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The crustal structure in Iberia inferred from P-wave coda

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ABSTRACT

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The crustal velocity-structure at four locations in the Iberian Peninsula is inferred from P-wave coda recorded by broad band stations. The crustal responses of the stations are derived from the P-wave coda of deep, teleseismic events, and the most pronounced features of the P and S velocity-structure are obtained by waveform modeling of the vertical and radial component of these responses.

The method is applied to data of NARS stations in Toledo, Alicante, and Granada in Spain, and Manteigas in Portugal. The optimum models of the waveform inversions are carefully evaluated, and compared to results of deep seismic sounding experiments. In general, the depths of large velocity gradients are well-resolved, but the actual values of the seismic velocities are less constrained.

Introduction

The crustal structure of Iberian Peninsula can be described as a Hercynian core surrounded by areas that have undergone Alpine deformation. Knowledge of both the crustal and upper mantle structure of Iberia is relevant for a better understanding of geodynamics of the region, such as the evolution of the Hercynian orogen in central Spain, or the complex tectonics of the Pyrenees or the Betic Cordillera. The Iberian Lithosphere Heterogeneity and Anisotropy (ILIHA) project was initiated from this perspective. Its aim is to improve our insight in the evolution of the Iberian lithosphere through seismological investigations and deep seismic refraction experiments. Thirteen stations of the portable Network of Autonomously Recording Stations (NARS) were deployed in Spain and Portugal for a period of nearly one year (1988–1989) as part of this project (see Fig. 1). The data of this broad band 3-component network are suitable for surface-wave as well as body-wave studies, which enables a variety of different experiments (e.g., Alvarez and Herraiz, 1990; Maupin, 1990; Paulssen et al., 1990; Payo et al., 1990).

The objective of the present study is to determine the most pronounced features of the crustal P and S velocity-structure beneath the NARS–ILIHA stations. The method that is employed for this purpose is based on a non-linear waveform inversion of the crustal responses of the stations. In this paper we present a brief description of the method, show data and results for 4 of the NARS–ILIHA stations, and compare the results with findings from deep seismic sounding studies.

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Method and data selection

For a detailed description of the procedure to determine the receiver response and of the non-linear inversion we refer to Paulssen et al. (1992); here we only present a brief outline of the method. The crustal receiver response is defined as the response of all multiples and converted phases that are generated in the crust near the receiver due to a plane P-wave incident from the mantle. We estimate the crustal response from the P-wave coda of teleseismic events by deconvolving the seismograms with an empirically determined "source wavelet" for each event. This wavelet is defined to include source–time function, effects from source and mantle, and the instrument response of the station. For each event, the wavelet is selected by visual inspection of the vertical component (NARS) recordings as the common wavelet representative of the direct P-wave signal. Evidently, such a wavelet can only be determined for events that have a short and simple source–time function. It is eliminated from the vertical and radial component of the seismogram by the deconvolution technique of Langston (1979) which includes an additional Gaussian filter.

The crustal P and S velocity-structure beneath a station is determined by modeling the (P–SV) crustal responses of that station. We use the propagator matrix method (Haskell, 1962; see e.g., Aki and Richards, 1980) and a non-linear inversion scheme (Nolet, 1987) to find the optimum plane layered model that explains the data. The crustal structure is described by the P velocity α, the S velocity β, the density ρ, and the thickness of each of the layers. The upper mantle parameters were kept fixed to $\alpha = 8.1\ \mathrm{km \cdot s^{-1}}$, $\beta = 4.3\ \mathrm{km \cdot s^{-1}}$, and $\rho = 3.3\ \mathrm{g \cdot cm^{-3}}$.

If several seismograms are available from a certain source region, it is favourable to stack their responses in order to enhance the signal-to-noise ratio. Stacking has the additional advantage that a measure of the variance of the data can be obtained.

As mentioned before, only carefully selected events can be used for the determination of the crustal response of a station. First of all, the source–time function of the event should be short and clear. If the source function is too long, then

