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Crust and upper mantle structure of the central Iberian Meseta (Spain)

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Summary. Quarry blasts recorded along three lines on the central Iberian Meseta are used in an attempt to interpret the crustal structure. The results of the interpretation of the data, together with published surface wave and earthquake data, suggest a layered structure of the crust having the following features: the basement, in some areas covered by up to 4 km of sediments, has a P-velocity of 6.1 km s^{-1}; a low-velocity layer, between 7 and 11 km depth, seems to exist on the basis of both P and S interpretation of seismic data; a thick middle crust of 12 km has a P-velocity of 6.4 km s^{-1} and overlies a lower crust with a mean P-velocity of 6.9 km s^{-1} and a possible slight negative gradient; the mean v_p/v_s ratio for the crust is about 1.75; the Moho is reached at about 31 km depth and consists of a transition zone at least 1.5 km thick. The P-velocity of the upper mantle is close to 8.1 km s^{-1} and the S-velocity about 4.5 km s^{-1}, which gives a v_p/v_s ratio of 1.8 for the uppermost mantle. A tentative petrological interpretation of the velocities and composition of the layers is given.

Introduction

The crustal structure of the Alpine provinces of the Iberian Peninsula, the Betic Cordillera and the Pyrenees, has been extensively studied by means of seismic refraction experiments during the last few years (Working Group for Deep Seismic Soundings in Spain 1974–1975 1977; Banda & Ansorge 1980; Explosion Seismology Group Pyrenees 1980; Gallart *et al.*

1980). Less attention has so far been paid to the stable part of the Peninsula, at least in Spain. In Portugal early work by Mueller *et al.* (1973) has been extended to the north by Mendes Victor, Hirn & Veinante (1981).

Since 1976, the Instituto Geográfico Nacional of Spain has been recording quarry blasts fired by the Fábrica de Cementos Asland at Yepes, near Toledo, together with other Spanish institutions (Payo & Ruiz de la Parte 1977), so that now some data on the central Spanish Meseta are available.

In addition, an interpretation of the phase and group velocities of surface waves from paths crossing the Iberian Peninsula and the travel times from near earthquakes has been carried out and published (Payo 1965, 1970, 1972; Payo & Ruiz de la Parte 1974; Perez, Payo & Ruiz de la Parte 1978; Sierra 1980). It is the aim of this paper to present the new data obtained from explosion seismology and to compare the results with those from earthquake seismology.

The Iberian Massif (Fig. 1), is part of the Hercynian belt of Western Europe which outcrops in larger areas of the Iberian Peninsula. Apart from some Tertiary basins, the uplift of the Iberian Massif during the Mesozoic has caused the sedimentary cover to be thicker and more complete on its margins. To the north and west, the Iberian Massif is overthrusted by its Mesozoic cover (Cantabrian shelf and Mesozoic cover of Lisboa). To the south it is in contact with the Betic Alpine chain, whereas to the east the Hercynian basement is covered by Tertiary and Mesozoic sediments affected by Alpine deformation (the Iberian Chain) (Fig. 1).

The seismic profiles reported here (Fig. 1) were carried out in different stages; recording explosions, whose size varied from 1500 to 5000 kg, with ten 3-component portable stations. Measurements at some of the sites had to be repeated due to noisy conditions or

