for 1-D S-wave velocity models that can be interpreted as estimates of average models of the structure between neighbouring stations. Two broad-band stations on Crete are used to investigate the structure beneath the island. The influence of strong lateral heterogeneity caused by the dipping slab beneath Crete on the fundamental Rayleigh mode propagating arc-parallel is detected and quantified. For a SW–NE profile crossing Crete a model of the tectonic development of the lithosphere is presented. It is based on the investigation of Rayleigh wave propagation presented, tectonic reconstructions by Dercourt et al. (1986) and Gealey (1988) and the tectonic model by Stöckhert (1999) for Crete.

2 DETERMINATION OF PHASE VELOCITY CURVES OF THE FUNDAMENTAL RAYLEIGH MODE

2.1 The two-station method

The displacement of the fundamental Rayleigh mode can be approximated by the Jordan–Wenzel–Kramers–Brillouin (JWKB) response to a moment tensor source on a spherically symmetric, isotropic earth model (Snieder & Nolet 1987; Tromp & Dahlen 1992):

\[ u(r, \omega) = p(r) \left[ - (i \omega)^{-1} M : E'(r, \omega) \right] (8 \pi k \sin \Delta)^{-1/2} \exp \left( k \Delta + \frac{\pi}{4} \right) \]

The complex polarization vector is given by \( p(r) = n_s U(r) + i n_\perp V(r) \), where \( U(r) \) and \( V(r) \) are real, radial eigenfunctions. The unit vectors \( n_s(r) \), \( n_\perp(r) \), and \( n_s(r) \) form a local Cartesian basis, which is dependent on the location \( r \). The unit vectors \( n_s(r) \) and \( n_\perp(r) \) point in the radial direction and in direction of increasing epicentral distance \( \Delta \), respectively. The third unit vector is defined by \( n_\perp(r) = n_s(r) \times n_\perp(r) \). The eigenfunctions are normalized according to \( c C_1 = 1 \), where \( c \) is the phase velocity and \( C_1 \) the group velocity. The integral \( I_1 \) is given by \( I_1 = \int_0^\infty \rho (U^2 + V^2) r^2 dr \). \( M \) is the moment tensor for a point source at the source location \( r_s \). \( E(r, \omega) \) is the source strain tensor. The colon indicates contraction over both indices. The wavenumber is denoted by \( k \). Cross correlating the vertical component of the displacement field excited by the same event and recorded at two stations with the same backazimuth yields

![Figure 2](https://example.com/image.png)

**Figure 2.** 1-D models of S-wave velocity are determined for the station combinations MARJ–GVD, GVD–SKD, KRIS–SKD, SKD–SANT, SANT–ISP and ISP–ANTO.

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The epicentre is smaller than 7° the two stations and the great circles through the stations and the paths from the sources to the stations are shown. Events indicating the paths from the sources to the stations are shown. Great circles inscribed for the path between the stations KRIS and Kovach (1978).

Figure 3. Source mechanisms and great circles between the epicentres and the stations for events used to analyse the propagation of the fundamental mode between the stations KRIS and SKD on Crete.

\[ \Phi(\omega) = U(r)^2 \left[ -(i\omega)^{-1} \right] \mathbf{r}_j \left| \mathbf{r}_j \right|^2 \]

\[ \times (8\pi k)^{-1} \sin \Delta \sin \Delta_2^{-1/2} \exp i [k(\Delta_1 - \Delta_2)] \]

if the JWKB approximation is adopted. The indices 1 and 2 refer to the two stations. The phase velocity may be obtained from the phase of the cross correlation function \( \Phi(\omega) \):

\[ c(\omega) = \frac{\omega(\Delta_1 - \Delta_2)}{\arctan \left( \frac{\left| \Phi(\omega) \right|}{\left| \Phi(\omega) \right| + 2n\pi} \right)} \]

Because of the ambiguity of the arctan function the phase velocity estimated from the cross correlation function of the vertical components is an array of curves, that means \( n \in \mathbb{N} \). The two-station method in the frequency domain was introduced by Sato (1955). Since then it has frequently been applied in local studies. Reviews of early applications are given e.g. by Knopoff (1972), Seidl & Mueller (1977) and Kovach (1978).

As an example, the determination of the phase velocities is described for the path between the stations KRIS–SKD on Crete. In Fig. 3 the events used for the analysis are depicted. Great circles indicating the paths from the sources to the stations are shown. Events are only considered if the angle between the great circle connecting the two stations and the great circles through the epicentre is smaller than 7°. The epicentral distances are lower than 120° and larger than 5°. The magnitudes of the events range from 4.9 to 8.0. These search criteria are applied to all the paths considered. Therefore, the events and the source regions are different for every path.