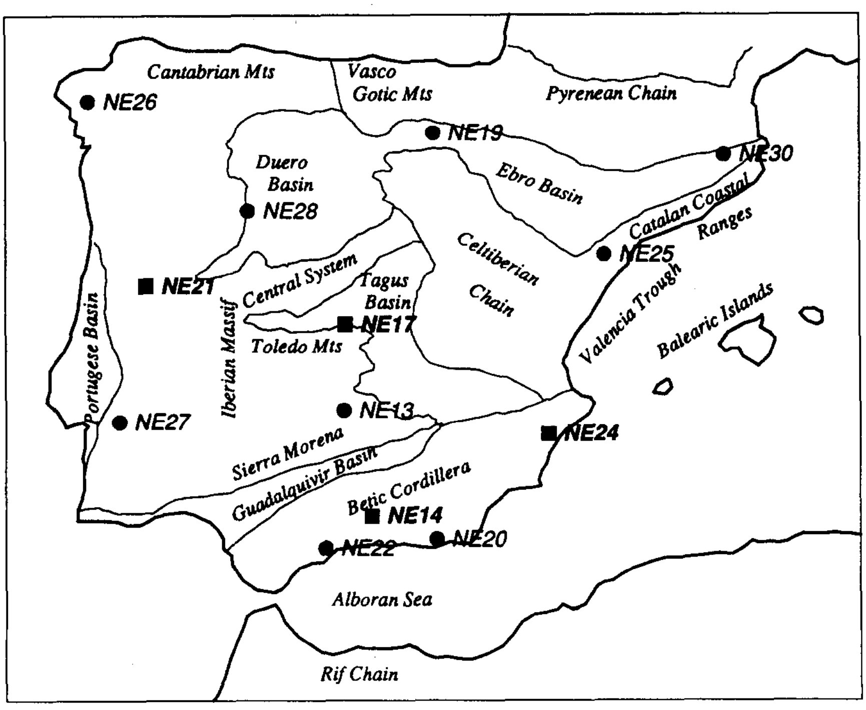

Fig. 1. Main tectonic units of the Iberian Peninsula and locations of the NARS (ILIHA) stations. Squares indicate the stations used in this study.

part of the crustal response may be included, and crustal information will be eliminated in the deconvolution procedure. Secondly, the seismograms must have a good signal-to-noise ratio because the phases that are generated in the crust are generally small in amplitude. Lastly, no primary phases (e.g., pP or PP) should arrive in the time window of interest. This implies that only deep, teleseismic events can be used. We only model the first 15 s of the response, because the most prominent crustal phases (first order multiples and converted phases) arrive in this time interval.

At this point we wish to emphasize that it is difficult, or practically impossible, to give a formal estimate of the resolution of the model that is obtained. In absence of noise and for a plane layered structure, the inversion will end at the actual velocity-structure. To avoid convergence to local minima in the model space, we developed an inversion strategy that involves a step-by-step inversion procedure where the number of free parameters gradually increases (see Paulssen et al., 1993). Convergence is considered stable if different starting models converge to the same final model. In absence of noise, the uncertainty

TABLE 1

Events used for the determination of the crustal responses

Response	Event code	Δ (°)	Backazimuth (°)	Depth (km)	Phase velocity ($km \cdot s^{-1}$)	Average phase velocity
Event parameters for station NE17						
1	880706	57.5	67.5	194	16.09	16.2
	870505	57.2	67.7	211	16.19	
	860426	57.5	67.4	187	16.21	
2	880920	74.2	−97.1	38	18.70	20.1
	861123	79.5	−103.3	126	20.64	
	860326	777.8	−110.1	609	20.88	
Event parameters for station NE21						
1	880806	59.7	66.2	194	16.41	16.4
2	880820	76.1	66.78	71	19.15	19.2
3	880728	82.7	−127.5.78	286	21.77	21.8
Event parameters for station NE24						
1	880806	55.5	68.1	194	15.78	15.8
2	880820	71.8	70.0	71	18.24	18.2
3	880920	76.9	−94.4	38	19.34	19.3
Event parameters for station NE14						
1	850501	78.2	−111.1	600	19.74	20.2
	861123	79.3	−102.5	126	20.55	
2	870401	84.0	−124.6	224	22.50	23.0
	851031	86.2	−130.6	595	23.46	
3	830912	58.2	66.0	208	16.33	16.3
	831230	58.1	66.4	222	16.31	
4	860618	91.3	−4.1	61	23.98	24.3
	841203	95.0	19.9	70	24.48	

in the model parameters is determined by the criterion at which the inversion is stopped (in our case when the norm is improved by less than 2% in the subsequent iteration). In practice, however, the data are contaminated by noise, and this may severely limit the resolution that can be achieved. A perfect fit between data and synthetics would then imply a projection of noise onto the model. It is clear that the accuracy of the model parameters is limited by the level of misfit due to noise. All models that give a misfit less than (or equal to) the misfit determined by the noise have an acceptable fit to the data. Unfortunately, in many cases we do not have a formal estimate of the noise, which makes it impossible to present formal error bars of the model parameters. We have