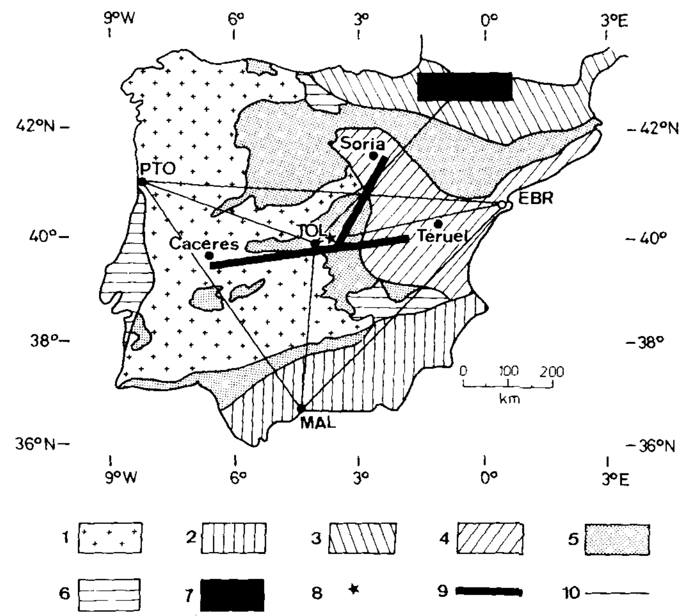

Figure 1. Simplified geotectonic map of the Iberian Peninsula, where the seismic profiles and the paths studied by surface waves are indicated. (1) Iberian (Hercynian) Massif. (2) Betic Alpine Chain. (3) Pyrenean range. (4) Iberian Chain. (5) Tertiary basins. (6) Mesozoic basins. (7) Epicentral zone for group-velocity studies. (8) Shot point. (9) Seismic refraction line. (10) Surface wave path.

failures. The profiles are roughly E–W and NE–SW oriented (Fig. 1). The profile to the west is more than 250 km long, whereas to the east and north-west most of the data suitable for evaluation are restricted to the first 150–200 km. The data are presented in the form of record sections (Figs 2a, 3, 4, 6 and 7) with a reduction velocity of 6.0 km s^{-1} for the vertical component and 3.5 km s^{-1} for the horizontal components. All profiles are unreversed.

P-wave interpretation

The results of the *P*-wave interpretation have been obtained by matching the interpreted correlations on the record sections with theoretical travel-time curves in a trial and error procedure. Synthetic seismograms have been computed to check the amplitude pattern of the phases, resulting in a refinement of the primary model obtained by travel-time methods.

The best information comes from the record section of the profile Toledo–Cáceres (Fig. 2a) where the main features are: (1) A clear P_g wave (refracted at the top of the basement), delayed by 0.8 s with respect to the origin, probably due to the presence of sediments around the shot point, which can be correlated from 10 to 110 km with a velocity of 6.1 km s^{-1}. (2) A phase arriving about 1 s later than P_g, correlatable between 30 and 130 km, interpreted as a reflection from an intracrustal discontinuity. (3) A phase interpreted as a reflection from an even deeper intracrustal discontinuity. (4) A conspicuous P_MP phase (reflected at the crust–mantle boundary) whose critical distance can be defined precisely, between 80 and 90 km, due to the sharp change in amplitudes between subcritical and over-