Fig. 4 shows an example for the determination of the phase velocity of the fundamental Rayleigh mode using the two-station method. The vertical components of ground velocity at the two stations are shown together with amplitudes as a function of time and frequency. They are sorted according to the direction of propagation. Scattering at lateral heterogeneity strongly influences the waveform in the frequency range between 50 and 80 mHz. Nevertheless, due to the applied data processing for some events it is possible to determine phase velocity curves up to approximately 100 mHz. In the example of Fig. 4, below 3 mHz the fundamental Rayleigh mode is hidden by noise.

The cross correlation function between the vertical components of the displacement at the two stations is plotted with the corresponding time-frequency representations. The time-frequency representations are used to characterize the long-period waveforms qualitatively. Strong scattering and noise are easily detected in these plots. For the calculation of \( c(\omega) \), the cross correlation function is filtered with a frequency-dependent Gaussian bandpass filter and then it is windowed in the time domain with a frequency-dependent Gaussian window centred around the maximum amplitude of the cross correlation function. This frequency-dependent time window is applied in order to enhance the signal-to-noise ratio. Side lobes in the cross correlation function that are due to noise and correlations between the fundamental mode and other parts of the waveform e.g. the body wave part are down weighted. Then it is transferred into the frequency domain where the phase velocity of the frequency considered is calculated from the phase of the cross correlation function. In Fig. 4 on the right the array of curves defined by eq. (3) is shown. Phase velocities are accepted in a frequency range that is chosen interactively. The dotted-dashed curve depicits the normalized amplitude spectrum of the cross correlation function, the dashed line the phase velocity curve for an initial model. These curves assist the interactive choice of the frequency range in which the estimate of the phase velocity is to be accepted. The plots on the right, in the middle and on the bottom show phase velocity curves for other events. They are sorted according to the direction of propagation. All determined phase velocity curves for that path are shown in Fig. 5.
Figure 4. Example for the determination of the phase velocity using the two-station method. The example is for the path between the stations KRIS–SKD on Crete. On the left, the waveforms at the two stations are shown together with amplitudes as a function of time and frequency calculated by the multiple filter technique (MFT). Automatically determined group traveltimes of the fundamental Rayleigh mode are indicated by solid white lines in the time-frequency representations. The cross correlation function of the two waveforms with the time-frequency representation is depicted on the left at the bottom. On the right in the upper plot the solid lines indicate the array of curves that is the result of the calculation of the phase velocity from the phase of the cross correlation function. Note that an additional frequency-dependent filtering and weighting was applied to the cross correlation function before phase velocities were determined in the frequency domain. The dotted-dashed curve in the upper right plot indicates the normalized amplitude spectrum of the cross correlation function. The dashed line depicts the phase velocity of the fundamental Rayleigh mode for an initial model. From the array of curves interactively the curve is selected that is comparable to the initial model. Furthermore, the frequency range in which the phase velocity is accepted is chosen interactively. The lower plots on the right show previously determined phase velocity curves sorted according to the direction of propagation. The grey curve on the right in the middle shows the phase velocity curve estimated for this example. All determined phase velocity curves for this path are shown in Fig. 5.

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2.2 Phase velocity curves of the fundamental Rayleigh mode

The two-station method was applied to the combinations of broadband stations shown in Fig. 2. Most of the stations belong to the Geofon network (Hanka & Kind 1994), MARJ to the MIDSEA project (van der Lee et al. 2001) and ANTO is a GSN station. The path MARJ–GVD crosses the Libyan Sea and runs from northern Africa to Crete. This path supplies information on the lithosphere beneath the Libyan Sea south of Crete. The station combination GVD–SKD allows one to study the structure of the subduction zone south of Crete. The path KRIS–SKD was added in order to investigate the deep structure of Crete. The paths SKD–SANT, SANT–ISP and ISP–ANTO complete the profile towards central Turkey. The distances between the stations are: MARJ–GVD, 393 km; GVD–SKD, 65 km; KRIS–SKD, 146 km; SKD–SANT, 174 km; SANT–ISP, 478 km; and ISP–ANTO 300 km.

Fig. 5 shows the measured phase velocity curves. The number of curves varies between 6 and 66. Because the orientation of the paths with respect to active source regions is similar, the varying number of phase-velocity curves is due to different times of observation.