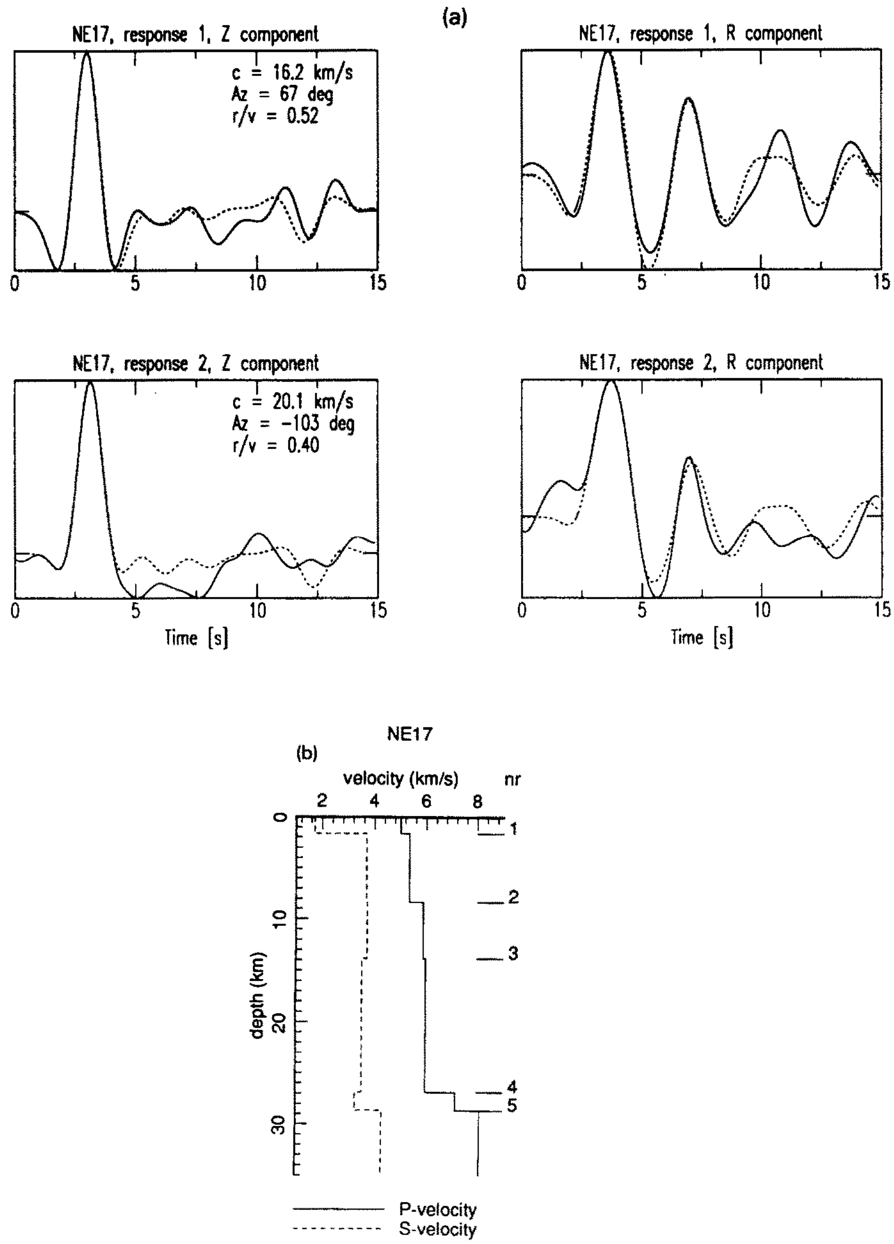

Fig. 2. (a) Observed (solid) and synthetic (dashed) crustal responses of station NE17. Phase velocity, (back)azimuth, and amplitude ratio of radial to vertical component are given in the upper right corner of the vertical component. (b) Final model of the inversion for which the synthetics are shown in (a).

tried to obtain an estimate of the sensitivity of the fit on the various model parameters by perturbing the final model parameters and looking how these variations influence the match between data and synthetics. In the description of the models below we indicate which features of the models are robust, and which elements are thought to be not very well-resolved.

Table 1 lists the events that were used for the determination of the crustal responses. Only for a limited number of stations we obtained sufficient good quality data, as the period of data acquisition was rather short for such an experiment which requires a strong data selection. Fortunately, we were able to include data of the previous NARS project (Nolet et al., 1986; Dost, 1987) for the stations NE14 and NE17. For station NE13 we obtained sufficient data, but we could not find a velocity-structure that adequately modeled the data of this station. This may be due to crustal heterogeneity in the vicinity of the station, or it might be an indication of errors in the crustal responses due to noise in the data.

In the following, we present data and results for NE17 and NE21, stations located on the crustal structure of the Iberian Massif, and for stations NE14 and NE24 in the Betic Cordillera. The locations of the stations and the main tectonic features of the Iberian Peninsula are shown in Figure 1.

The Iberian Massif

The Iberian Massif is part of the Hercynian belt of western Europe which outcrops in large areas of the Iberian Peninsula. Refraction experiments in central Spain (Banda et al., 1981; Suriñach and Vegas, 1988) have shown that the Hercynian crust of the Iberian Massif can be described by an upper crustal basement ($\alpha = 6.0$–6.1 km $\cdot$ s^{-1}), a low-velocity layer extending from about 7 to 11 km depth ($\alpha = 5.6$ km $\cdot$ s^{-1}), a mid-crustal layer to a depth of approximately 24 km ($\alpha = 6.4$ km $\cdot$ s^{-1}), and a lower crust ($\alpha = 6.9$ km $\cdot$ s^{-1}). The Moho appears to be a transition zone of at least 1.5 km thickness, located at a depth of about 31 km. The crust is locally overlain by a layer of Mesozoic sediments of varying thickness. A low (S) velocity layer at midcrustal depths is also found from Rayleigh wave measurements for the path between the WWSSN stations TOL and MAL (roughly corresponding to the path NE17–NE14 in Fig. 1; Payo, 1970, 1972; Badal et al., 1990). The mid-crustal low-velocity zone may correspond to a detachment level of late Hercynian extensional events, as was suggested by Casquet et al. (1988).

Station NE17 (Toledo, Spain)

Station NE17 (39.881°N, 4.049°W) is located in the Tagus Basin, where the Hercynian crust is overlain by Tertiary deposits. The crustal structure beneath this station is derived from two stacked crustal responses that are obtained from 3 seismograms each (Fig. 2a). The standard deviation of the signal is on average 10% of the normalized amplitude on each component. The waveform modeling results indicate that the most pronounced features of the data are well-explained by a 3-layer model consisting of a sedimentary cover, an upper, and a lower crust.