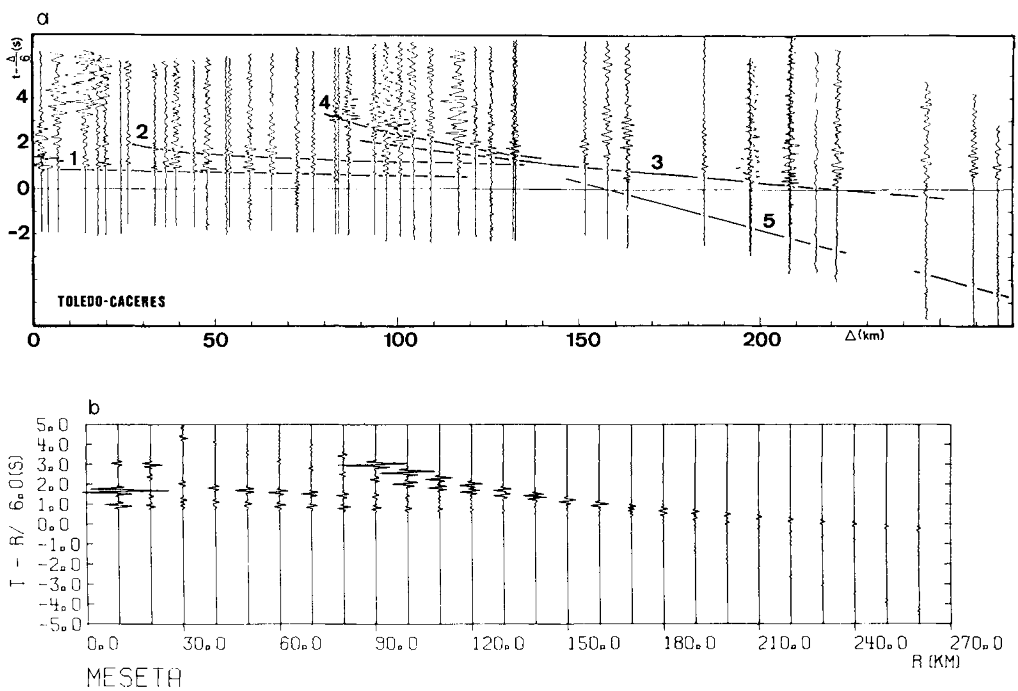

Figure 2.(a) Record section of the vertical component, reduced to 6 km s^{-1}, for the profile Toledo–Cáceres. (1) P_g phase. (2) Reflection at a discontinuity in the upper crust. (3) Reflection at the top boundary of the lower crust. (4) P_MP reflections. (5) P_n phase (see text for a more detailed explanation). (b) Synthetic section, computed using a ray method (Červený, Molotkov & Pšenčík 1977), corresponding to the model shown in Fig. 5 for a dominant frequency of 4.5 Hz.

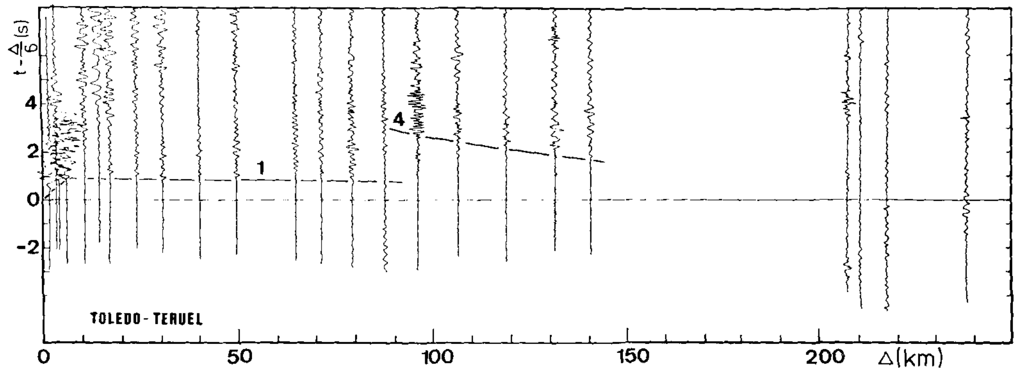

Figure 3. Record section of the vertical component, reduced to $6\,\mathrm{km\,s^{-1}}$, for the profile Toledo–Teruel. Phases are labelled as in Fig. 2(a).

critical reflections. (5) P_n onsets that are weak but visible in some seismograms giving an apparent upper mantle velocity of 8.0–$8.1\,\mathrm{km\,s^{-1}}$.

The profile Toledo–Teruel (Fig. 3) also shows a delayed P_g phase (between 10 and 90 km) due to the sedimentary basin extending from the shot point to the east. Intermediate phases are not visible and only P_MP reflections are correlated.

The profile Toledo–Soria (Fig. 4) shows an additional delay of the P_g beyond 50 km due to the thickening of the sedimentary cover in the Madrid basin. An alternative interpretation would be that the P_g disappears and a secondary phase takes over at about 70 km. We prefer the first because it fits better with the thickness of sediments reported in that region (Julivert *et al.* 1972). On this record section, the deep intracrustal reflection and the P_MP phase are visible between 80 and 200 km.

