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Crustal structure under the central and eastern part of the Betic Cordillera

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Summary. In 1974 and 1975 a network of crustal seismic refraction profiles across the Betic Cordillera was established with shotpoints at sea off the Spanish coast. The longest profile with a length of 440 km runs parallel to the general strike through the centre of the Betic Cordillera. Two other profiles lie south of the main profile closer to the coast and perpendicular to the east–west strike, respectively.

The interpretation of *P* waves indicates pronounced lateral variations of the structure and thickness of the crust and upper mantle. Two different crustal blocks seem to exist on either side of the north-east trending Palomares and Alhama de Murcia fault system in the eastern Betic Cordillera.

East of this fault system the crust is only 23 km thick with a shallow high velocity layer of 6.9 km s^{-1}. West of this line in the central and southern part of the Betic Cordillera the crust–mantle boundary lies at depths of 39 km and 24 km, respectively. In this area a velocity of 6 km s^{-1} is reached at about 3 km depth. A layer of reduced *P*-wave velocity follows between 7 and 12 km depth with a velocity of 5.4 km s^{-1}. At its lower boundary the velocity increases to 6.2 km s^{-1} and reaches values between 6.6 and 6.7 km s^{-1} at a depth of 18 km under the south coast and 24 km under the centre of the Betic Codillera east of Granada. A thin lid of about 6 km thickness and a P_n velocity of 8.1 km s^{-1} lies on top of a rather thick layer with the reduced velocity of 7.8 km s^{-1}. At a depth of 63 km the velocity increases again to 8.3 km s^{-1}. Deep-reaching structural differences mark the east–west trending boundary between the internal and external zones of the central Betic Cordillera.

Introduction

The Betic Cordillera in southern Spain and the Rif Mountains in northern Africa in conjunction with the Alboran Sea, which they embrace, represent the westernmost part of the Alpine chain in Europe. The continental part of the Iberian plate interacts with the African plate in a highly complicated manner which is not yet completely understood. A great variety of studies has been published about the tectonic evolution of the western Mediterranean Sea and the surrounding areas. For some areas quite clear concepts seem to exist, e.g. the development of the Provence basin and the South Balearic Sea as shown by Auzende, Bonnin & Olivet (1973). A detailed study of the geodynamic structure of the areas surrounding the Balearic Islands has been given by Mauffret (1976). Boccaletti & Guazzone (1974) have presented a whole sequence of palinspastic maps in terms of marginal basins and remnant island arcs for the whole western Mediterranean. Whereas there is less disagreement about the tectonic development of the western Mediterranean basins, quite different models have been proposed for the Alboran Sea and the bordering mountain ranges in northern Africa and southern Spain, i.e. the Rif Mountains and the Betic Cordillera. Araña & Vegas (1974) proposed a subduction of oceanic lithosphere towards the north under the Iberian plate in late Cretaceous to early Miocene time from the chemistry of Tertiary and Quaternary volcanism in south-eastern Spain. For a later stage Andrieux, Fontboté & Mattauer (1971) and Andrieux & Mattauer (1973) postulate an Alboran subplate which separates the African and European plates by its relative westward movement. This may explain the folding and shortening of external and internal zones on either side of the Alboran Sea. Subsequently, Le Pichon, Pautot & Weill (1972), Auzende *et al.* (1973) and Olivet, Auzende & Bonnin (1973) propose the opening of the two Alboran basins by the breaking of the Alboran plate into two parts along faults trending ENE–WSW. From a new map of aeromagnetic anomalies Galdeano *et al.* (1974) deduce the opening of the Alboran Sea along NNE–SSW trending faults, i.e. features perpendicular to the previous mechanism.