The signal generated at the sediment–basement interface has a strong influence on the two responses. The transition from the sedimentary cover (thickness of 1.5 km, $\beta = 1.7$–1.8 km $\cdot$ s^{-1}, α is not well-resolved: 5.0–5.5 km $\cdot$ s^{-1}) to the Hercynian basement rock causes an "apparent delay" of about 0.7 s of the first arrival on the radial component relative to the P-arrival on the vertical component. Such a delay was also observed for other NARS stations located on sediments (see Visser and Paulssen, 1993). The first strong arrival observed on the radial component is the P-to-S converted phase from the sediment–basement interface. This interface also produces the high-amplitude reverberatory signal on the radial component: it comprises its S-wave multiples.

The high-amplitude signal from the sediment–basement interface masks the signal from deeper structure, and, apart from the Moho transition, details in the S velocity-structure below this discontinuity are not well-resolved. The Moho is located at a depth of about 29 km, as is inferred from the first P-wave Moho multiple on the verti-

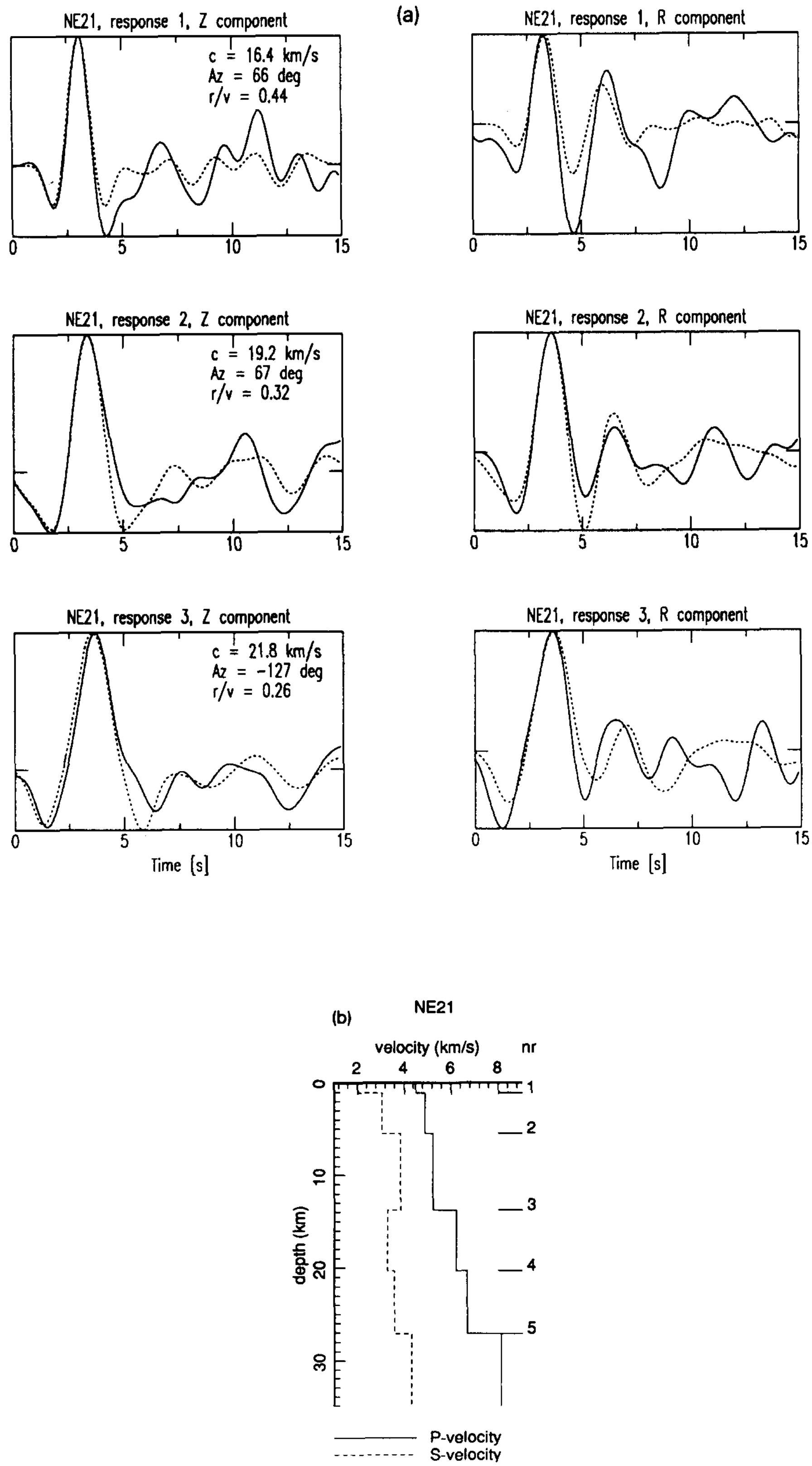

Fig. 3. Station NE21. For explanation see Figure 2.

cal component and the Moho P-to-S conversion on the radial component. Of the other features of the model, only the interface at a depth of approximately 8 km contributes to a consistent significant improvement of the fit between data and synthetics. The final model of the inversion is shown in Figure 2b.