The mean velocity of the sediments is $3.3\,\mathrm{km\,s^{-1}}$, and from the intercept time of the P_g phase, their thickness is estimated to increase from 1.5 km at the shot point to about 3.0 km towards the east, when entering the Tagus basin. Towards the north the sedimentary thick-

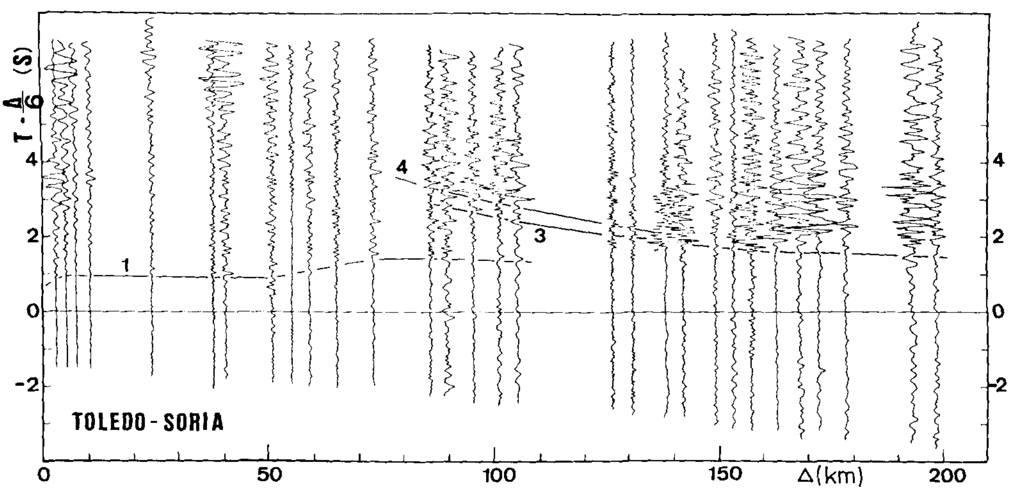

Figure 4. Record section of the vertical component, reduced to $6\,\mathrm{km\,s^{-1}}$, for the profile Toledo–Soria. Phases are labelled as in Fig. 2(a). Note the delay of P_g arrivals beyond the 50 km.

ness increases even more to about 4.0 km. To the west the basement outcrops at only 15 km distance from the shot point.

The crystalline basement has a mean velocity of 6.1 km s^{-1} (P_g phase), but a gradient from 6.05 to 6.15 km s^{-1} seems to fit better the amplitude ratio between the P_g phase and secondary arrivals. The basement overlies a low-velocity layer (5.6 km s^{-1}) detected between 7 and 11 km depth (Fig. 5a). The criteria to introduce such a velocity inversion come from the amplitude of the reflection, at the intermediate discontinuity, at about 40 km distance. Models without the inversion generate amplitudes far beyond 40 km. We have chosen the low-velocity layer interpretation because a high velocity underneath the basement would increase the thickness of the crust. As a result, the critical distances of the deep intracrustal reflection and the P_MP would be beyond the observed ones. An alternative to that is to disregard the deep intracrustal reflection but, in our opinion, these reflections are clear enough in the profiles Toledo–Cáceres and Toledo–Soria, in both P and S record sections, as to be taken into account. In Fig. 2(b) the synthetic section corresponding to our model shows a reasonable fit with the observed amplitude pattern. In addition, the S-wave interpretation supports the existence of such a low-velocity layer (see next section).

A layer with a velociy of 6.4 km s^{-1}, with its top at 11 km depth, reaches a depth of 24 km where another discontinuity is found thus subdividing the crust which in its lower part has a mean velocity of 6.9 km s^{-1}. As can be seen on the record section (Fig. 2a), the P_MP dies out at about 165 km which would not be the case with a lower crust having a constant velocity. One way to make the P_MP phase disappear (Banda 1979) is to introduce a low-velocity layer or a negative gradient in the lower crust. In our case a weak negative gradient from 6.9 to 6.8 km s^{-1} accounts for the lack of P_MP reflections at large distances (see synthetic section in Fig. 2a).

The Moho is reached at a depth of about 31 km which we believe to be a fairly accurate value due to the precise determination of the P_MP critical distance. The upper mantle velocity is not well defined because of the weak P_n arrivals, but a value of 8.0–8.1 km s^{-1} (phase 5, Fig. 2a) seems to fit the data and coincides with the values obtained in other Palaeozoic domains such as the Pyrenees or Brittany. The mentioned abrupt change of subcritical and overcritical amplitudes reflected at the Moho discontinuity is accounted for, in our model, by introducing a transition zone at the crust–mantle boundary rather than a first-order discontinuity which would generate visible subcritical reflections (Braile