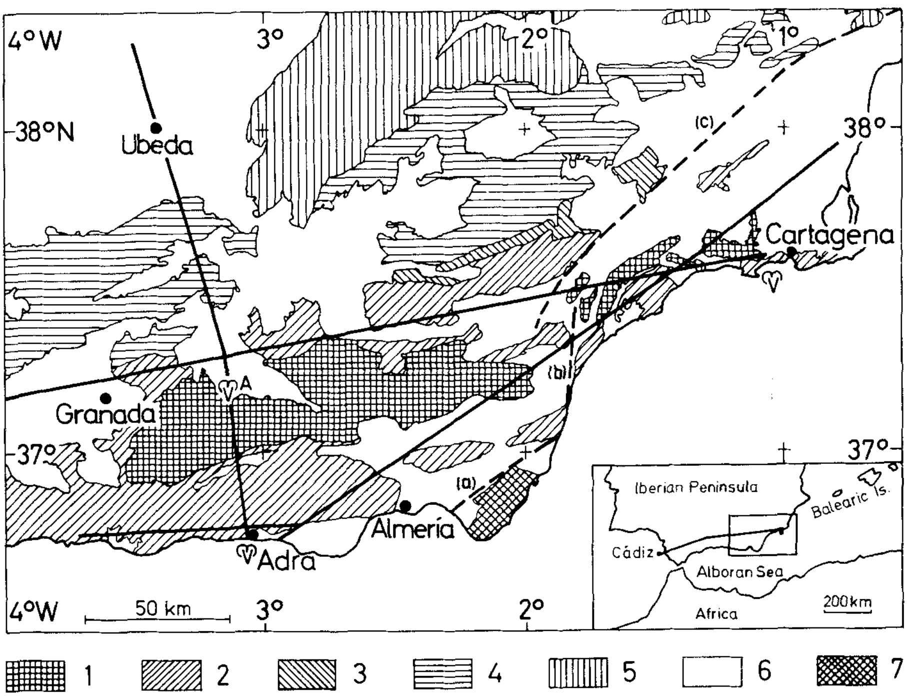

Figure 1. Location of shotpoints Adra, Alquife (A), Cádiz, Cartagena and profiles in the eastern part of the Betic Cordillera of Spain. Tectonic units after Julivert *et al.* (1972). Internal zones: (1) Nevado–Filabride complex; (2) Alpujarride complex; (3) Malaguide complex. External zones: (4) Subbetic zone; (5) Prebetic zone; (6) Post-orogenic sediments; (7) Volcanic rocks, (a) Carboneras fault, (b) Palomares fault, (c) Alhama de Murcia fault after Bousquet & Philip (1976).

According to Loomis (1975) the Alboran Sea opened by the counter-clockwise rotation of the Iberian Peninsula resulting in an oceanization of the crust between Spain and Africa. From the distribution of focal mechanisms and the seismicity in the Alboran region, Udías, López-Arroyo & Mezcua (1976) considered the previously mentioned Alboran plate as a buffer plate between Iberia and the African plate. But no interpretation of the Betic orogeny could be given.

In the present paper we report the latest results of seismic refraction observations derived from several profiles in the central and eastern part of the Betic Cordillera. The data presented are part of a larger project which comprises the entire Betic Cordillera in southern Spain from Cádiz to Cartagena (Fig. 1), the Balearic Islands and to some extent the Alboran Sea. Earlier interpretations have been given by the Working Group for Deep Seismic Sounding in Spain 1974–1975 (1977), Ansorge *et al.* (1976, 1977) and Banda (1979). Udías (1977) has published an extensive report about the experiments containing technical and preliminary scientific results together with a complete list of references on this subject. The Working Group for Deep Seismic Sounding in the Alboran Sea – 1974 (1978), Hatzfeld (1976, 1978) and Boloix & Hatzfeld (1977) have determined the crustal structure in the Alboran Sea from seismic refraction observations and other geophysical data.

Geological setting

Since we will refer to some details of the geological structure in the later discussion the main characteristics of the geology in the eastern part of the Betic Cordillera are summarized briefly. The Betic Cordillera can be divided into two major zones (Fig. 1): (1) The internal or Betic zone in the south and (2) the external zone to the north, which in itself is subdivided into the Subbetic zone and the Prebetic zone bordering the Hercynian Massif of the Spanish Meseta. The present investigation is restricted to the eastern part of the Betic Cordillera with the main part of the refraction lines lying in the internal zone. In this area the internal zone can be grouped into the three main complexes of Nevado–Filabride, Alpujarride and Malaguide with the highest grade of metamorphism in the lowest complex, i.e. the Nevado–Filabride (Egeler & Simon 1969a). All three complexes consist of Permo-Triassic sediments with additional rocks of pre-Permo-Triassic age in the Alpujarride and Malaguide complexes. The individual complexes show a nappe-type structure. The direction of thrusting is still debated, i.e. whether a north- to north-westward overthrusting (Egeler & Simon 1969b; Fontboté 1970) or a south to south-eastward overthrusting (Durand Delga 1966) led to the present tectonic situation of geological complexes. Kampschuur & Rondeel (1975) propose a two-phase development. An older deformation of Mesozoic age produced NW–SE fold axes followed by a younger period with NE–SW to E–W axes of Neogene age. According to Kampschuur & Rondeel (1975) only the younger Neogene NE–SW trending deformations are associated with the convergence of the African and European plates.

The experiments

The locations of shotpoints, refraction profiles and observation points are shown in Figs 1 and 2. All the data which will be discussed here were obtained during two field campaigns in 1974 and 1975. A detailed description of these experiments including shot sizes, exact locations of explosions and experimental procedure has been given by the Working Group for Deep Seismic Sounding in Spain 1974–1975 (1977). Except for the explosions at Alquife (A) east of Granada (Fig. 1) all shots were fired offshore by the Spanish Navy which also provided the explosives. The charge size ranged from single depth charges with 136 kg of explosives to composite units of depth charges of up to 1632 kg of explosives.