If we compare our results with models obtained from refraction experiments of the Iberian Massif, we find that our inferred Moho depth is in agreement with these results (Banda et al., 1981; Suriñach and Vegas, 1988; ILIHA DSS Group, 1993-this issue). The average crustal P velocity of 6.0 $km \cdot s^{-1}$ is slightly lower than indicated by the refraction models. We do not find evidence for a low-velocity zone between 7 and 11 km depth with a marked velocity increase at its base as inferred from the refraction data: the response from the other interfaces may be too large to identify its signal, or the discontinuity may locally be less pronounced or even absent. Another striking observation is that there is no evidence for a strong discontinuity at a depth of approximately 24 km as suggested from the refraction surveys. Apparently, the signal from this interface is too small to be identified by the inversion. It can either be masked by the high amplitude signal of the other phases, or the discontinuity may locally be less pronounced than inferred from the refraction results.

Station NE21 (Manteigas, Portugal)

The crustal structure beneath station NE21 (Manteigas: 40.403°N, 7.537°W), located in the western part of the Iberian Massif, is derived from 3 crustal responses (Fig. 3a). They were each determined from one seismogram only, which implies that the signal-to-noise ratio is not as good as for station NE17.

Our inversion results indicate that the most prominent features of the crustal response of station NE21 can be explained by a 3-layer crust: an approximately 1 km thick upper layer, an upper crust to a depth of about 14 km, and a lower crust with the Moho at about 27 km. Small improvements to the fit are obtained by introducing additional layers and discontinuities at depths of 6 and 20–21 km, but in the model these features are less resolved. The final model of the inversion is shown in Figure 3b.

The P velocity is very poorly resolved. Velocity variations of about 0.4 $km \cdot s^{-1}$ degrade the fit only marginally. Moreover, there is a large misfit between data and synthetics on the vertical component of the responses. The P velocity model of Figure 3b may therefore not be a very reliable representation of the velocity-structure at NE21. The S velocity is better resolved, and the depths of the discontinuities is mainly determined by the depths of large velocity variations in the S velocity-structure. The low S velocity zone between 14 and 20 km is a feature that is required by the data as different starting models (without low-velocity zone) converge to similar models with a low S velocity zone. The S velocity decrease with depth at 14 km produces a P-to-S conversion that contributes (together with the S-wave multiple from the first interface) to the negative peak on the radial component at approximately 5 s.

Until recently, there were no refraction experiments carried out in the near vicinity of station NE21. Lines were shot towards the southwest across the Tagus valley (Mendes Victor et al., 1980) and along the coast (Moreira et al., 1980). Closer coverage was obtained through the ILIHA deep seismic sounding experiment: one of the lines sampled the crustal structure northeast of NE21 (ILIHA DSS Group, 1993-this issue). The results of this experiment indicate that the crystalline upper crust near NE21 can be divided in an upper layer, with an average P velocity of approximately 6 $km \cdot s^{-1}$, and a lower layer of approximately 6.3 $km \cdot s^{-1}$, starting at 14 km depth. The top and base of the lower crust are located at depths of 21–23 and 28–30 km, respectively, and the average velocity of the lower crust is approximately 6.8 $km \cdot s^{-1}$. Our average uppermost crustal P velocity is lower than inferred from the ILIHA refraction line, but the crustal velocity-structure beneath 14 km is in good agreement with the deep seismic sounding results.

Striking is the overall agreement between our S velocity-profile and the P velocity-structure found by Moreira et al. (1977, 1980) for refraction lines along the Portugese coast. These data show