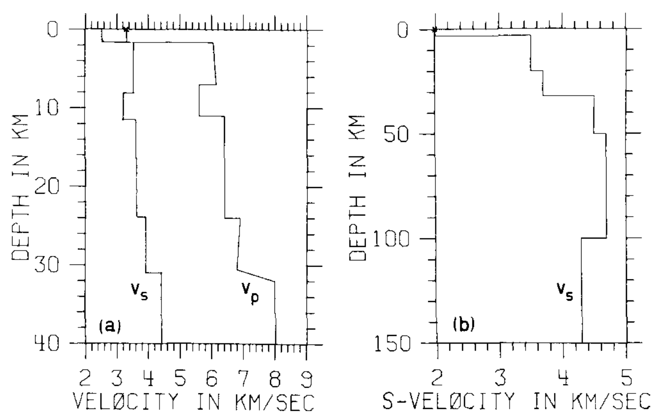

Figure 5.(a) P- and S-velocity–depth functions as derived from travel-time and amplitude interpretations. The corresponding synthetic sections is shown in Fig. 2(b). (b) Shear velocity–depth function obtained from surface-wave interpretation (see text for references).

& Smith 1975). In this case a transition zone of only 1.5 km thickness is sufficient as can be seen in the synthetic section of the proposed model (Fig. 2b).

S-wave interpretation

Transverse and in-line components of the profile Toledo–Cáceres are shown in Figs 6 and 7. It is obvious that it would be difficult to derive a crustal model from these record sections. Nevertheless we found it worth looking at the visible phases and, as a matter of fact, some information could be extracted. The main characteristics on the record sections are: (1) An S_g phase is barely seen. (2) A clear reflection, interpreted as the reflection at the top of the low-velocity layer, deduced from *P*-waves in the previous section, is visible between 20 and 100 km distance. (3) The reflection from the bottom of the low-velocity layer is not visible at all. (4) Reflections from the deep intracrustal discontinuity and the S_MS phase can be tentatively correlated.

The velocity–depth function deduced from the *P*-wave interpretation was used as a starting model, assuming a Poisson's ratio of $\sigma = 0.25$ ($v_p/v_s = \sqrt{3}$). This model was then modified until a reasonable fit of theoretical and interpreted travel-time curves was reached (Figs 6 and 7). The final model has only slightly different depths of the discontinuities with respect to the *P*-model (Fig. 5a) which we interpret as a confirmation of the model determined in the previous section. From the mean average velocity of the *P* and *S* models the v_p/v_s ratio turns out to be close to 1.75.

The most striking feature in the *S*-wave data is the energy which we interpret as returned from the top of the low-velocity layer. Usually the *P*-wave data show a strong reflection from the bottom of the low-velocity layer, whereas the reflection from the top is very seldom resolved because of the small difference in travel time as compared to the P_g-wave. In our case this holds for *P*-waves, but *S*-waves seem to show a stronger reflection from the top of the inversion zone. If this interpretation is correct, it could imply that the *S*-velocity distribution differs slightly from the *P*-velocity distribution, a detail which cannot be resolved with the present data.

Surface waves

The structure of the crust and upper mantle from surface-wave data has been reported in several papers (Payo 1975, 1970; Payo & Ruiz de la Parte 1974; Perez *et al.* 1978). In Fig. 1 the paths used for these studies are shown. Recently, Sierra (1980) has reinterpreted some of the lines and interpreted new data using multiple filtering technique to obtain more accurately the dispersion curve and to extend the period range suitable for inversion. Sierra (1980) used the phase velocity of teleseismic events for the path TOL–EBR and the group velocity of near earthquakes for the line Pyrenees–TOL (Fig. 1). The inversion of the data yielded a generalized model of the Iberian Massif having a sedimentary cover of 3 km overlying a crystalline basement with a shear velocity of 3.5 km s^{-1}. Nevertheless, Payo (1972) found evidence to include a low-velocity of 3.3 km s^{-1} and the thickness of the entire crust has a value of about 30 km. The *S*-velocity of the subcrustal lithosphere is reported to be 4.5 km s^{-1} increasing to 4.7 km s^{-1} at about 50 km depth. The lithosphere–asthenosphere boundary is reached at about 100 km depth, where the shear velocity decreases to 4.2–4.3 km s^{-1}. A similar lithospheric thickness of about 90 km is reported by Panza *et al.* (1980) from a unified interpretation of surface-wave data in Western Europe. The *S*-velocity–depth function of the lithosphere in the Iberian Massif is shown in Fig. 5(b).