A synthetic test was performed in order to estimate errors in the measured phase velocities due to random noise. For the paths GVD–SKD and KRIS–SKD synthetic waveforms were calculated for the events considered in Fig. 5 and Gaussian white noise was added with a standard deviation that corresponds to 5 per cent of the maximum amplitude of the synthetic waveforms. Fig. 6 shows average dispersion curves and their standard deviations. The grey curves denote the phase velocity determined for synthetic waveforms without noise. In this case, the standard deviation is so small that the average curve and the standard deviations are graphically indistinguishable.

For noisy data, the standard deviation of the phase velocity is larger for shorter paths (e.g. GVD–SKD) and frequencies lower than approximately 20 mHz. However, the average phase velocities are unbiased due to random noise. It can be concluded that for the path GVD–SKD the measured phase velocities for frequencies lower than approximately 20 mHz are not well determined because of the limited number of phase velocity curves and the small distance between the stations. That means the structure below approximately 100 km depth cannot be determined for this path. On the other hand side, the small interstation distance allows one to determine phase velocities for frequencies up to 120 mHz (Fig. 5).

In Fig. 7 measured average phase velocity curves are plotted together with the standard deviation. The phase velocity curves of the fundamental Rayleigh mode were determined in the frequency range from approximately 3 to 100 mHz. This corresponds to depths of approximately 400 to 15 km. The observed dispersion indicates increasing the S-wave velocity with depth. Remarkable are the dispersion curves for GVD–SKD and KRIS–SKD. High velocities due to the slab are clearly visible in the dispersion curves between 15 and 40 mHz for GVD–SKD and between 15 and 30 mHz for the path KRIS–SKD. A strong change in the phase velocity is found for GVD–SKD between 30 and 60 mHz and for KRIS–SKD between 20 and 50 mHz. They are caused by the change in velocity across the oceanic Moho within the African lithosphere. The phase-velocity curves reflect the fact that the slab is deeper directly beneath Crete than south of Crete because the change occurs at deeper frequencies for the path KRIS–SKD. The estimated standard deviation varies between 0.1 and 0.2 km s$^{-1}$. For shorter paths and for the path KRIS–SKD, which runs almost parallel to the strike of the subduction, the standard deviation is increased to approximately 0.2 km s$^{-1}$.

Lateral heterogeneity due to the subducting African lithosphere that dips beneath Crete at an angle between 15$^\circ$ and 20$^\circ$ towards NE leads to a complicated propagation of the fundamental mode that is supposed to cause the enlarged standard deviation for the path KRIS–SKD.
3 INVERSION OF PHASE VELOCITY CURVES OF THE FUNDAMENTAL RAYLEIGH MODE

The dispersion curves were inverted for 1-D models by a gradient search using a Gauss–Newton algorithm. That means, the relation between the model parameters and the dispersion curve is linearized in the vicinity of the considered values of the parameters. Starting from initial estimates, model parameters are iteratively improved until a good fit between theoretical and measured curves is obtained. Theoretical phase velocities for elastic, isotropic, 1-D models are calculated using a Thomson–Haskell matrix formalism and an earth-flattening approximation according to Schwab & Knopoff (1972). Parameters of the inversion are coefficients of basis functions into which the velocity perturbations of the background model are developed. Following Nolet (1990) the basis functions are box car functions and triangles. Below the crust, the velocity in 3 to approximately 10 individual layers might be varied by a single basis function. Furthermore, the perturbation of the Moho depth is a parameter of the inversion. The total number of inversion parameters

Figure 6. Synthetic test of the determination of phase velocities of the fundamental Rayleigh mode using the two-station method. The black lines denote average phase velocities and their standard deviations for the two paths GVD–SKD and KRIS–SKD and the events considered in Fig. 5. Synthetic waveforms were perturbed by Gaussian white noise with a standard deviation corresponding to 5 per cent of the maximum amplitude of the waveform. For comparison, the grey lines show results for synthetics without noise.

Figure 7. Measured average phase velocity curves (black solid line) and standard deviation (black dashed line). Theoretical phase velocity curves for the final 1-D models of Fig. 7 (thick grey line).
Figure 8. Final 1-D models obtained by the inversion of the measured average phase-velocity curves (thick black lines). Initial 1-D models (thin black lines). For the path MARJ–GVD, the final 1-D model is compared with shear wave velocities expected for oceanic lithosphere older than 110 Myr (thick black dots between 30 and 100 km depth). The filled grey polygons show the results of the resolution analysis indicating maximum acceptable perturbations of the final model.

