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The crustal structure from teleseismic *P*-wave coda—II. Application to data of the NARS array in western Europe and comparison with deep seismic sounding data

Jeanette Visser* and Hanneke Paulssen

Department of Geophysics, Institute of Earth Sciences, P.O. Box 80021, 3508TA Utrecht, The Netherlands

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SUMMARY

The crustal structure beneath a seismic station has a large influence on the *P*-wave coda recorded by that station. In this study we employ the vertical and radial component of the crustal receiver response to determine the most important features of the crustal velocity structure beneath stations of the NARS array. The receiver response is estimated from the *P*-wave coda of teleseismic events by deconvolution with a source wavelet and by stacking responses of different events. The crustal velocity structure at the station is derived from these data by non-linear waveform inversion.

The responses of some of the NARS stations show anomalous features such as an 'apparent delay' of the first arrival on the radial component relative to the onset on the vertical component. This appears to be a combined effect of very low velocities in the top layer of the model and a strong velocity discontinuity in the uppermost part of the crust. A high amplitude coda on the radial component is observed for stations on a structure with strong *S*-velocity gradients in the upper crust. The receiver responses of the NARS stations are generally well modelled by the synthetics of the final models of the inversions. The method provides an adequate procedure to estimate the dominant effects of the crustal structure at the site with the models representing the most significant velocity gradients of the crustal structure.

Key words: body waves, crustal structure, Europe, inversion.

INTRODUCTION

Knowledge of the crustal structure under seismological stations is important to assess the effects of near receiver structure on the recorded seismograms. Unfortunately, for many stations the crustal structure is not well known, which may hamper a detailed interpretation of the body wave part of the seismogram. For instance, it is difficult to isolate source effects or to interpret later arriving phases if crustal receiver effects are insufficiently known. In a foregoing paper (Paulssen, Visser & Nolet 1993), we have described a technique to determine the *P*- and *S*-velocity structure beneath a broad-band station from seismograms of teleseismic events. The method is based on a non-linear waveform inversion of the crustal receiver response which is obtained from the *P*-wave coda of teleseismic events. In this study we apply the method to data of the Network of Autonomously Recording Stations (NARS) in order to determine the receiver effects of these stations, but also to obtain insight into the variations of the crustal structure in western Europe.

NARS is a portable network of 14 broad-band (0.01–1 Hz), three-component stations (Dost, van Wettum & Nolet 1984; Nolet, Dost & Paulssen 1986). From 1983 to 1987 NARS was deployed as a linear array in western Europe with 14 stations installed at 18 different locations (Fig. 1). The stations were situated in different tectonic domains, varying from Precambrian (Scandinavia), to Caledonian (northwest Europe) and Variscan (central and southwestern Europe) regions, with some areas that have undergone Alpine deformation (Pyrenees and southern Spain). The crustal structure beneath most of the NARS stations is not known in detail, although a large number of deep seismic sounding experiments—notably those carried out within the context of the European Geotraverse Project

* Presently at: Faculty of Civil Engineering, Technical University Delft, P.O. Box 5048, 2600 GA Delft, The Netherlands.

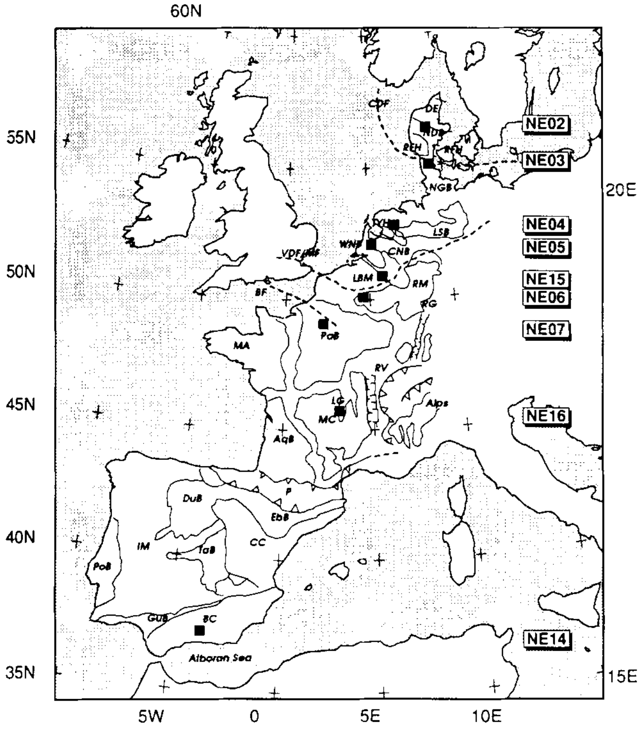

Figure 1. Geologic provinces and important tectonic features in western Europe. AqB = Aquitaine Basin, BC = Betic Cordillera, CC = Celtiberian Chain, CNB = Central Netherland Basin, DE = Danish Embayment, DuB = Duerro Massif, EbB = Ebro Basin, GuB = Gualdalquivir Basin, IM = Iberian Massif, LBM = London Brabant Massif, LG = Limagne Graben, LSB = Lower Saxony Basin, MA = Massif Amoricain, MC = Massif Central, NDB = Norwegian–Danish Basin, NGB = North German Basin, P = Pyrenees, PaB = Paris Basin, PoB = Portugese Basin, RFH = Ringköbing–Fyn High, RG = Rhine Graben, RM = Rhenish Massif, RV = Rhone Valley, TYH = Texel–Ysselmeer High, TaB = Tagus Basin, WNB = West Netherlands Basin. Fault zones (dashed): CDF = Caledonian Deformation Front, VDF/MF = Variscan Deformation Front, Midi Fault, BF = Bray Fault. Squares indicate the locations of NARS stations with the names indicated to the right of the figure.

(Freeman & Mueller 1990)—have contributed to a better insight into the crustal structure beneath western Europe.

In this paper, we present a brief outline of the method, and show data and results for eight of the stations of the NARS array. We discuss the models that are inferred from the data, and compare these with results of refraction and reflection investigations.

METHOD AND DATA SELECTION

For a detailed description of the method we refer to Paulssen *et al.* (1993); here we only present a brief outline of the procedure. The crustal response of a station is defined as the signal generated in the crust near the receiver due to a plane *P*-wave incident from the mantle. It contains the crustal reverberations and (multiple) conversions that arrive in the coda of the direct *P*-wave. The vertical and radial component of the crustal response are estimated from the *P*-wave coda of teleseismic events by deconvolution with their source wavelets. The source wavelet of an event is assumed to represent the *P*-wave signal at the base of the crust convolved with the instrument response, so it should incorporate source time function, effects from source and mantle, and the instrument response at the receiver. It is empirically picked from the data, and eliminated from the *P*-wave coda by the deconvolution technique of Langston (1979).

Individual crustal responses, obtained from different events in a certain source region, can be stacked to improve the signal-to-noise ratio. Stacking is allowed for clusters of events for which the crustal responses are expected to be similar. An estimate of the variance of a stacked crustal response can be obtained from the rms error of the data.

The synthetic seismograms used to model the data are calculated by the propagator matrix method. This implies that a plane layered crustal structure is assumed which is described by *P*-velocity α, *S*-velocity β, density ρ, and thickness z for each of the layers of the model. The model that gives the optimum fit between data and synthetics is determined by non-linear waveform inversion (Nolet 1987). The parameters of the upper mantle structure were kept fixed with $\alpha_{\mathrm{mantle}} = 8.1\ \mathrm{km\,s^{-1}}$, $\beta_{\mathrm{mantle}} = 4.3\ \mathrm{km\,s^{-1}}$, and $\rho_{\mathrm{mantle}} = 3.3\ \mathrm{g\,cm^{-3}}$. Because the density is ill-constrained by the data, it was coupled to the *P*-velocity by the ρ/α ratio of the starting model.

There are several selection criteria for the data that can be used for the determination of the crustal response. First of all, no other phases may arrive in the time window that is used for the determination of the crustal response. We considered a 15 s time window that includes the direct *P*-wave and the most prominent phases of the crustal structure at the receiver. We limited the data set to seismograms of deep events (>50 km) in order to avoid interference with the *pP* phase, and used only teleseismic events ($\Delta > 40°$). At distances larger than 72°, the *PcP* phase arrives in the time window of interest, but this phase has negligible amplitude (<5 per cent of the direct *P*-phase) at distances beyond 76° (surface source). Data from events at distances from 72° to 76° were therefore discarded, or equivalently, data with phase velocities of the direct *P*-wave ranging from 18.6 to 19.1 $\mathrm{km\,s^{-1}}$.