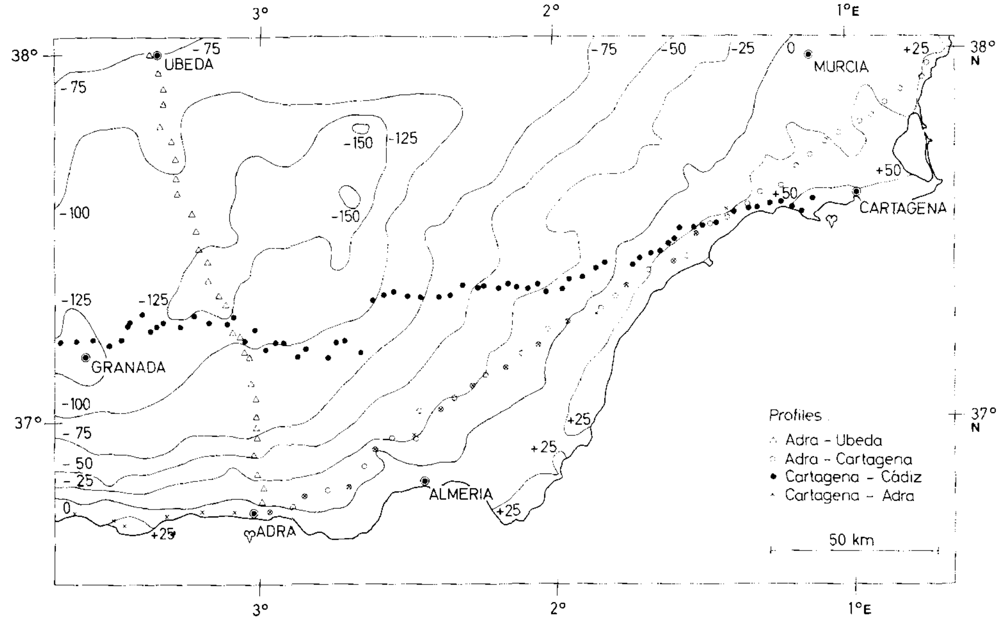

Figure 2. Map of Bouguer anomalies as determined by the Instituto Geográfico Nacional Madrid (unpublished) in the area under investigation together with the exact location of observation points and shotpoints Adra and Cartagena.

The explosions off the coast of Cádiz (Fig. 1) could not be observed beyond a distance of 130 km east of the shotpoint. A deep basin of young and partly unconsolidated sediments in addition to a steep rise of the crystalline basement off the coast caused this high attenuation. Therefore, this investigation is restricted to the central and eastern part of the Betic Cordillera. More detailed data have been obtained westwards from Adra to Cabo de Trafalgar south of Cádiz in 1977 which are the subject of a separate study (Ansorge & Banda 1980).

The two main profiles for the crustal structure of the internal zone from Cartagena to Cádiz and Cartagena to Adra run more or less parallel to the general strike of the tectonic units (Fig. 1). Beyond a distance of 240 km the profile from Cartagena to Cádiz enters the external zone. It was tried to reverse the southern profile from Adra to Cartagena. However, because of the off-line position of the shotpoint near Cartagena, this profile cannot be treated as exactly reversed. The profile from Adra to Ubeda (Fig. 1) crosses the Alpujarride and Nevado–Filabride complexes from south to north and reaches the external zone at a distance of about 65 km.

Three explosions in an iron ore mine (Minas del Marquesado, C.A.M.) near Alquife (shotpoint A in Fig. 1) were recorded to a distance of 50 km close to the main profile along the northern edge of the Nevado–Filabride complex to the east and west.

The record sections Cartagena–Cádiz, Cartagena–Adra, Adra–Cartagena, Adra–Ubeda and Alquife are shown with a reduced time-scale (reduction velocity 6 km s^{-1}) in Figs 2–8. The correlated travel-time curves will be discussed in detail subsequently. All the analogue tape records, except for the Alquife observations, were digitized and plotted with the amplitudes normalized to a maximum value in each trace.