Figure 9. Grid search for the optimal model of the path KRIS–SKD regarding the two parameters depth of the oceanic Moho and average shear wave velocity above the Moho. Rms-values of the misfit smaller than one are plotted. Contour lines denote 0.05 intervals of the rms-misfit. The misfit is the difference between the measured and theoretical phase-velocity curves weighted by the standard deviation of the measured curve.

The parameters are the average velocity above the Moho and the Moho depth, respectively. For different combinations of these two parameters the rms value of the misfit is determined. The misfit is calculated for the entire dispersion curve though perturbations of the parameters considered only partly vary the phase velocity curve. The rms-error is plotted for values smaller than one. From this plot the average depth of the oceanic Moho is estimated to be approximately 46 km. The average velocity above the Moho is approximately 3.5 km s$^{-1}$. The error function is a relatively simple one. This shows that a linearized inversion scheme may be applied. The grid search validates the findings of the gradient search (Fig. 8).

The two-station method is based on the great-circle approximation. The fundamental mode is assumed to propagate along a great circle through epicentre and station and the phase is assumed to represent the integral over the local wavenumber along the path. The great-circle approximation is valid for smooth and long-wavelength
lateral heterogeneity. Estimates of the phase velocity by the two-station method are too fast if the direction of wave propagation and the great circle between the stations form an angle larger than expected from the great-circle approximation. If this deviation amounts to 7° the phase velocity is too fast by a factor of 1.008. For 15° the factor is 1.035. Multipathing might cause additional errors in the estimated phase velocity curves. Furthermore, strong lateral heterogeneity will cause finite-frequency effects due to the long wavelengths of Rayleigh waves and deviations from the great-circle approximation (e.g. Wielandt 1993; Friederich & Wielandt 1995; Meier et al. 1997; Marquering et al. 1998; Friederich 1999; Yoshizawa & Kennett 2002). In this case, the phase of the fundamental mode deviates from the integral over the local wavenumber along an estimated path and the phase velocities determined by the two-station method may be biased. This effect is expected especially for paths that run parallel to strong lateral heterogeneity e.g. KRIS–SKD. Meier et al. (1997) detected anomalous waveforms of the fundamental Rayleigh mode for a path parallel to the Philippine trench. They were explained by conversion of the fundamental Love mode to the fundamental Rayleigh mode.

In this study, phase velocities for many events are averaged to reduce the influence of lateral heterogeneity along the path from the epicentres to the stations. Results of the inversion are interpreted as estimates of the average structure on the path. Results of the resolution analysis and hints for deviations from the great-circle approximation are taken into account when interpreting the results.

4 1-D MODELS OF THE SHEAR-WAVE VELOCITY

The models in Fig. 8 exhibit considerable lateral heterogeneity along the profile. The models are interpreted with regard to average properties of the deeper crust and of the upper mantle down to a depth of approximately 200 km. In the frequency range considered, the phase velocity curves are strongly sensitive to the depth of the Moho.

4.1 The slab

The models for the paths GVD–SKD, KRIS–SKD, SKD–SANT show significantly increased velocities in the upper mantle. They are interpreted as high velocities within the subducting African oceanic mantle lithosphere. According to the resolution analysis the increased velocities in the final models are significant. Furthermore, on the paths GVD–SKD and KRIS–SKD a discontinuity is observed due to a significant increase of the phase velocities between approximately 15 and 40 mHz. For the path GVD–SKD a discontinuity is observed at a depth of approximately 40 km. It is interpreted as oceanic Moho within the African lithosphere. Crustal velocities above this discontinuity are due to the African oceanic crust and the Eurasian continental crust. In the model KRIS–SKD the oceanic Moho is found to be deeper at a depth of slightly lower than 50 km. The resolution analysis indicates possible errors in the estimation of the Moho depth of approximately 5 km. Therefore, the observed difference in the Moho depth between the two paths is significant. The high-velocity slab is also imaged in the model SKD–SANT for the path between western Crete and Santorini, but deeper than in the model KRIS–SKD. It follows that the slab is dipping towards the NE as expected. That means enlarged velocities due to a subducting slab may be imaged by the two-station method for paths similar to those considered in this study.