Another constraint is that the source time function must be short and simple. If the source wavelet is too long, then part of the crustal response may be included in it, causing crustal information to be eliminated in the deconvolution procedure. Moreover, a simple source time function helps to confidently identify a clear source wavelet in the seismogram. Using data of the NARS array, a visual comparison of the *P*-waveforms recorded by the different stations helps to distinguish between source and receiver effects and to identify the (common) source wavelet of the event.

An obvious selection criterion is that the signal-to-noise ratio must be sufficiently high to observe the relatively low-amplitude crustal phases. We required that the noise level on the seismogram prior to the *P*-phase was less than about 10 per cent of the *P*-amplitude on the vertical component.

The last criterion is that the effects of lateral heterogeneity should be small, because the data are modelled by a plane layered structure. This condition is obviously not met when the *P*-wave coda on the transverse component is large in amplitude. The signal on the transverse component was therefore checked, and in most cases appeared to be less than about 10 per cent of the amplitude of the direct *P*-wave. Data of two stations showed non-negligible energy on the transverse component, and it will be shown that their crustal (*P–SV*) responses are indicative of lateral heterogeneity by azimuth dependence.

INVERSION RESULTS

Crustal models have been obtained for 9 of the 18 locations of NARS stations. The other stations did not record sufficiently good quality events for a reliable determination of their crustal response. The results of station NE05 are described in a foregoing paper (Paulssen *et al.* 1993). In this paper, we describe the inversion results of the other eight NARS stations after a brief introduction on the geological setting of each of the locations. Event parameters of seismograms which were used to obtain the stacked crustal responses are given in Table 1. The locations of the stations and the most important geologic provinces and features which are mentioned in the text are shown in Fig. 1.

Station NE02 (Monsted; northern Denmark)

Station NE02 (56.459N, 9.170E) is located in the Norwegian–Danish basin. This basin was formed during the Permian when the post-orogenic uplift of the Variscides was paralleled by the subsidence of two large intra-cratonic basins: the Norwegian–Danish basin and the North German Basin (Ziegler 1990). An extensive overview of the geology and crustal structure of Denmark can be found in a paper by the EUGENO-S Working Group (1988).

The crustal velocity structure at NE02 is determined by the inversion of three (single, unstacked) crustal responses (see Table 1). The data and synthetics of the final model of the inversion are shown in Fig. 2a; the model is presented in Fig. 2b. A quick inspection of Fig. 2a reveals that the second response shows a pulse at about 13 s which is not modelled by the synthetics and which is also not present on

Table 1. Event parameters.

response	event code	Δ (°)	backaz. (°)	depth (km)	phase vel. (km/s)	average phase vel.
			Event parameters for station NE02			
1	860426	45.3	88.9	187	13.98	14.0
2	850423	88.7	63.9	188	23.60	23.6
3	830412	92.6	-94.9	107	24.17	24.2
			Event parameters for station NE03			
1	840423	71.8	28.8	415	18.58	18.8
	860608	75.5	30.8	61	19.01	
2	860511	82.0	43.3	223	21.79	22.3
	840424	84.1	42.1	424	22.76	
Table 1. (*continud*)						
			Event parameters for station NE04			
1	871003	47.3	83.2	80	14.18	14.2
	860426	47.0	83.5	187	14.27	
	870505	46.8	83.8	211	14.26	
2	870518	73.4	25.5	554	18.55	19.2
	870714	72.8	25.3	591	19.06	
	841203	77.8	27.3	70	19.63	
	840423	74.5	26.9	415	19.92	
3	840424	86.7	40.1	424	23.40	23.0
	860511	84.5	52.2	223	22.66	
			Event parameters for station NE15			
1	850802	47.9	81.6	103	14.32	14.4
	860426	47.8	81.1	187	14.43	
2	870714	74.8	24.5	591	19.53	20.1
	860416	80.0	27.0	52	20.70	
3	870422	84.1	34.2	49	23.14	23.5
	871212	90.2	38.6	183	23.87	
			Event parameters for station NE06			
1	850802	48.7	80.1	103	13.76	13.8
2	840420	75.7	22.9	581	19.74	20.4
	860719	78.8	22.5	151	20.38	
	841203	80.8	25.6	70	21.06	
	870114	80.8	29.9	99	21.09	
3	830412	89.1	-94.6	107	23.62	23.7
	850501	83.0	-106.7	600	23.80	
4	840424	89.6	38.4	424	23.98	24.0
	840306	91.1	38.7	457	24.15	
			Event parameters for station NE07			
1	870505	50.2	77.5	211	14.98	15.0
	831230	50.3	77.6	222	15.02	
2	870518	77.7	22.4	554	20.75	21.2
	870321	79.3	-0.2	97	20.47	
	860416	82.6	24.5	52	21.76	
	860608	82.8	25.5	61	21.87	
3	840306	92.9	36.9	457	24.38	24.5
	870425	95.3	58.3	124	24.54	
			Event parameters for station NE16			
1	870505	50.4	75.4	211	14.84	14.9
	850802	50.6	75.6	103	14.94	
	860426	50.6	75.1	187	15.03	
2	870518	80.5	22.7	554	21.83	22.4
	860416	85.3	25.0	52	22.72	
	870114	85.1	28.6	99	22.69	
			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	

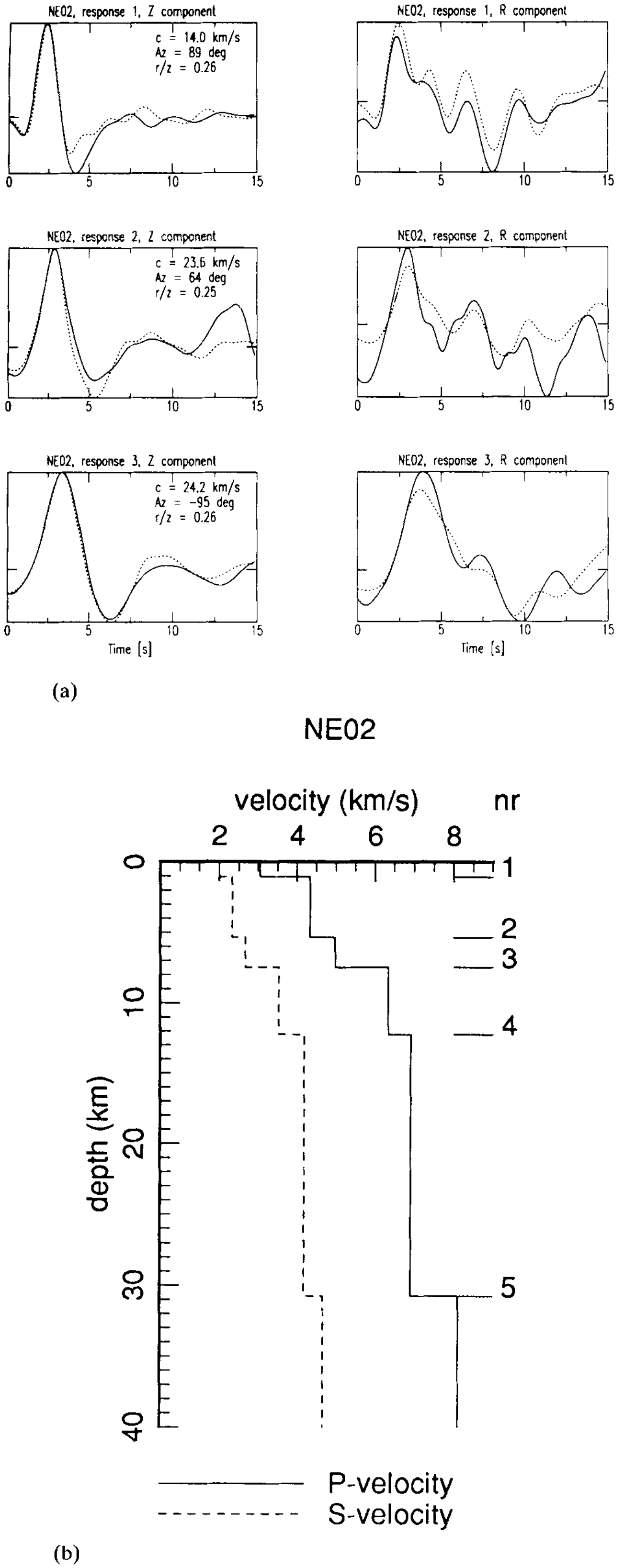

Figure 2. (a) Observed (solid lines) and synthetic (dashed lines) crustal responses for station NE02. Phase velocity, (back)azimuth and ratio of radial to vertical component are given in the upper right corner of each vertical component. Data are obtained from events specified in Table 1. (b) Final model of the inversion for which synthetics are shown in (a).

the other two observed responses. This signal can probably be considered as a noisy spike, generated at the source or in the mantle, and may be due to an incorrect estimate of the source wavelet. Unfortunately, since we only have one good quality seismogram from this source region, we cannot check this assumption or reduce its effect by stacking.