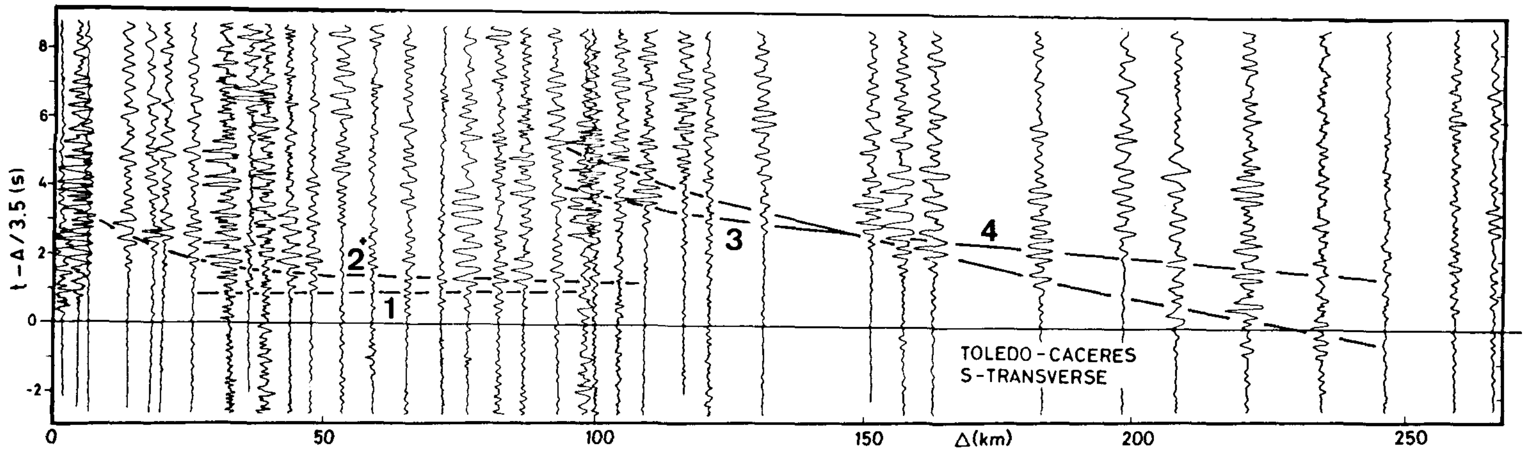

Figure 6. Record section of the transverse component, reduced to 3.5 km s^{-1}, for the profile Toledo–Cáceres. (1) S_g phase. (2$^+$) Reflection at the top of the low-velocity layer. (3) Reflection at the top boundary of the lower crust. (4) S_MS phase.