4.2 From Santorini towards central Turkey

The model SANT–ISP shows a Moho at approximately 40 km depth and no significant anomalies in the upper mantle probably because this long path crosses different tectonic units in the Aegean Sea and in western Turkey. The Moho along the path ISP–ANTO crossing central Turkey is poorly resolved. Anomalous low velocities are notable in the upper mantle down to a depth of 130 km. They might be associated with low densities causing the high topography in central Turkey. Similar results are obtained e.g. by Marone et al. (2003).

4.3 The path MARJ–GVD crossing the Libyan sea

The path MARJ–GVD runs from northern Africa to the island Gavdos south of Crete. Between 50 and 150 km depth it shows upper mantle velocities slightly below 4.5 km s⁻¹. This finding is confirmed by surface wave tomography (Marone et al. 2003) Oceanic lithosphere just north of the African passive margin is expected to be older than 110 Myr (e.g. Dercourt et al. 1986). In Fig. 8 the model for the path MARJ–GVD is compared with theoretical calculations of the S-wave velocity for 110 Myr old oceanic lithosphere (Stein & Stein 1996). The observed velocities are significantly different from the theoretical values for 110 Myr old oceanic lithosphere. A thick oceanic lithosphere that is due to cooling of lithosphere that was produced by seafloor spreading at mid-ocean ridges is missing. That means that this path crossing the Libyan Sea is not dominated by oceanic lithosphere and the present coastline of northern Africa is obviously not associated with a sharp transition from continental lithosphere to old and cold oceanic lithosphere. If this were the case, velocities above 4.5 km s⁻¹ would be expected in the upper mantle. Velocities lower than 4.5 km s⁻¹ point to a transition between pure continental and pure oceanic mantle lithosphere as it is expected for a passive continental margin. For west Africa the transition between continental and oceanic lithosphere was found by Marillier & Mueller (1982) to be approximately 500–600 km wide. A passive continental margin is formed during continental rifting and breakup. Properties of a passive margin are influenced e.g. by extension during rifting, sedimentation and magmatic intrusions.

Because of the low velocities observed for this path it might by estimated that at most one quarter of the path crosses oceanic mantle lithosphere with high S-wave velocities. Therefore, the southern border of oceanic lithosphere is assumed to be situated not more than approximately 100 km SW of Gavdos. This means, that the African oceanic lithosphere is almost completely subducted. It follows that the active continental margin of the Hellenic subduction zone retreated back to the passive margin of northern Africa.

It is interesting to note that differences in the structure of the crust and the uppermost mantle between the eastern and western Mediterranean have been discussed since the early studies of Payo (1967, 1969). A continental character of the lithosphere was assumed for the eastern Mediterranean, while an oceanic character was detected for the western Mediterranean. Here we interpret the low velocities along the path discussed as hints for a passive margin. Details concerning the internal lithospheric structure e.g. the question of whether or not this is a volcanic or non-volcanic margin have yet to be resolved.

4.4 The path KRIS–SKD on Crete

The determination of the phase velocity curves by the two-station method assumes smooth and long-wavelength lateral
heterogeneity. Since the path KRIS–SKD runs arc-parallel, the propagation of surface waves is influenced by strong lateral heterogeneity due to the NE-dipping slab. In Fig. 7 the average phase velocity curve for the path KRIS–SKD is shown together with the estimated standard deviation and the phase velocity curve of the final model. Between 10 and 30 mHz the phase velocity curve of the final model is systematically slower than the measured phase velocity curve. This is because the damping was chosen in such a way that the velocities within the slab are not unrealistically high. The difference between the phase velocity curve of the final model and the measured phase velocity curve shows that for this path and in the discussed frequency range phase velocities measured by the two-station method are biased by approximately 0.2 km s⁻¹ towards larger values. The amount of the bias corresponds to the standard deviation of the measured phase velocity. This bias indicates a deviation of the propagation of fundamental mode in subduction zones from the great-circle approximation. That means anomalous propagation of the fundamental Rayleigh mode is detected for this path.

It is assumed that the deviation from the great-circle approximation is strongly influenced by finite-frequency effects of surface wave propagation. It is therefore concluded that the wavefield in subduction zones should be studied by a dense arrays of broad-band stations that would be capable to reveal details of the wave propagation. Modelling of surface wave propagation in 3-D media is needed to understand the causes of the observed anomalous propagation of the fundamental Rayleigh mode.