A three-layer starting model typical for a Caledonian crust (Meissner 1986) was used for the inversions, representing a layer of sediments, an upper crystalline basement, and a lower crust. Extra layers were added to model the response on the radial component in more detail. The signal on the radial component appears to be dominated by *P*-to-*S* conversions and (first order) multiples from discontinuities at mid-crustal depths and the Moho (interface 3, 4 and 5 of Fig. 2b). The finer detail in the velocity structure introduced by the second interface contributes to an improvement of the fit between data and synthetics immediately after the direct *P*-wave. The actual value of the low velocities of the first layer is not very well constrained: variations of about 10 per cent degrade the fit only marginally. The relatively good fit between data and synthetics for east (response 2 and 3) and west (response 1) backazimuths indicates that the hypothesis of lateral homogeneity of the crustal structure is in agreement with the data.

The model for station NE02 can be compared with the results of a dense network of refraction profiles carried out by the EUGENO-S Working Group (1988). This also facilitates an interpretation of the velocity structure in terms of the character of the different layers (Fig. 2b). Refraction experiments and well-logging data have revealed that the Norwegian–Danish basin is filled with an almost continuous succession from Mesozoic to Quaternary sediments, on top of a sequence of Paleozoic strata, overlying the crystalline crust of supposedly Caledonian origin. The uppermost, 1.1-km thick layer of our model is interpreted as a layer of unconsolidated Tertiary deposits. It overlies a 4.3-km-thick sequence of consolidated Mesozoic sediments with an average *P*-velocity of 4.3 km s^{-1}. This is in good agreement with the refraction data, which yield an average velocity 4.5 km s^{-1} for the approximately 5-km-thick Mesozoic sedimentary (second) layer. The refraction data indicate that the Lower Paleozoic sequence has a thickness of 2 to 3 km and a *P*-velocity of 5.2 to 5.8 km s^{-1}. Our modelling results give a 2.1-km-thick layer of slightly lower *P*-velocity (5.0 km s^{-1}). The actual crystalline basement of our model (layer 4) is only 4.8 km thick, which is in excellent agreement with the 4 to 5 km thick basement in the vicinity of the station inferred from the refraction data. The Conrad discontinuity of our model at 12.3 km depth is in good agreement with the refraction results. Note that our model does not show any lower crustal discontinuities. This is in accordance with the refraction data, which are indicative of a gradual increase in *P*-velocity of 6.6 to 6.9 km s^{-1}. Our inferred Moho depth at 30.9 km is also in fair agreement with the interpretation of the refraction profiles (28–30 km).

Unfortunately, we cannot assess the reliability of the *S*-velocity structure from independent evidence. The overall good agreement between our *P*-velocity model and the refraction results indicates that the method adequately determined the crustal (*P*-velocity) structure from the (unstacked) receiver responses.

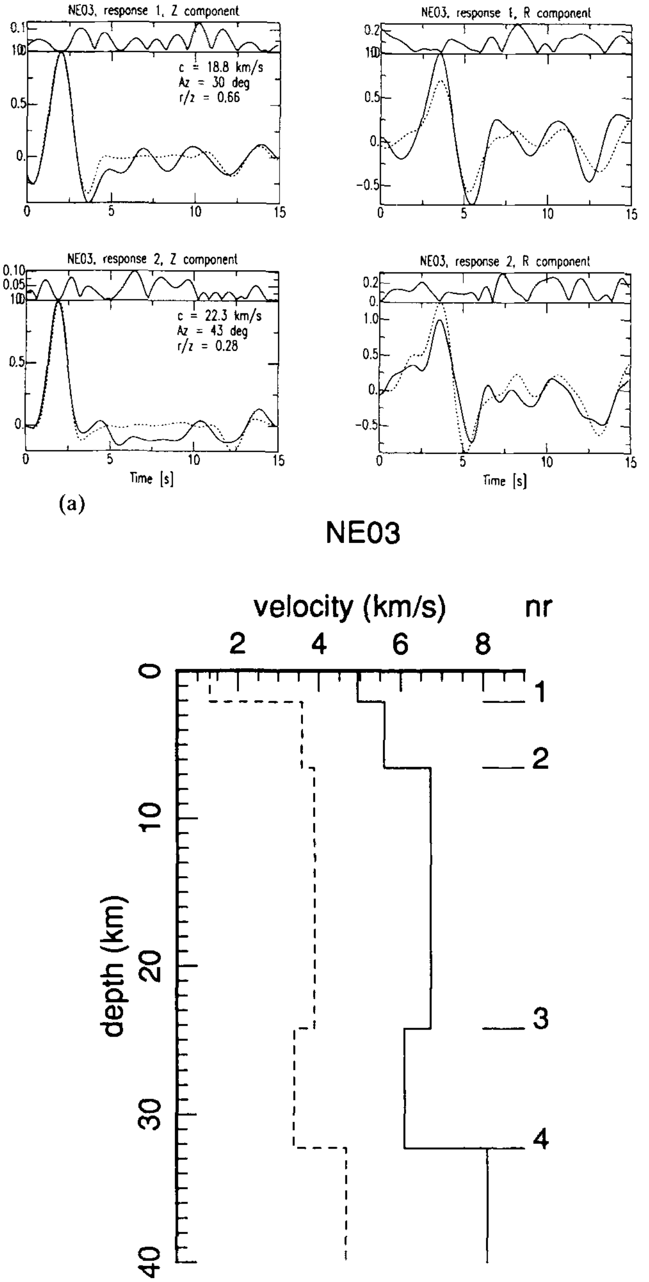

Figure 3. Same as Fig. 2 but for station NE03. The upper panels of figure (a) give time dependent variance of the data.

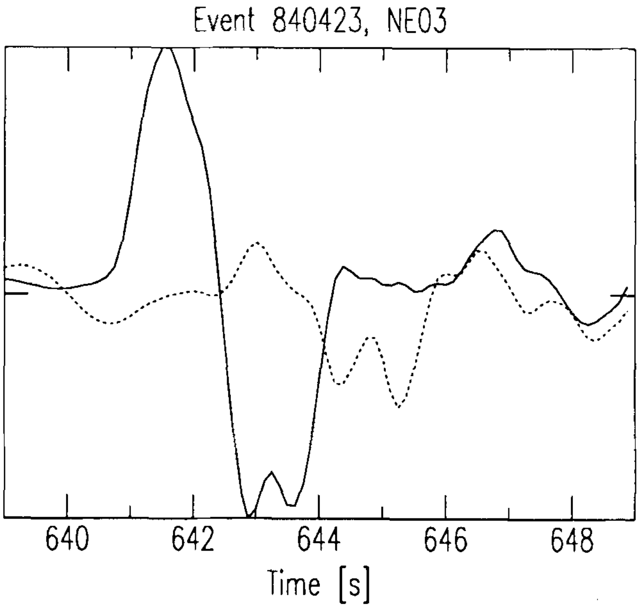

Figure 4. Initial part (vertical component solid and radial component dashed line) of a seismogram recorded by station NE03 for an event at the Kuriles (see Table 1 for event parameters).

Station NE03 (Logumkloster; southern Denmark)

Station NE03 (55.045N, 9.153E) is located close to the Caledonian Deformation Front (CDF). This front links the North Sea Caledonides with the Polish Caledonides through southernmost Denmark and defines the northern border of the North German Basin. The CDF has been considered to mark the transition from the Precambrian crystalline crust in the north to the Caledonian crust in the south, but the refraction profiles by the EUGENO-S Working Group (1988) indicate that this transition occurs at the (hypothetical) Trans European Fault, about 50 km to the south.