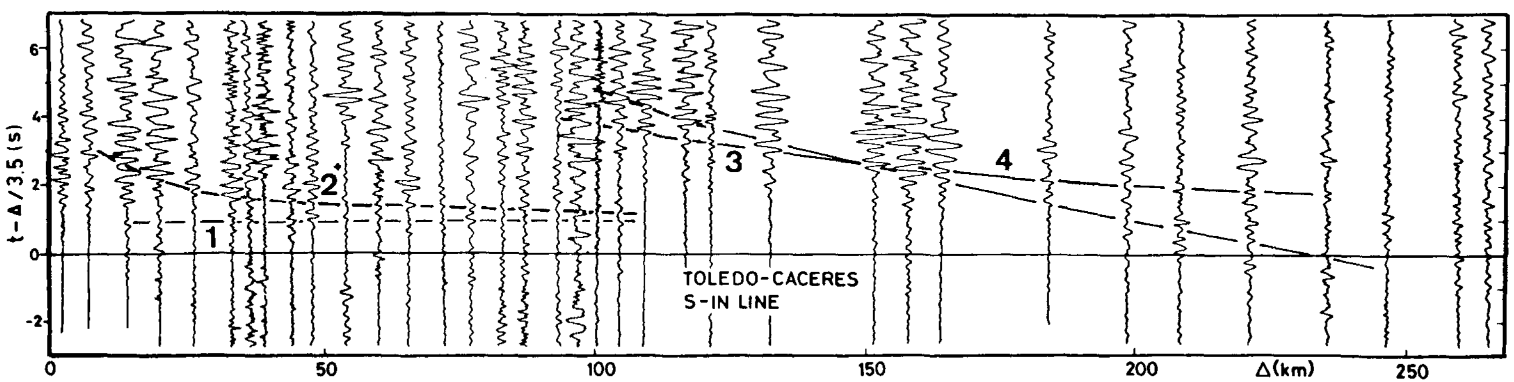

Figure 7. Record section of the in-line component, reduced to 3.5 km s^{-1} for the profile Toledo–Cáceres. Phases are labelled as in Fig. 6.

Taking into account the resolution of surface-wave studies, these results can be considered to be consistent with those obtained by explosion seismology. Only the values of the sub-Moho velocities seem to be different if one assumes Poisson's ratio (σ) of 0.25. Our seismic data do not define an S_n velocity with precision, but a value of $4.4\,\mathrm{km\,s^{-1}}$ fits the observed critical distance assuming the Moho depth derived from the *P*-wave interpretation. This value is even lower than the $4.5\,\mathrm{km\,s^{-1}}$ reported from surface waves. Our *P*-wave data are incompatible with such an *S*-velocity assuming $\sigma = 0.25$, which leads us to suggest that a Poisson's ratio higher than 0.25 and closer to 0.3 ($v_p/v_s = 1.80$) holds for this region of the uppermost mantle immediately beneath the crust–mantle boundary.

Earthquake travel-time data

Payo (1972) published a compilation of earthquake travel-time data which he used to determine crustal velocities between the epicentral areas (Gulf of Cadiz and the Alboran Sea) and the stations PTO, TOL and MAL (Fig. 1). Payo finds velocities ranging from 5.5 to $5.9\,\mathrm{km\,s^{-1}}$ for the P_g phase, whereas our data indicate a P_g velocity of $6.1\,\mathrm{km\,s^{-1}}$ with an error which is not likely to be greater than $0.1\,\mathrm{km\,s^{-1}}$. It is not the first time that such a discrepancy between earthquake data and explosion seismology has been noted. Besides the inaccuracy of the epicentral parameters, one explanation could be given in terms of Gutenberg's (1954) ideas that the earthquake waves generated at some depth in the crust produce as a first arrival a wave which is channeled in a low-velocity layer; some energy leaks out and is observed at the surface. This could possibly be the case since we have indications of the existence of a low-velocity layer in both the southern part of the Iberian Peninsula (Mueller *et al.* 1973; Banda & Ansorge 1980) and in the Iberian Massif (this paper).

Velocities of $6.3\,\mathrm{km\,s^{-1}}$ in the middle crust and 6.6–$6.9\,\mathrm{km\,s^{-1}}$ in the lower crust, reported by Payo (1972) are in fairly good agreement with our results.

The most important disagreement between Payo's results and those from explosion seismology arises from the velocity immediately below the Moho. Payo reports an average velocity of 7.6–$7.65\,\mathrm{km\,s^{-1}}$, which, as pointed out before, lies well beyond the possible error of the P_n velocity obtained in our study. One could claim that the P_n onsets are not clear on the record sections, but the critical distance of the P_MP, well documented in our data, would fall far beyond the observed one, if a P_n velocity of $7.6\,\mathrm{km\,s^{-1}}$ is assumed (more than 120 km distance against the 80–90 km observed). Therefore we believe that the average *P*-velocity inferred by Payo, in his travel-time data, is strongly influenced by the epicentral areas, Gulf of Cadiz and Alboran Sea, where a low-velocity upper mantle is known to exist from the results of refraction experiments (Purdy 1975; Hatzfeld 1978). Possible strong dips in the Moho are also excluded from our data and from the smoothness of the Bouguer anomaly along the profiles.