Owing to the complicated propagation of the fundamental Rayleigh mode found for the path KRIS–SKD, the velocity structure within the slab cannot be determined by the two-station method. However, the average depth of the oceanic Moho may be estimated because it causes a change in the phase velocity of approximately 1 km s⁻¹ which is significantly larger than the observed bias. The estimate of the Moho depth might partly be influenced by the bias of the measured phase velocity curve. Therefore, the estimate of approximately 45 km depth for the oceanic Moho might be slightly too small. An average depth of approximately 50 km is supported by receiver function studies (Knapmeyer & Harjes 2000; Li et al. 2001). Low velocities above this discontinuity are due to the African oceanic crust and the continental Aegean crust. If the Aegean continental Moho is assumed between 30 and 35 km depth as indicated by a wide angle seismic study (Bohnhoff et al. 2001), additionally an approximately 10 km thick wedge of Aegean mantle lithosphere might be present above the African plate.

5 TECTONIC INTERPRETATION

5.1 A tectonic model of the lithosphere

In the following a model of the lithosphere is assumed to be based on the results presented in Fig. 8, together with a tectonic interpretation. Fig. 10 (bottom) summarizes the present lithospheric structure of the Hellenic subduction zone on a SW–NE profile from northern Africa towards the Cretan sea. This is a schematic cross-section that shows the main properties of the lithosphere imaged by the models in Fig. 8. The vertical and horizontal scales are equivalent. Beneath Crete subducted African oceanic lithosphere is indicated. Above the slab a mantle wedge with relative low velocities may be found. The Eurasian continental crust is thickened beneath Crete and is thinned towards the Cretan Sea (e.g. Bohnhoff et al. 2001). North of the African coastline the passive African continental margin can be found. Details of its internal structure are unknown. It consists probably of extended African lithosphere that may be altered by intrusions.

Tectonic reconstructions allow conclusions to be drawn on the geometry of the lithospheric plates at the surface and at different times. In this study, the tectonic reconstructions along the considered profile are extrapolated into depth in such a way that the present lithospheric structure may be explained. The general tectonic evolution between 130 and 45 Ma was adopted from Dercourt et al. (1986) and Gealey (1988). For the last 45 Myr Fig. 10 is based on the model suggested by Thomson et al. (1998) and Stöckhert (1999). In Fig. 10 oceanic mantle lithosphere is distinguished from continental lithosphere. Oceanic lithosphere shows thin crust, while continental lithosphere has either thinned, normal or thickened crust. The thickness of the lithosphere is varied accordingly.

The continental crust underlying Crete is assumed to have been part of the extended lithosphere of northern Africa since the Jurassic (e.g. Dercourt et al. 1986). The Neotethys was located north of Africa. In the earliest Cretaceous, at the latest, two microcontinents were rifted off northern Africa and drifted northwards towards Eurasia, with the new oceanic realm between these microcontinents referred to as Pindos and that between the southern microcontinent and present northern Africa referred to as Mesogea. The continental crust beneath Crete is interpreted to represent the southern of the two microcontinents (Thomson et al. 1998; Stöckhert 1999). After the subduction of the Neotethys oceanic lithosphere beneath the southwestern margin of Eurasia, the northern microcontinent became accreted to Eurasia (e.g. Gealey 1988) and subduction of the Pindos ocean commenced. The southern microcontinent entered the subduction zone at the southern margin of Eurasia approximately 50 Ma. Some sedimentary cover of this microcontinent was pulled down to depths of up to ca. 35 km during collision between 25 and 20 Ma. This is a hint for a thickening of the crust at the southern Eurasian margin at this time and for a coupling between the subducting mantle lithosphere and the crust of the second microcontinent. Around 20 Ma subduction of the oceanic lithosphere of the Mesogea started south of the southern microcontinent. Gravitational retreat of the subducted slab caused extension of the southern Eurasian margin. The thermochronometric record of the sedimentary cover of the second microcontinent indicates that parts of the crust of the microcontinent detached from the subducting lithospheric mantle after collision. It returned to upper crustal depths at between 20 and 15 Ma, filling the space created by continuous roll back of the subduction zone (Thomson et al. 1998; Stöckhert 1999). The former sedimentary cover of the second microcontinent is now exposed on Crete, while its continental basement probably constitutes the upper continental crust beneath Crete and continental crust south of Crete.