The crustal structure under station NE03 is determined from two crustal responses obtained from two different seismograms each. Data and synthetics of the optimum model are presented in Fig. 3a; the model is shown in Fig. 3b. The most striking feature of the responses is an 'apparent delay' on the radial component of 1.7 s relative to the signal on the vertical component. Fig. 4 illustrates this phenomenon on the raw data very clearly: the onset of direct *P*-wave cannot be observed on the radial component of the seismogram, and the signal on this component appears to be 'delayed' by 1.7 s. This observation points to two features: vertical incidence of the direct *P*-wave at the surface, and a high-amplitude *P*-to-*S* converted phase which arrives 1.7 s later. The inversion results indeed converged to a model that explains these phenomena. The low *P*- and *S*-velocities in the uppermost layer of the model cause near-vertical incidence at the surface. The 'delayed' arrival on the radial component is a *P*-to-*S* conversion from the second interface (the delay of which is caused mainly by the low *S*-velocity in the uppermost layer). The strong negative amplitude at about 6 s on the radial component of the responses is produced by the first *SV*-multiple reflected at the interface at 2.1 km depth. The *S*-velocity in the first layer is therefore very well constrained. The first *P*-wave multiple from this interface is not large in amplitude, indicating that the *P*-velocity constrast must be smaller than the *S*-velocity contrast.

None of the parameters of the model are as well constrained as the *S*-velocity of the uppermost layer. For some of the model parameters, variations of about 5 per cent give a nearly equally good fit to the data. Features of the model that are relatively well resolved are the *S*-velocity of the third layer and Moho depth.

It is not simple to make a detailed comparison between the optimum model for NE03 and the EUGENO-S refraction data (1988) because the crustal structure changes rapidly in the vicinity of the station. Yet, the velocity structure shows a remarkably good correlation with the interpretation of the refraction data. Station NE03 is located in the southward extension of the Brande Trough, which separates two blocks of the Ringköbing–Fyn high, the eastern one of which (the Glamsbjerg block) is sampled by our data. The Ringköbing–Fyn high is a structural high which has a shield-like crustal structure. The refraction experiments have shown that the Brande Trough is a circa 30-km-wide graben filled with a 2-km cover of Mesozoic–Cenozoic deposits overlying a 4-km thick infill of Paleozoic sediments.

The two uppermost layers of our model exhibit exactly the same velocity structure as the refraction results for the Brande Trough: a 2.1-km-thick layer of unconsolidated sediments on top of a 4.5-km layer with a *P*-velocity of 5.6 km s^{-1} (the refraction data give a velocity of 5.2 to 5.8 km s^{-1} for this layer of Paleozoic sediments). According to the refraction data, a sharp gradient then marks the transition to the on average circa 20-km-thick Precambrian crystalline crust. Beneath the Glamsbjerg block, the upper part of this layer first has a rather uniform velocity (6.4 km s^{-1}); the *P*-velocity gradually increases to 6.9 km s^{-1} in the lower part. Our model yields an average *P*-velocity of 6.7 km s^{-1} over this 17.6-km-thick layer. The low-velocity zone of the fourth layer is also present in the refraction results. The 10-km-thick low-velocity zone of the profile dies out or dips beneath the Moho from the Glamsbjerg block towards the Brande Trough. The interpretation of this feature is presently not clear (see the tectonic interpretation of Profile 5 of the EUGENO-S experiment, 1988).

From the above discussion, it seems that our crustal model displays the most prominent features of the crustal structure under NE03, although the actual velocities may be slightly in error.

Station NE04 (Witteveen; northern Netherlands)

Station NE04 (52.813N, 6.668E) is situated in the North German basin, just outside the Lower Saxony basin. Although the crust is of presumably Caledonian origin, it is highly affected by periods of subsidence in the Mesozoic, which were caused by wrench movements between the Danish–German block in the north and the Variscan Massifs in the south (Ziegler 1990).

The crustal structure under station NE04 is derived from three sets of stacked crustal responses. The observed and synthetic responses are depicted in Fig. 5a, and the final model of the inversion is shown in Fig. 5b.

Similar to the data of NE03, the radial component of the crustal response of NE04 is characterized by a 'delay' of about 1.6 s. In this case, the delayed arrival on the radial component is clearly broader than the *P*-phase on the vertical. This is best explained by the interference of several *P*-to-*S* conversions, i.e. from the first three interfaces of the final model of the inversion.

The crustal responses of station NE04 are characterized by high coda amplitudes on the radial and vertical component. This is attributed to the high reflectivity of the fifth interface, which keeps nearly all energy trapped within the first five layers of the crust. As a consequence, the inversion was mainly sensitive to the velocity structure of the upper layers. Although the two lowermost layers in the

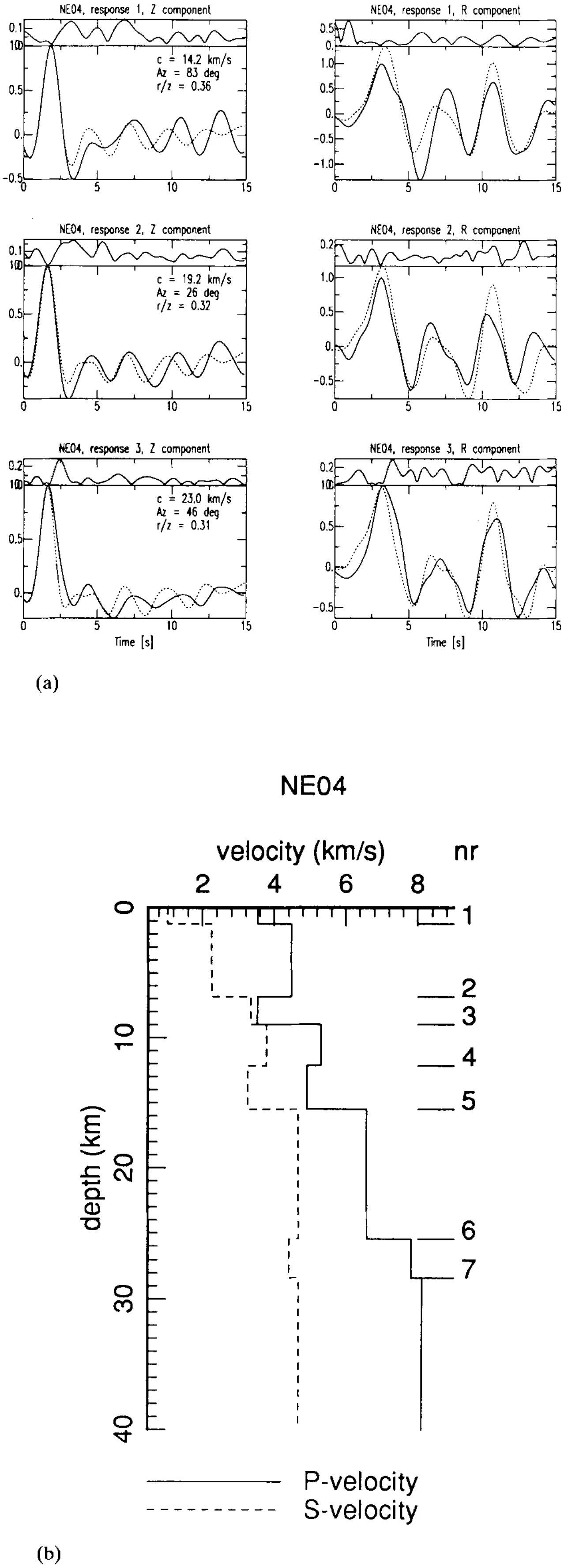

Figure 5. Same as Figs 2 and 3 but for station NE04.

model did contribute to a better fit between the data and the synthetic seismograms, many different lower crustal velocity structures were found that gave a nearly equally good fit. Layers 6 and 7 of our model are therefore very poorly resolved. We also note that the vertical component of the responses is poorly modelled, suggesting that the *P*-velocity structure may not be very reliable. Furthermore, the combination of *P*- and *S*-velocity of the third layer of the model is obviously not realistic, as this would imply a negative Poisson's ratio. It indicates that the unconstrained inversion converged to a physically unacceptable model due to inconsistencies or noise in the data. Instead of coupling or constraining the model parameters to realistic values, we chose to let the model itself be indicative of errors in the data, because it is unclear whether the main source of error occurs on vertical component (thereby mainly affecting the inferred *P*-velocity structure) or on the radial component (affecting the *S*-velocity structure). Increasing the *P*-velocity of the third layer to a high velocity, or decreasing the *S*-velocity to a low velocity, in both cases results in a significant mismatch between data and synthetics.