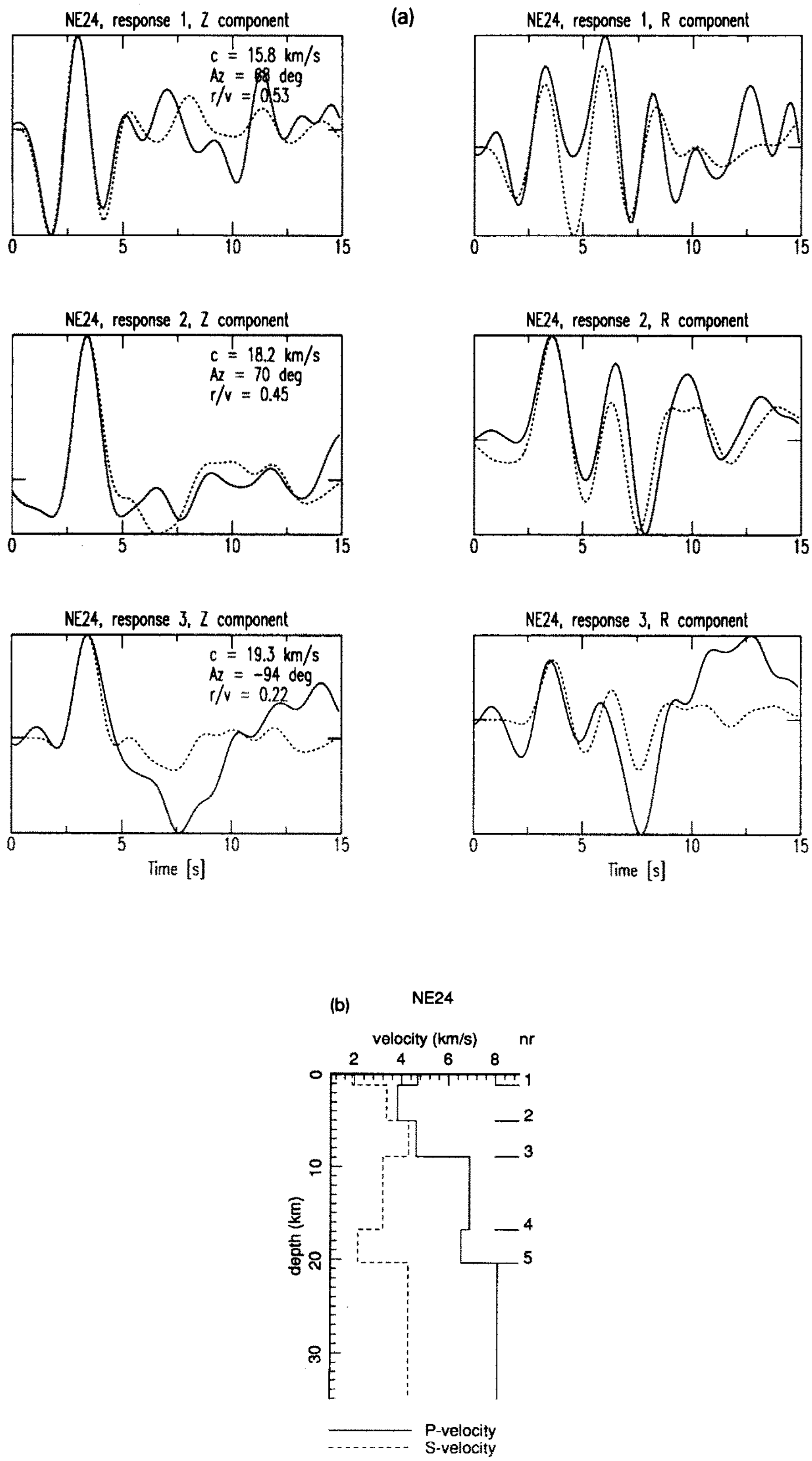

Fig. 4. Station NE24. For explanation see Figure 2.

evidence for a low P velocity zone at depths between roughly 10 and 20 km, which may correspond to our low S velocity layer. Moreira et al. (1980) suggested that this zone may represent a granitic layer of the Iberian Massif that collapsed at the end of the Paleozoic or beginning of the Mesozoic, when the opening of the Atlantic Ocean began.

The Betic Cordillera

The Betic Cordillera, in the south of the Iberian Peninsula, delineates the westernmost part of the Alpine chain in Europe. Together with the Alboran Sea and the Rif Mountains in northern Africa, it forms a tectonically complex zone that constitutes the boundary between the African and Eurasian plates. Refraction and surface-wave studies indicate that the crustal and upper mantle structure beneath the Betic Cordillera are highly heterogeneous (Working Group for Deep Seismic Sounding in Spain, 1974–1975, 1977; Banda and Ansorge, 1980; Paulssen et al., 1990). A thinned continental crust and anomalously low seismic mantle velocities have been found beneath the Alboran Sea (Working Group for Deep Seismic Sounding in the Alboran Sea, 1974, 1978; Marillier and Mueller, 1985). Despite the identification of distinct stages of Alpine deformation (e.g., Vegas and Banda, 1982), and a number of kinematic and geodynamic models that have been proposed for the area (e.g., Dewey et al., 1989; Platt and Vissers, 1989), the geodynamic evolution of the southern border of the Iberian Peninsula is still unclear.

NE24 (Alicante, Spain)

Station NE24 (38.355°N, 0.487°W) is located in the northeastern part of the (Pre-)Betics, where it merges with the extension of the Balearic Island chain. The crustal responses of this station are not of very high quality: they are each inferred from a single seismogram, and we have no estimate of the noise level in the data. The low-frequency signal of the third response (Fig. 4a) is obviously due to noise in the data, and, therefore, it was given a weight of 0.1 instead of 1.0 in the inversion. In spite of the noisy character of the responses, there are some features in the data that are consistent, and that give important information about the crustal structure at Alicante.

The data are best modeled by a thin (1 km) layer of unconsolidated sediments, a strong mid-crustal reflector at 9 km depth, and an anomalously shallow Moho. These features are well-determined, the other details of the model shown in Figure 4b give a smaller improvement to the fit between data and synthetics.

The relatively shallow Moho depth of 20 ± 2 km appeared to be a very stable result from the inversion: very different starting models all converged to a model with the Moho in this depth range. The Moho depth is mainly determined by the Moho P-to-S conversion which is identified as the high-amplitude arrival following the direct P-wave on the radial component. The values of P and S velocity for most of the layers are ill-constrained, and velocity variations of 0.4–0.8 $km \cdot s^{-1}$ degrade the fit only slightly. We therefore consider neither the P nor the S velocities very reliable. We have chosen for final inversion runs with uncoupled P and S velocity to obtain the optimum fit to the data. The combination of P and S velocity of the second and third layer of this inversion is obviously not realistic physically (negative Poisson's ratio). Because the P velocity-structure is mainly determined by the signal on the vertical component (crustal P-wave multiples), and the S velocity-structure by the signal on the radial component (P-to-S conversions), this indicates that there are inconsistencies in the data. However, since we do not know if the vertical component or the radial component is contaminated by noise, we have chosen not to couple the P and S velocities, as one of the two profiles may be more reliable than the other.

Although the velocities are ill-constrained, the character of responses indicates that there must be large velocity contrasts present beneath Alicante. The inversion results showed that velocity discontinuities are required at depths of approximately 5, 8–9, 17 and 20 km to explain most of the prominent features of the data.