After accretion of the southern microcontinent, subduction of the Mesogea oceanic realm continued to the present day. Today, the Mesogean oceanic lithosphere is almost completely subducted and the passive continental margin of northern Africa is about to enter the subduction zone. This conclusion is based, for example, on the observation by LePichon et al. (1995) that the external deformation front south of the Mediterranean ridge is deformed due to the collision of continental Africa with the accretory complex (Fig. 1). Mascle et al. (1999) studied the structure of the Mediterranean ridge south of Crete and found evidence for incipient collision. Furthermore, the absence of N–S extension in the Hellenic arc might be an indication of the beginning collision of Eurasia and northern Africa (Lyon-Caen et al. 1988). The view that the Aegean plate is about to collide with the passive continental margin of northern Africa is further supported by findings of his study namely that the mantle lithosphere on the path between MARJ and GVD crossing the
Rayleigh waves in eastern Mediterranean

130 Ma
Jurassic–Cretaceous

110 Ma
Cretaceous (Aptian)

80 Ma
Cretaceous (Campanian)

65 Ma
Cretaceous–Paleogene

35 Ma
Paleogene (Eocene–Oligocene)

20 Ma
Paleogene–Neogene (Oligocene–Miocene)

10 Ma
Neogene (Miocene, Pliocene)

0 Ma

Figure 10. Tectonic model of the lithosphere on a NE–SW profile crossing Crete from the Late Cretaceous to the present. Light grey symbolizes crust, deep grey continental mantle lithosphere, black oceanic mantle lithosphere. The hatched regions indicate transitions between oceanic and continental mantle lithosphere. The following abbreviations are adopted: NA, Northern Africa; C, Crete; NT, Neotethys; M, Mesogea (Libyan Sea); P, Pindos ocean; E, Eurasia; DH, Dinaro–Hellenic chain.

Libyan Sea shows shear wave velocities below 4.5 km s\(^{-1}\). While the forearc of the Hellenic subduction zone is a tectonically active region, the African passive continental margin is at least 110 Myr old.

The essential points of the tectonic model of the lithosphere along the profile are: while the crust of the two microcontinents was accreted to the active southern margin of Eurasia after subduction of the intervening oceanic realms, the underlying mantle lithosphere forms parts of a continuous slab that was imaged by seismic studies of the mantle. This subducted continuous slab comprises both oceanic and continental sections, the approximate present location of which can be inferred from the original width of the subducted oceanic realms and the accreted microcontinents.

5.2 Length of the slab

In the model presented, the length of the subducted oceanic lithosphere of Mesogea is assumed to be approximately 550 km. First estimates of the length of the slab were provided by LePichon & Angelier (1979). They assumed that the depth extent of the slab is indicated by, for example, the maximum depth of the intermediate depth seismicity and that the subduction started approximately 10 Ma. They concluded that the slab is a short one. Spakman et al. (1993, 1988), Meulenkamp et al. (1988) and Wortel et al. (1990) found that the slab is significantly longer than previously assumed and that the subduction consequently started earlier, at least 26 Ma. Results of seismic tomography (e.g. Spakman et al. 1993; Karason & van der Hilst 2000) show that the slab penetrates into the lower mantle down to a depth of at least 1200 km.

An independent estimate of the length of the Mesogean slab may be obtained if the interception of the subduction is taken from petrological evidence and if the kinematics in the eastern Mediterranean is taken into account. The subduction of the Mesogea started approximately 20 Ma (Thomson et al. 1998; Stöckhert 1999). Since then Africa has moved towards stable Eurasia with a relative horizontal velocity of approximately 1 cm a\(^{-1}\) (e.g.
Dercourt et al. 1986). This yields approximately 200 km of subducted slab.

In addition, the southward migration of Crete as part of the forearc relative to Eurasia, for example, indicates that more than 200 km of African lithosphere was subducted. The kinematics of the Anatolian–Aegean plate can be described by a counterclockwise rigid rotation and internal extension (e.g. McClusky et al. 2000). The average velocity of the horizontal displacement of the Anatolian–Aegean plate in respect to Eurasia was estimated by LePichon et al. (1995) to be approximately 3 cm a\(^{-1}\) since the beginning of the extrusion of the Anatolian–Aegean approximately 10 Ma. This results in a relative horizontal displacement along the North Anatolian fault of approximately 300 km which corresponds to a rotation of the plate of approximately 20° around the Euler pole of rotation determined by McClusky et al. (2000). To estimate the resulting contribution to the length of the subducted slab it is necessary to account for the angle between the movement of the Anatolian–Aegean plate and the direction of subduction. Assuming that the counterclockwise rotation indicates a movement of the active continental margin from approximately 10 Ma. This results in a relative horizontal displacement along the North Anatolian fault of approximately 300 km which corresponds to a rotation of the plate of approximately 20° around the Euler pole of rotation determined by McClusky et al. (2000).