There are no wide-angle deep seismic refraction experiments carried out in The Netherlands, but we can compare our results with deep reflection data (Remmelts & Duin 1990). In the vicinity of station NE04, Remmelts & Duin model a circa 1.6-km-thick layer of Cenozoic deposits on top of 2.4-km Mesozoic and 3.3-km Permian sediments. They tentatively infer the top of basement at a depth of about 7.1 km. Furthermore, from *P*-delay times, Souriau (1979) estimated the sedimentary thickness at station WIT (same location as NE04) to be 6 to 8 km. The first two, or possibly three layers, of the model of Fig. 5b can therefore be considered as being of sedimentary nature. The low *P*- and *S*-velocities, and the high Poisson's ratio ($\sigma = 0.45$) of the uppermost layer are in agreement with an interpretation of unconsolidated (probably Cenozoic) sediments. The lower crustal structure is badly constrained; the good agreement of the Moho depth of 28.7 km with the reflection data (29 km, Remmelts & Duin 1990) may be coincidental.

Station NE05 (Utrecht; central Netherlands)

For a description of the inversion of the crustal responses of station NE05, we refer to Paulssen *et al.* (1993).

Station NE15 (Valkenburg; southern Netherlands)

Station NE15 (50.86 N, 3.103E) is situated in eastern London–Brabant Massif. It is enclosed by two branches of a major fault system just north of the Midi fault. This fault is considered to be part of the Variscan Deformation front which separates the Caledonian basement in the north from the Variscan basement in the south. The area around NE15 is enclosed by faults and forms a structural high in the pre-Permian subsurface (Bless & Bouckaert 1988).

The crustal model for NE15, depicted in Fig. 6b, is determined by inversion of three crustal responses that are obtained from two seismograms each. The synthetic and observed seismograms are displayed in Fig. 6a. The data of this station do not show an apparent delay as found for stations NE03, NE04 and NE05, and there is no evidence for high-amplitude SV-multiples. The coherence of the

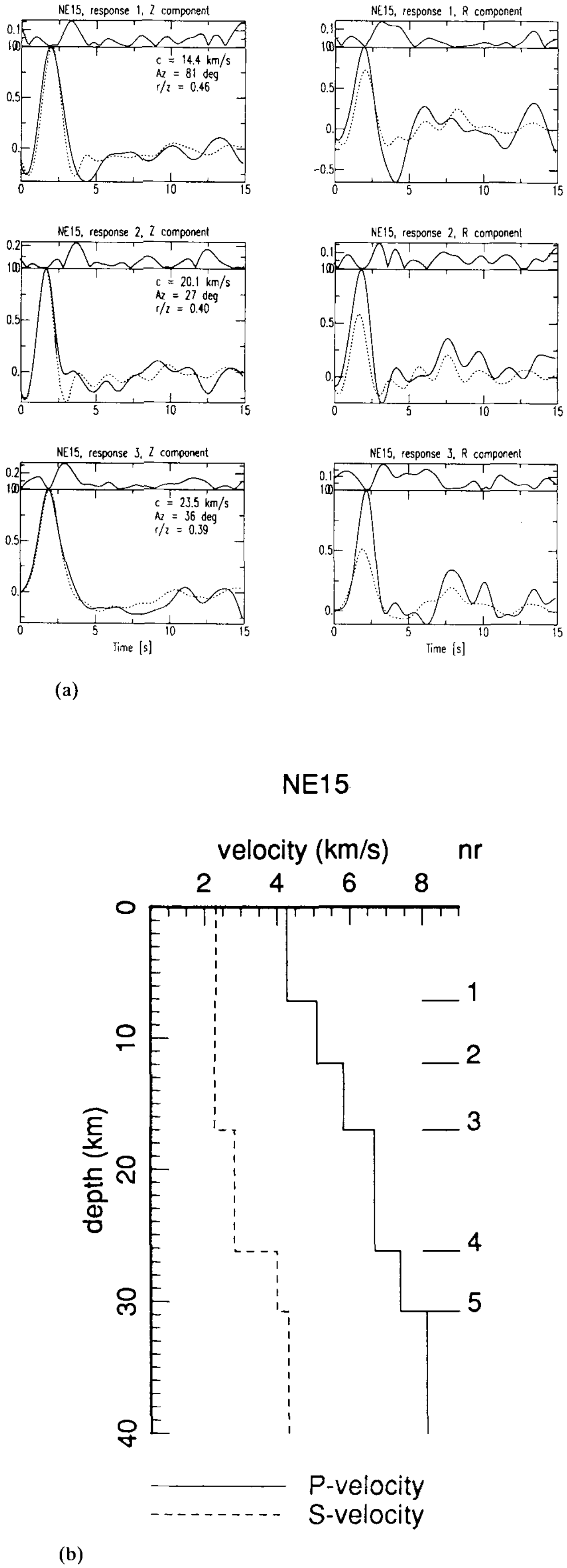

Figure 6. Same as Figs 2 and 3 but for station NE15.

responses is rather poor, although there is no evidence for lateral heterogeneity from the data on transverse components of the seismograms. The only outstanding feature is the arrival at about 7.5 s on the radial component of the second and third response, but it is not present on the first response. We consider the model inferred from the data therefore not very reliable. The poor resolution is also apparent from the inversions as no strong minimum in the misfit could be found, as well as from the large misfit between data and synthetics. Moreover, the amplitude of the direct *P*-wave on the radial component appeared difficult to match. The model presented in Fig. 6b should consequently not be considered as a reliable representation of the crustal velocity structure at station NE15, and a further interpretation of the inversion result is therefore omitted. We only note that the absence of clear phases on the crustal responses indicates that there are no prominent velocity discontinuities in the crustal structure under NE15. This is consistent with reflection results (DEKORP Research Group 1990a,b) that show a very 'transparent' crustal structure and a rather weak band of Moho reflections in the area of the London–Brabant Massif close to the station.

Station NE06 (Dourbes; Ardennes, Belgium)

Station NE06 (50.097N, 4.595E) is situated in the Ardennes forming the western extension of the Rhenish Massif. The Rhenish Massif is bordered to the north by the Midi (-Aachen) Fault that constitutes part of the Variscan Deformation front. This is a major low-angle thrust fault system that was activated during the Carboniferous folding of the Rhenohercynian zone of the Varisan Orogen. The Rhenohercynian fold-belt forms the suture between the North European Craton and the Central Amorican–Saxothuringian terranes, which were separated by a back-arc ocean in Devonian times (Ziegler 1989).

The crustal structure under station NE06 was determined from four crustal responses. The data are shown in Fig. 7a, and the final model of the inversion is given in Fig. 7b. The responses of this station do not show an 'apparent delay', but the first dominant arrival appears to be broader on the vertical and radial component when compared to other stations of the NARS array. This broadening must be caused by the uppermost structure, or, more specifically, by the interference of the direct *P*-wave with a *P*-wave multiple reflected at the first interface. The multiple has the same polarity as the direct *P*-wave and must therefore be reflected at an interface that defines a velocity decrease with depth. The high amplitude of the *P*-wave multiple and the absence of a clear *SV* multiple or *P*-to-*S* converted phase are indicative of a large *P*-velocity contrast which is accompanied by a small *S*-velocity gradient.

The best resolved parameters of the model are the *S*-velocity in the first and third layer and the depth of the third interface (10 per cent variations in the model parameters give an approximately 20 per cent increase of the misfit). When the *S*-velocity of the third layer is changed to a high value (no low-velocity layer), the misfit increases significantly. The *P*-velocity and depth of the second layer are reasonably well resolved, while the other parameters are less constrained.