Conclusions and petrological interpretation

Explosion seismology and earthquake data dealt with in this paper lead to suggest a lithospheric structure for the Iberian Massif whose main features are (Fig. 8):

(a) The thickness of the crust is about 31 km, with an average velocity of $6.2\,\mathrm{km\,s^{-1}}$ and a v_p/v_s ratio of about 1.75 ($\sigma = 0.26$).
(b) The crystalline basement has a mean velocity of $6.1\,\mathrm{km\,s^{-1}}$ with a slight positive gradient and possibly overlies a low-velocity layer ($5.6\,\mathrm{km\,s^{-1}}$) situated between 7 and 11 km depth.
(c) The middle crust is about 12 km thick with a mean velocity of $6.4\,\mathrm{km\,s^{-1}}$.

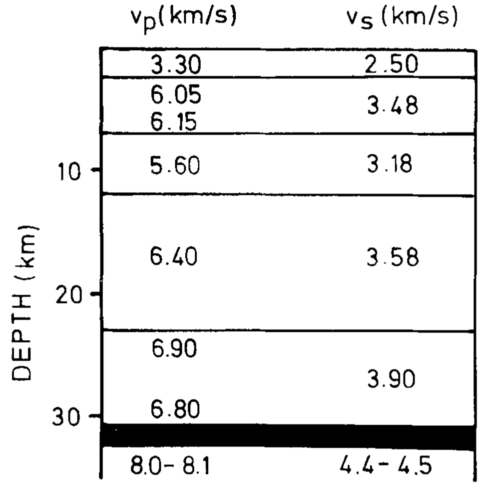

Figure 8. Schematic model of the crust in the Iberian Meseta.

(d) The lower crust is some 8 km thick and has an average velocity of 6.9 km s^{-1} with a possible negative gradient.

(e) The Moho is a transition zone at least 1.5 km thick.

(f) The *P*-velocity below the Moho is 8.0–8.1 km s^{-1} and the *S*-velocity 4.4–4.5 km s^{-1} which suggests v_p/v_s close to 1.8.

(g) The asthenosphere is reached at about 90–100 km depth where the shear velocity decreases from 4.7 to 4.2–4.3 km s^{-1}.

The general features of the crustal structure of other Hercynian provinces in Western Europe, such as Brittany (Sapin & Prodehl 1973), show a remarkable coincidence with our model, excepting the upper crustal low-velocity zone (see a compilation of crustal structure of Hercynian domains in Gallart, Banda & Daignieres 1981). However, the structure deduced in Portugal is somewhat different (Mueller *et al.* 1973; Mendes Victor *et al.* 1981). This can be explained because the crust in Portugal has been modified during its rift flank position during the initial stage of the opening of the Atlantic or possibly because of its frontal position at the end of the Hercynian orogeny.

From the seismic model one is tempted to interpret the distribution of layers and velocities from a petrological point of view. The surface geology of the Iberian Meseta shows that Tertiary and Mesozoic sediments unconformably overlie low grade Palaeozoic metamorphic rocks, basically shales and quartzites which could be the seismic layer with velocity 6.1 km s^{-1}. Underneath the basement, where a low-velocity zone has been proposed, granitic material, which outcrops in several places of the Iberian Massif, may be present, as suggested by Mueller (1977) in a generalized model of the continental crust. The presence on the surface of rocks such as migmatities, adamellites (quartz-monzonite) and granodiorites, suggest that the middle crust, the seismic layer with velocity 6.4 km s^{-1}, may have such composition (Mueller 1978). In fact, paragenetic data from those migmatitic facies indicate a 12 km depth origin for them (Aparicio 1971), a depth which coincides roughly with the 11 km estimated for the interface between the low- and high-velocity layers. Finally, the lower crust, according to the model of Smithson & Brown (1977), would consist of different igneous and metamorphic rocks, in our case, granodiorites, diorites and granulites. The Moho, in this model a transition zone 1.5 km thick, would then be the boundary between the crust and a perioditic mantle. We are aware that this petrological interpretation is rather speculative, and more detailed studies from both seismic and petrological points of view are necessary to check it.

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