At present, the internal extension of the Aegean plate due to roll back, amounts to \(\sim 1\) cm a\(^{-1}\) in a SW direction (LePichon et al. 1995; McClusky et al. 2000). If this extension is assumed to represent an average value for the last 20 Myr, another 200 km of African lithosphere had to be subducted. Meulenkamp et al. (1988) estimated that the internal extension and the rotation together indicate approximately 400 km of subducted slab. ten Veen & Meijer (1998) and ten Veen & Kleinspehn (2003) assume a southward movement of Crete within the last 11 Myr of slightly more than approximately 300 km based on numerical modelling, geodetic and palaeomagnetic data. These estimates are comparable to that given above. According to our estimates the total maximum length of the Mesogean oceanic lithosphere subducted south of Crete is approximately 550 km.

This estimate is in contrast to the length of the slab deduced from global tomography. Following the model in Fig. 10, the slab imaged by seismic studies down to approximately 400 km depth consists of oceanic lithosphere of Mesogea. That means enlarged mantle velocities in Fig. 8 are due to oceanic lithosphere of Mesogea. According to the model presented in Fig. 10, between 400 and 660 km depth mantle lithosphere of the southern microcontinent and lithosphere of the Pindos ocean may be imaged. The causes for the missing deep seismicity (Fig. 1) may thus be found in the subduction history and the composition of the oceanic lithosphere of the Pindos ocean and the continental mantle lithosphere of the southern microcontinent. Below 660 km high velocities in the tomographic cross-sections are probably due to the subducted oceanic lithosphere of the Neotethys. Continental mantle lithosphere of the northern microcontinent may also form a part of the slab. This means that the length of the Mesogean slab is relatively short, while the longer slab imaged by structural investigations of the mantle consists of oceanic lithosphere of several ocean basins.

6 CONCLUSIONS

Phase velocity curves of the fundamental Rayleigh mode determined by the two-station method may be inverted for 1-D models of shear wave velocity. For the considered path, the comparison of the resulting 1-D models allows conclusions to be drawn on the structure and tectonics of the lithosphere and the upper mantle down to a depth of approximately 200 km. Results of the two-station method might thus supplement tomographic studies at regional and global scales and yield additional constraints on the local average structure between neighbouring stations. In this study, the subducting African oceanic lithosphere was imaged and the depth of the oceanic Moho beneath Crete was estimated by the two-station method. In the case of smooth lateral heterogeneity the 1-D models may be interpreted as average models of the structure between the stations.

Strong lateral heterogeneity might cause a bias of the phase velocity estimated by the two-station method. For the path KRIS–SKD anomalously propagated of the fundamental Rayleigh mode is detected: the phase velocities between 10 and 30 mHz are biased by approximately 0.2 km s\(^{-1}\). In this frequency range measured phase velocities are systematically faster than expected. Therefore, details of the velocity distribution within the dipping slab cannot be determined using a two-station method. A dense array of broadband stations is needed to investigate the details of the structure within the slab.

However, the average depth of the oceanic Moho beneath Crete can be estimated at approximately 50 km depth with an accuracy of approximately 5 km. Above this discontinuity the average S-wave velocity is approximately 3.5 km s\(^{-1}\) which corresponds to crustal velocities. A \(\sim 10\) km thick mantle wedge can be present between the African oceanic crust and the Eurasian continental crust. Anomalously low velocities are found in the uppermost mantle beneath central Turkey down to a depth of 130 km. The S-wave velocity model for the path from the Libyan coast towards the island Gavdos also shows relatively low velocities. This indicates that the African oceanic lithosphere is almost completely subducted and that the active margin of the Hellenic subduction zone approaches the passive African continental margin. The internal structure of the passive African continental margin remains unknown.

A tectonic model of the lithosphere along a NE–SW profile across Crete is suggested that indicates accretion of continental crust of two microcontinents to Eurasia and progressive subduction of oceanic lithosphere of the Neotethys, the Pindos ocean and the Mesogea. Mantle lithosphere of the microcontinents was detached from the overlying crust and subducted. This could explain the continuous slab that is found in tomographic images of the Hellenic subduction zone down to a depth of approximately 1200 km. While the mantle lithosphere of the southern microcontinent was subducted, the continental crust of the southern microcontinent was detached after collision and is now found in the continental crust south of Crete and beneath Crete. The length of the oceanic lithosphere of Mesogea that was subducted south of Crete is estimated to be not greater than approximately 550 km.

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