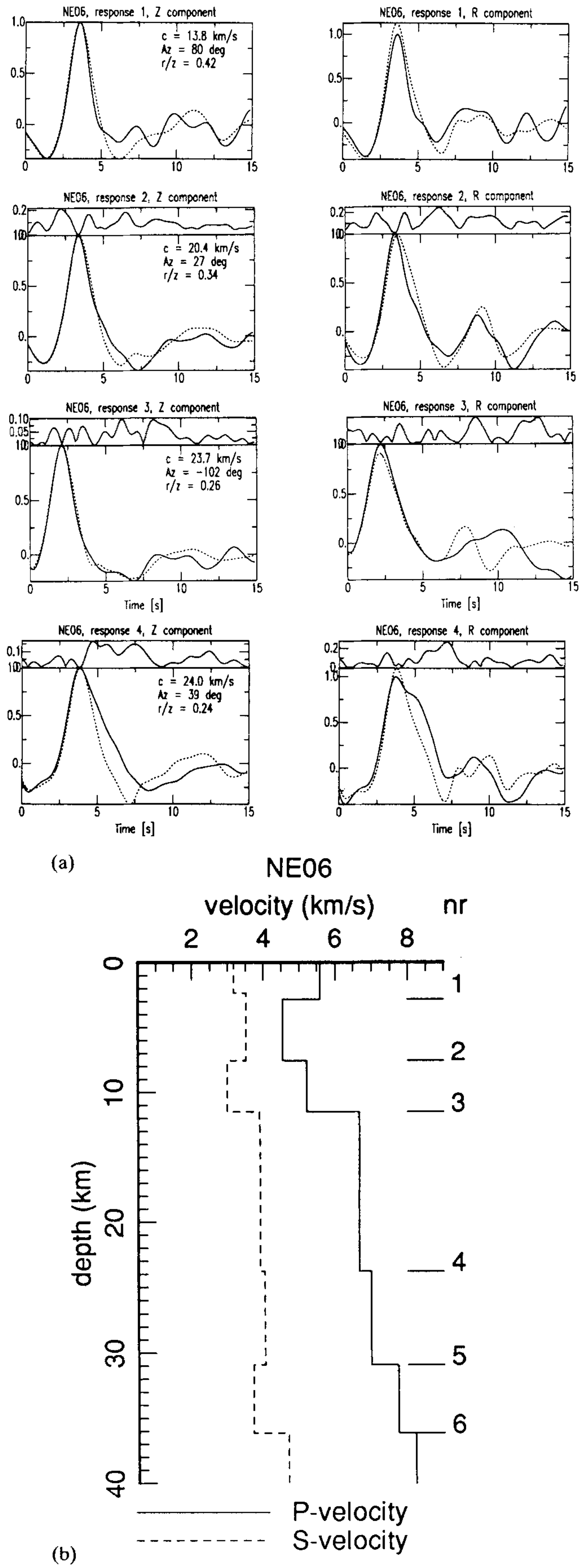

Figure 7. Same as Figs 2 and 3 but for station NE06.

Comparison of the model for NE06 with the results from reflection and refraction surveys (Cazes *et al.* 1986; Hirn *et al.* 1987; Matte & Hirn 1988) suggests that the upper two layers probably represent the strongly folded, autochtonous Dinant–Ardennes nappe; a *P*-velocity of 5.5 km s^{-1} in this depth range is inferred from the refraction data. The reflection data indicate that this unit is separated from the underlying autochtonous unit by a south-dipping decollement surface which emerges at the Midi Fault 50 km north of NE06. This decollement surface is located at a depth of 5–7 km 100 km east of NE06, and probably corresponds to interface 2 of our model. The third interface could correspond to the interface between the autochtonous Paleozoic sediments (note the low velocities in layer 3) and the reflection-seismically relatively transparent Caledonian basement of the Brabant block. The refraction data give a *P*-velocity of 6.7 km s^{-1} in the lower crust down to the Moho. Although the Moho is not visible on the vertical reflection data, deep refraction data indicate that it occurs at a depth of roughly 35 km, which is in good agreement with our model.

Station NE07 (Villiers-Adam; Paris-Basin, France)

Station NE07 (49.074N, 2.232E) is situated in the Mesozoic Paris Basin, which is underlain by a Variscan basement complex consisting of Precambrian crustal elements, Paleozoic metamorphics and intrusives. The basement, which consists of north-verging nappes, is transsected by Permo–Carboniferous strike-slip fault systems (Cazes *et al.* 1986; Bois *et al.* 1988; Matte & Hirn 1988). Some of these NW–SE trending faults became reactivated during the Mesozoic and Cenozoic evolution of the Paris Basin. One of these is the Bray Fault, located immediately to the north of station NE07. This fault is associated with a large surface anticline formed during the Tertiary.

The crustal structure under station NE07 is inferred from three crustal responses. The data and the optimum model obtained by the inversion are shown in Fig. 8. The crustal responses of NE07 are, like those of NE03, NE04 and NE05, characterized by an 'apparent delay' of the first arrival on the radial component. The onset of the signal on the radial component seems to be delayed by about 0.6 s with respect to the vertical component. This observation is indicative of (near) vertical incidence of the direct *P*-wave at the surface, and of a high amplitude *P*-to-*S* converted phase from the first interface that arrives 0.6 s later. The large velocity contrast across the first interface generates strong *SV*-multiples that gradually decrease in amplitude. The *S*-velocity structure below the first interface is poorly resolved, because its response is obscured by the high amplitude multiples on the radial component. There is, however, little evidence for large *S*-velocity discontinuities at greater depths, except for the Moho transition at 36 km depth.

The uncertainty in the *P*-velocity structure is largely due to the low amplitudes on the vertical component: the amplitude level on this component is generally smaller than the variance in the data. The low amplitude signal indicates that there are no large gradients in the *P*-velocity structure at NE07, as is evident from the final model of the inversion. The most prominent features of the model are the

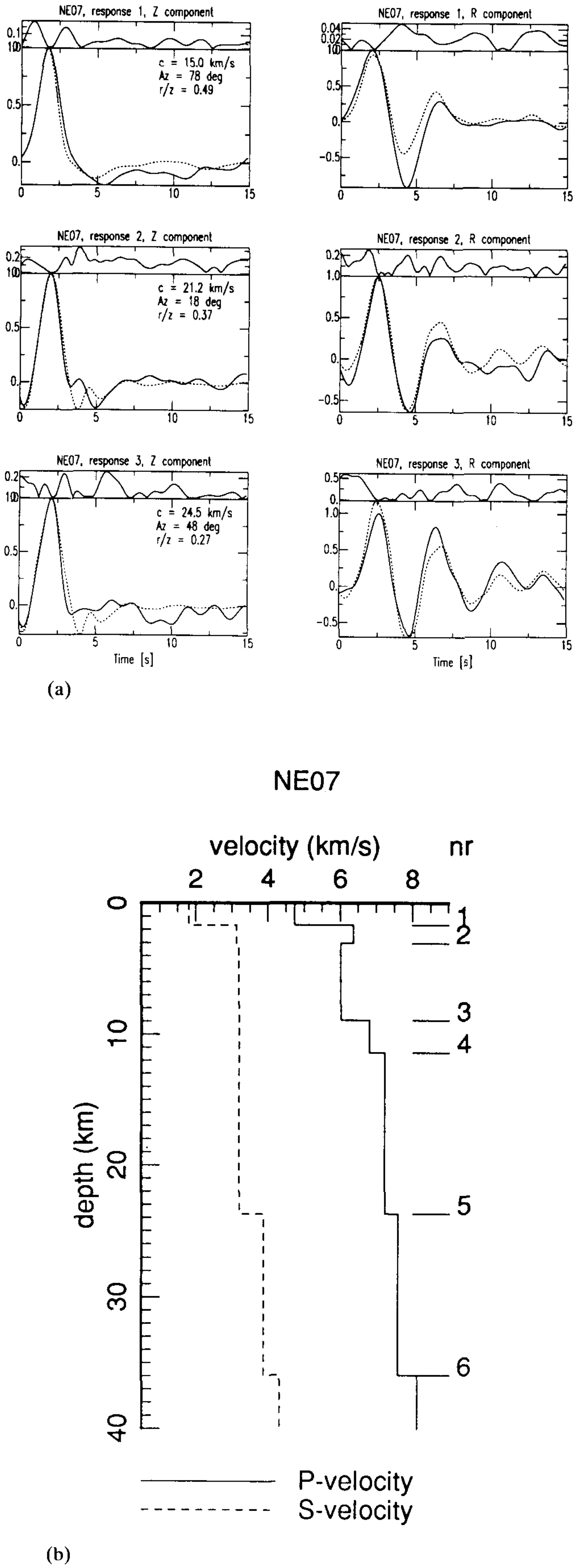

Figure 8. Same as Figs 2 and 3 but for station NE07.

transitions at 1.7, 9 and 36 km depth. The finer detail introduced by the other layers and interfaces only slightly improves the match between data and synthetics, and is probably not really warranted by the data.