If we compare our model with the crustal structure obtained from refraction experiments: for the Balearic Sea (Banda et al., 1980), and for the southwestern part of the Celtiberian Chain (Zeyen et al., 1985), we note that there is a good agreement between these studies and our investigation concerning the anomalously shallow Moho depth. Banda et al. (1980) located the Moho at a

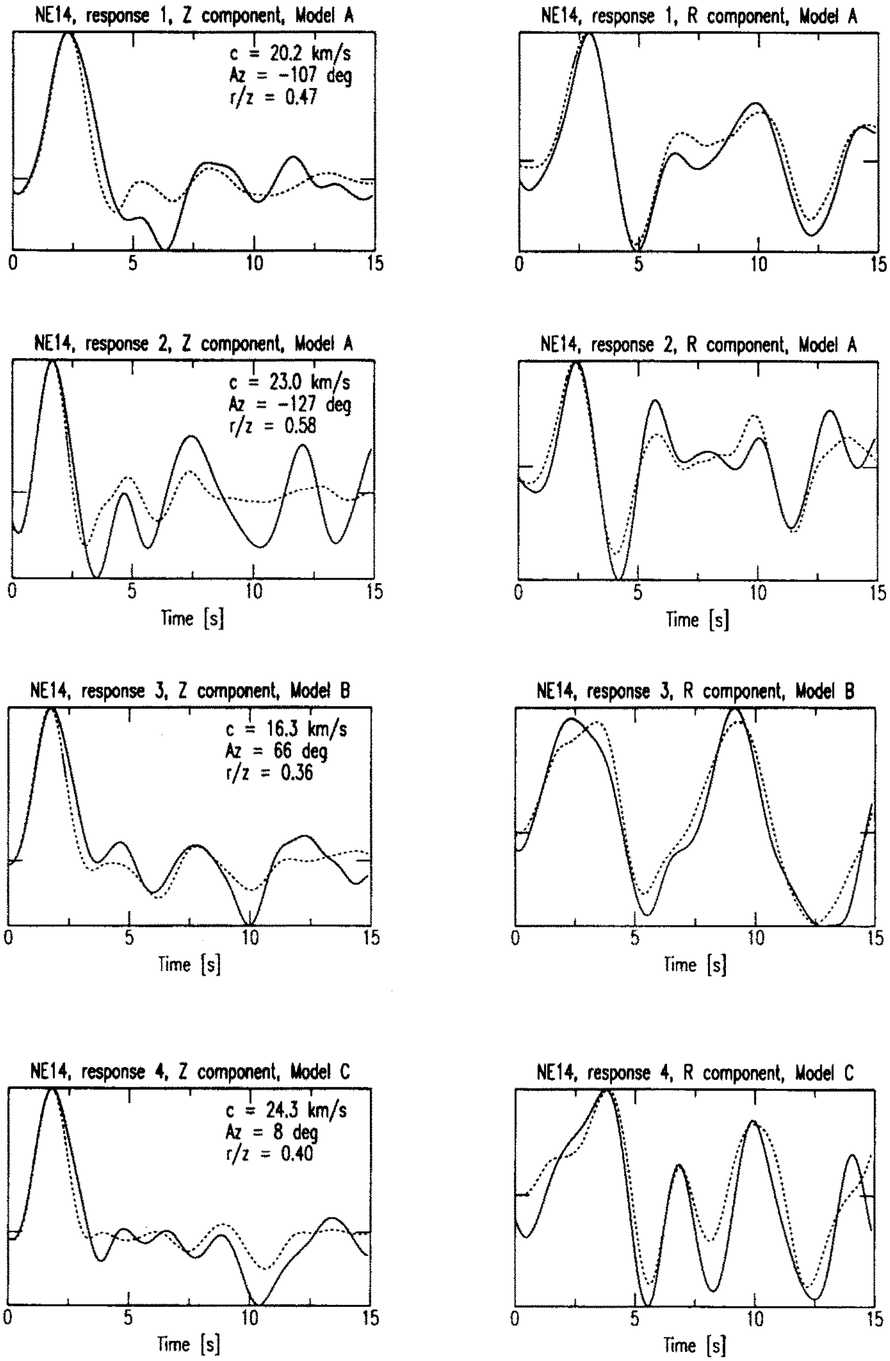

Fig. 5. Station NE14. For explanation see Figure 2. Responses 1 and 2 are modeled by model A, response 3 by model B, and response 4 by model C.