Reflection and refraction data show evidence for a complex velocity structure of the Paris Basin and its underlying crust (Matte & Hirn 1988). The Mesozoic sediments are approximately 2 km thick. The upper crust (circa 2–10 km depth) is interpreted as a complex assemblage of Precambrian rocks, and is associated with the allochtonous Central Amorican block. The transition from the upper to the lower crust at a depth of approximately 10 km coincides with a subhorizontal thrust plane, which carries the Central Amorican block onto lower Paleozoic sediments. The lower crust appears as a laminated zone on the reflection profile with the Moho at a depth of about 35 km. Although it is difficult to compare our model with the refraction and reflection results because of the large uncertainty in the velocity structure, it is possible to interpret the most distinct features of the model. The first discontinuity marks the transition from the Mesozoic cover to the (allochtonous) upper crust, and the third interface is probably associated with the transition from the upper to the lower crust. The Moho depth (36 km) shows a surprisingly good agreement with the reflection data, considering that the velocities of our model strongly deviate from the refraction results (Cazes *et al.* 1986).

Station NE16 (Clermont-Ferrand; Massif Central, France)

Station NE16 (45.763N, 3.103E) is situated in the Limagne Graben at the border with the Massif Central. The Limagne Graben is a Tertiary rift in the Massif Central, the crust of which was consolidated during the Variscan orogenic cycle.

The *P*-wave coda of station NE16 show a high noise level and a strong azimuthal dependence. Although this is indicative of lateral heterogeneity, we have applied our method to data of this station to investigate consistent elements of the crustal structure. We have only considered responses that were obtained by stacking data of (at least) three events. The inversion was initially performed for the two available responses (Fig. 9a) at the same time. The model that was obtained by this inversion matched the most outstanding features of the data, but still gave a relatively large misfit. This 'average' model shows *P*- and *S*-velocity increases at depths of 0.8, 8.2 12.5 and 27.3 km. A better fit to the data was achieved by modelling the two responses separately, and the synthetics and final models of these inversion runs are shown in Fig. 9. The velocities and depths of these two models differ generally by less than 3 per cent from the 'average' velocity structure, and the features of the models can therefore be considered as representative of the general velocity structure at NE16.

The signal on the radial component of the first response is dominated by the effects of the topmost layer. The *P*-to-*S* conversion from the first interface causes an 'apparent delay' on the radial component of 0.5 s, and the *SV*-multiples from this discontinuity are clearly visible as the wavetrain decreasing in amplitude on the radial component. The effects of the structure below the first (sedimentary) layer are more pronounced on the second response. The velocity increases for these data are located at approximately the

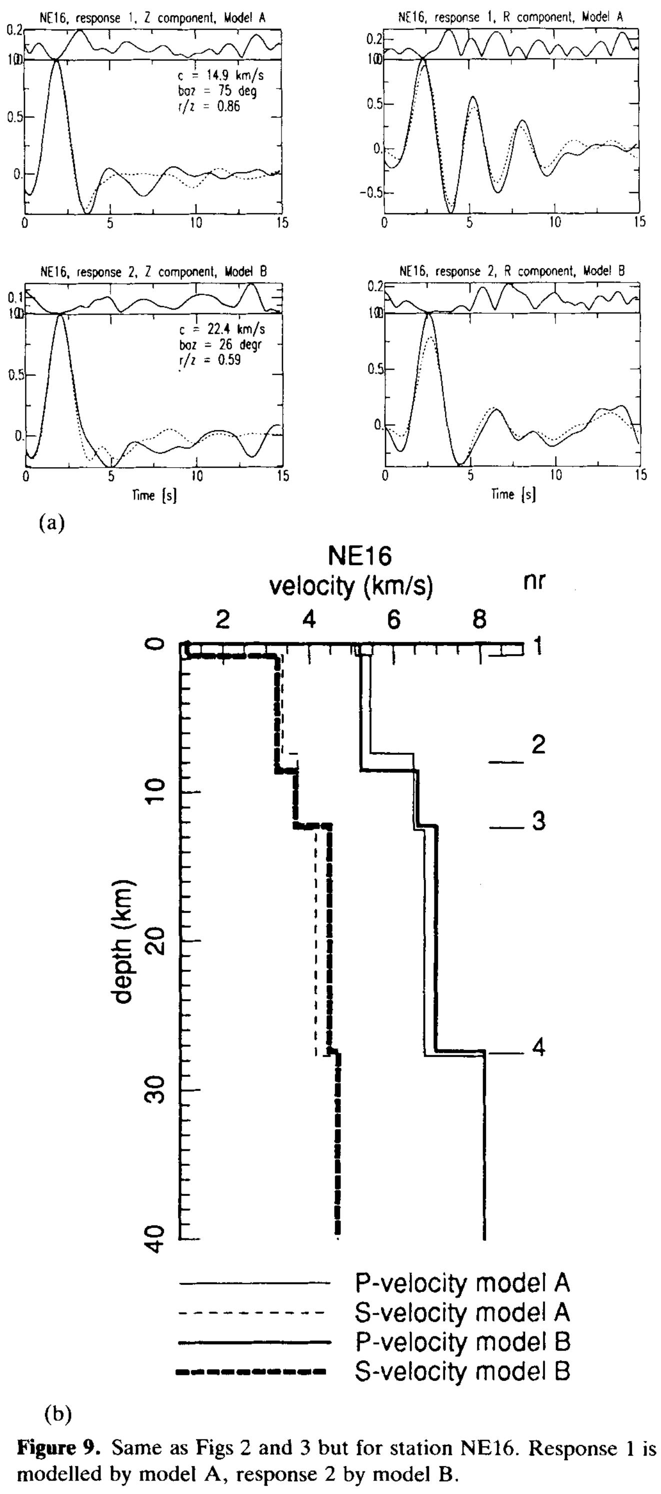

Figure 9. Same as Figs 2 and 3 but for station NE16. Response 1 is modelled by model A, response 2 by model B.

same depths as for the first response. The *S*-velocity structure is in both cases better resolved than the *P*-velocity structure. This is also evident from the modelling results: the match between data and synthetics is good for the radial (*SV*-) component, but is very poor for the vertical (*P*-) component of the responses.

As mentioned before, station NE16 is located at the border of the Limagne Graben in the Massif Central. Refraction experiments have been carried out (Perrier &

Ruegg 1973) which indicate that the Moho becomes shallower beneath the graben (24 km) and descends in the eastern part of the Massif Central (27 km). Our models, which indicate a Moho depth of 27 km are, however, so far not in agreement with the refraction results for the eastern Massif Central, as they show no evidence for the presence of an upper crustal low velocity zone as suggested by the refraction data. Moreover, they strongly deviate from the refraction results for the Limagne Graben. We note that our *P*-velocity structure is very poorly resolved, which might explain the discrepancy with the refraction data. Yet, the general features of the *S*-velocity appear to be reliable, as the inversions always converge to models with *S*-velocity increases at about 0.8, 8, 12 and 27 km depth.

Station NE14 (Granada; southern Spain)

Station NE14 (37.190N, 3.959W) is located in the Betic Cordillera at the transition from the Internal Zone in the south to the External Zone in the north. The Betic Cordillera represents, together with the Rif fold-belt in northern Africa and the Alboran Sea, the westernmost part of the Alpine chain of Europe. The tectonic evolution of the western Mediterranean area is still a subject of debate (see e.g. Vegas & Banda 1982; Dewey *et al.* 1989; Platt & Vissers 1989) and information about the (deep) crustal structure in the Betics may help to increase our knowledge of the geodynamics of the region.

The crustal structure under NE14 was investigated using four crustal responses. The transverse component of the seismograms showed a higher amplitude level than for the other stations (roughly 30 per cent of direct *P*-wave), this would point to (structural) lateral heterogeneity or to scattering effects in the vicinity of the station. Two of the responses are obtained for west to southwest backazimuths, one for a north, and one for an east backazimuth. It was impossible to obtain a reasonable fit to all four responses by a single model. The inversions appeared to be unstable: they converged to velocity structures that modelled just one or two of the data sets, and the final model of the inversion depended on the choice of the starting model. The data for the three different azimuth ranges were therefore modelled separately. The data might be inverted for dipping interfaces whereby heterogeneity is taken into account, but this has not been attempted at this point.

The velocity structure for west to southwest azimuths is fairly well constrained by the inversion of two responses. The data (responses 1 and 2 in Fig. 10a) show an 'apparent delay' of about 0.8 s, and the first dominant phase on the radial component can be identified as the *P*-to-*S* conversion from the first discontinuity at 1.3 km depth (see model A in Fig. 10b). The high amplitudes on the radial component are mainly caused by *SV*-multiples from this interface. The responses are dominated by the upper crustal structure, and the model below the third interface is not as well resolved as its upper part.

The responses for east and north backazimuths (responses 3 and 4 in Fig. 10a, respectively), are dissimilar from the western/southwestern responses. An important difference is the width of the first dominant arrival on the radial component which is broader than on the other two responses. The variations of this initial signal are mainly due to variations in the topmost layer of the model. The north and east responses are best modelled by a thicker sedimentary layer (circa 2.3 km). The velocity structure for the third and fourth response (model B and C of Fig. 10b) differ from each other mainly in the responses on the radial component between 5 and 10 s. The structure below approximately 27 km is not very reliable on any of the three models.

Refraction experiments in the central part of the Betic Cordillera (Banda & Ansorge 1980) show evidence for an upper crustal low *P*-velocity zone at depths between 7 and 12 km which disappears towards the east and north. Our model shows a low *S*-velocity zone in this depth range which seems to be more pronounced in the (south)west and east than to the north. The *P*-velocity structure of our model is less well constrained, but does not suggest the presence of a low *P*-velocity zone in the critical depth interval. All four responses indicate a large *P*-velocity increase at about 27 km. This is consistent with the refraction data which show, however, no evidence for the occurrence of a low velocity zone above this level. Whereas our models indicate a Moho depth of 27 km, the refraction data suggest that the Moho is located at a depth of 39 km under the central part of the Betics (Banda & Ansorge 1980). The velocity structure below the fifth interface is only very poorly resolved and therefore not reliable.

DISCUSSION AND CONCLUSION

The method to determine the crustal structure beneath broad-band stations is developed to be able to calculate the most important effects of the crustal structure beneath the receiver. As applied to the NARS stations, it appears that in most cases the (non-linear) inversion rapidly converges to a model that gives an adequate fit to the most pronounced features of the crustal responses. The approach is therefore successful, but it implies that the accuracy of the models heavily depends on the accuracy of the data that are used as input to the inversion. To achieve an optimum result, responses of clusters of events should be stacked to improve the signal-to-noise ratio of the data. In this way, spurious artifacts are suppressed that would otherwise greatly influence the final model of the inversion. A second, not less important, aspect is that the model parameters become better constrained if several responses are available for different phase velocity and azimuth. This helps to assess the consistency of the data for an interpretation in terms of a plane layered crustal structure and to better constrain the velocity model.

It was found that many of the NARS stations showed an apparent delay of the first arrival on the radial component. For these stations it is observed that there is only very little signal on the radial component (of the raw data) prior to the arrival of the *P*-to-*S* converted phase from the first interface. The low seismic velocities for the uppermost layer that are needed to explain the data are in agreement with values obtained for unconsolidated sediments. Poisson's ratio of the uppermost layer of these models (NE03, NE04, NE05, NE07, NE16 and NE14) varies between 0.41 and 0.47: values that are indeed typical for unconsolidated sediments.

Another striking feature is the relatively high amplitude

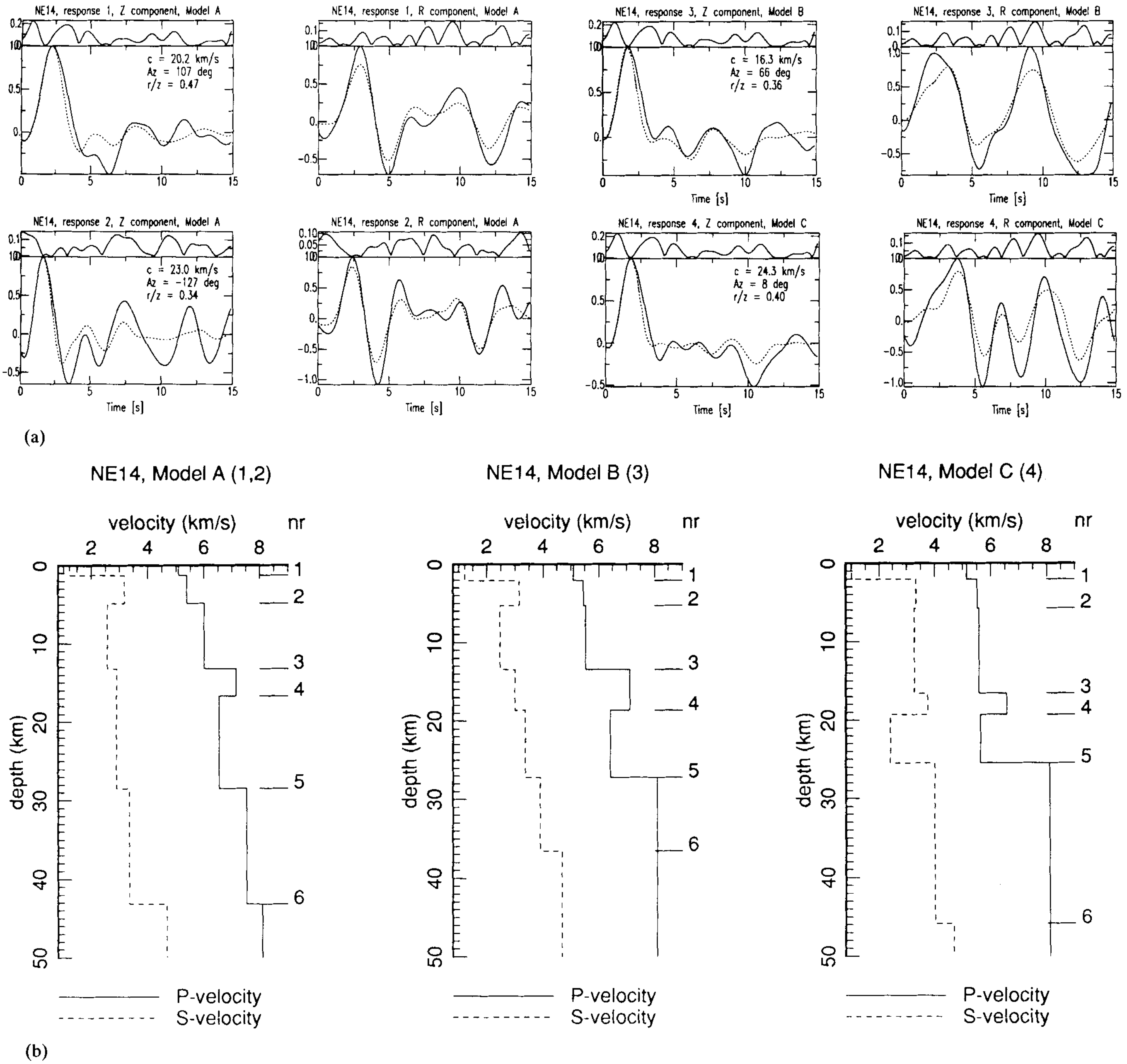

Figure 10. Same as Figs 2 and 3 but for station NE14. Response 1 and 2 are modelled by model A, response 3 by model B, and response 4 by model C.

level on the radial component for many of the stations. This points to the presence of strong S-wave reflectors in the upper crust. The response of the upper crust obscures the signal generated in the lower crust, which hampers accurate modelling of the lower crustal structure. It is evident that the inversion is mainly sensitive to large velocity contrasts which produce high-amplitude reflected and converted phases.

Considering the generally quite reasonable fit between our modelling results of teleseismic data and refraction data, the method described here is adequate for calculating the most important local crustal effects. The method can therefore also be viewed as an alternative to other types of crustal exploration techniques, such as deep seismic reflection or refraction or surface wave studies. It is an inexpensive technique to investigate the crustal P- and S-velocity structure, since it makes use of available teleseismic data. Comparison with refraction and reflection data indicates that the actual velocities may not always be accurately resolved, but the presence of strong velocity gradients and their approximate depths are generally well-determined.

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