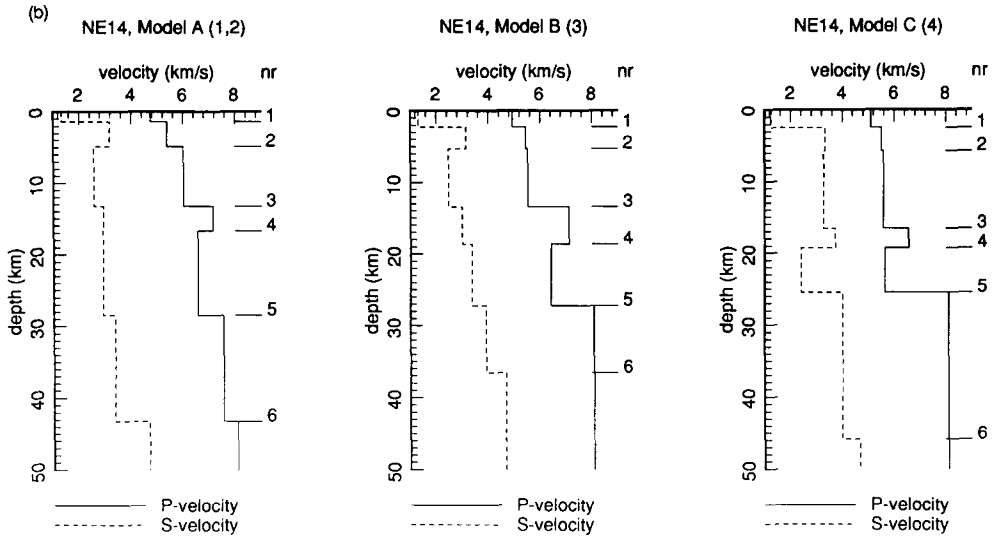

Fig. 5 (continued).

depth of 20 km for the western part of the Balearic Sea, and Zeyen et al. (1985) found a fast transition from normal continental Moho depths (30–32 km) beneath the Celtiberian Chain to depths of 20 km at the coast. They associated the thinning of the crust at the eastern border of the Iberian Peninsula to the opening of the Valencia Trough in Oligo-Miocene times. We do not elaborate on the other features of the model shown in Figure 4b, as the velocities are ill-constrained.

NE14 (Granada, Spain)

Station NE14 (37.190°N, 3.595°W) is located in the Betic Cordillera at the border of the Internal Zone and the External Zone. A detailed description of the inversion was given by Visser and Paulssen (1993); here we concentrate on the interpretation of the modeling results.

The four crustal responses of NE14 (Fig. 5a) that were used for the inversion were obtained by stacking the responses of two seismograms each. The standard deviation of the stacked signal is on average 10% of the normalized signal on each component.

The crustal structure at Granada is heterogeneous, because it was not possible to model the responses of NE14 adequately by a single velocity-structure. The two crustal responses obtained for westerly to southwesterly backazimuths (response 1 and 2) are best modeled by a layer of unconsolidated sediments of 1.4 km thickness, whereas the responses for northerly and northeasterly azimuths (response 3 and 4) require a thicker sedimentary cover of about 2.3 km (see Fig. 5b). This finding correlates with the tectonic map of the Betic Cordillera (by Julivert et al., 1972; see also Banda and Ansorge, 1980) that shows that Granada is located in a sedimentary basin with the sedimentary cover thinning towards the north and east.

A pronounced feature of the crustal structure at NE14 is a mid-crustal reflector at the base of a low S velocity layer. It is located at a depth of approximately 13 km for (south)western and (north)eastern azimuths (responses 1–3; models *A* and *B*), but modeled at a depth of 16 km for the crustal response with a northern azimuth (response 4; model *C*). This feature may be explained as a difference in the crustal structure between the Internal Zone of the Betic Cordillera in the south (sampled by responses 1–3) and the External Zone in the north (sampled by response 4). The low S velocity layer, which is more pronounced in the models *A* and *B* (Fig. 5b) than for the northern model *C*, corresponds to the low P velocity layer inferred from refraction experiments in the southern to central part of the Betic Cordillera (Banda and Ansorge, 1980). Striking is the agreement between the high P velocity layer and the prominent reflector found by Banda et al. (1993-this issue) in the Internal Betics. Banda et al. interpret the reflector at about 12 km depth as the base of the Nevado-Filábride complex and/or as a detachment surface.

The responses are dominated by effects of the upper crustal structure, so the velocity-structure below the third interface is less resolved. However, the large P velocity increase at about 27 km is a consistent feature of our models that is required by all four responses. A low P velocity zone bottoming at this depth (layer 5 in the models) is not found by the refraction data, although a strong velocity increase is located at this depth beneath the central to eastern Betics (Banda and Ansorge, 1980). This discontinuity may correspond to the Moho, but may also represent another major transition as the refraction data suggest that the Moho is found at greater depths beneath the central part of the Betics (ca. 39 km). Because it is an important feature of the crustal structure, it needs to be explained in tectonic models of the crustal structure of the Betic Cordillera.

Conclusions

The method of determining the crustal structure beneath a station by inversion of its receiver response is most sensitive to large velocity contrasts. These features produce high-amplitude multiples and converted phases that are modeled in the inversion are therefore best resolved in the models presented above. The actual values of the seismic velocities may not always be very accurate. For most models the S velocity-structure appears more reliable than the P velocity-structure because the (normalized) signal on the radial component is larger in amplitude than that on the vertical component.

Some of the most striking results from this study include the anomalously shallow Moho depth beneath NE24 (Alicante: ca. 20 km), and the good overall agreement between the models of NE21 (Manteigas) and NE14 (Granada) with their velocity-structures obtained from deep seismic sounding experiments. The inferred Moho depths agree well with those determined by refraction data, and other strong intracrustal velocity variations are generally also well-resolved. The method proves to be a good (complementary) alternative to other types of crustal studies, because the most pronounced features in the P and S velocity-structure can efficiently be determined from existing seismological data. Intracrustal low S velocity layers are generally well-resolved, and information on the existence of such layers may of be crucial importance for the tectonic interpretation of the structure and evolution of the crust.

Acknowledgements

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