Fig. 2 shows the newly surveyed Bouguer anomaly by the Instituto Geografico Nacional Madrid (unpublished) for the area of this investigation together with the location of seismic observation points. The minimum of the gravity anomaly lies north of the main profile from Cartagena to Cádiz, whereas the profile Adra–Ubeda more or less traverses the minimum

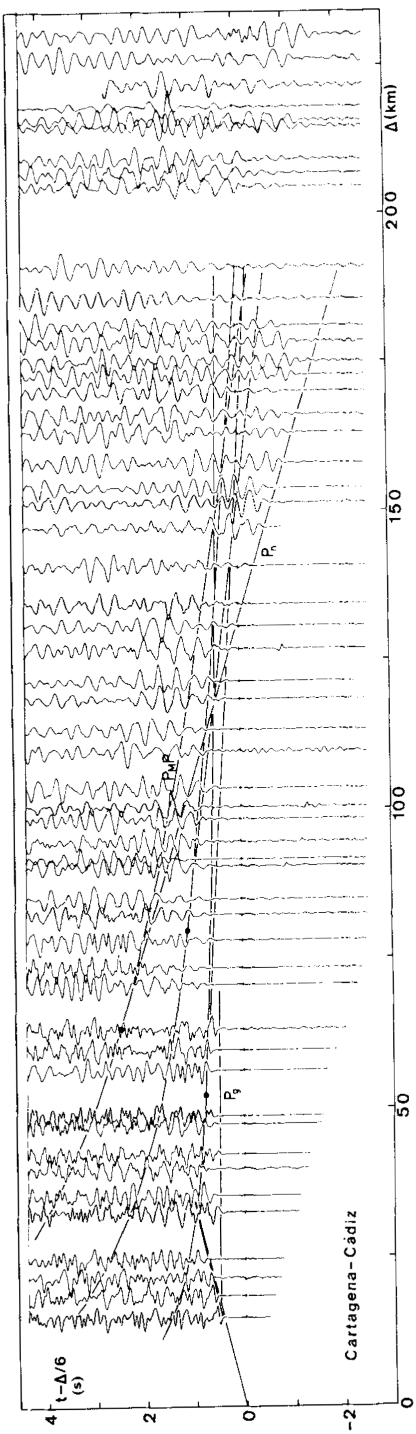

Figure 3. Record section Cartagena – Cádiz plotted with a reduction velocity of 6 km s^{-1}, amplitude normalized. Travel-time correlations are shown.

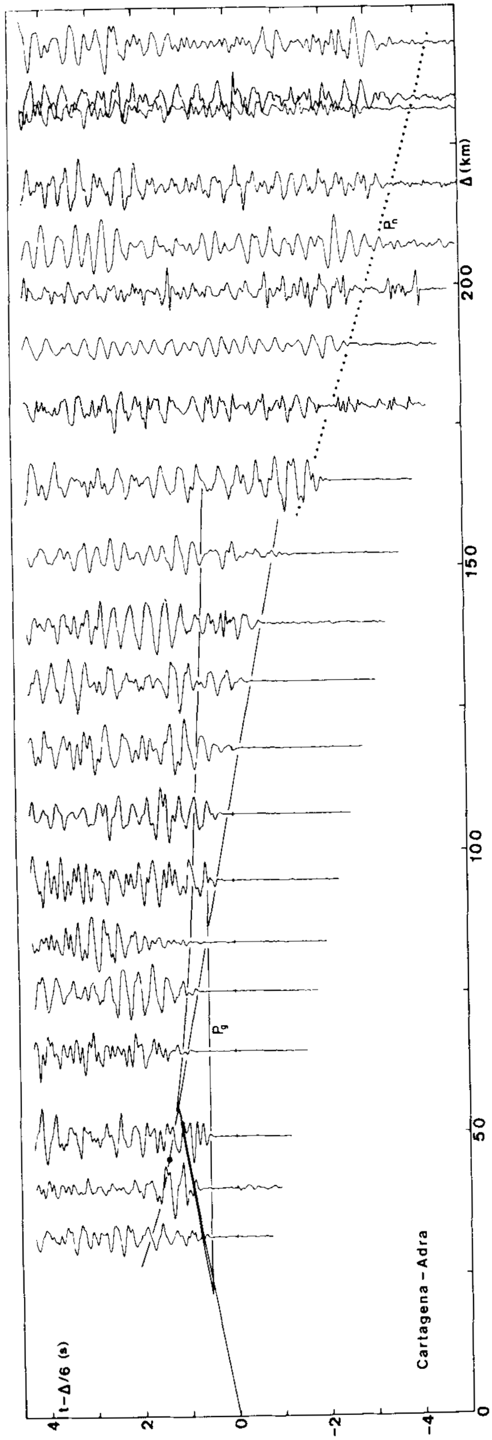

Figure 4. Record section Cartagena – Adra with travel-time correlations.

Bouguer anomaly of −125 to −150 mgal. The contour lines of the gravity anomaly lie almost parallel to the coast line with an increasing negative gradient towards the interior of the Betic Cordillera from north-east to south-west.

Interpretation

THE UPPER CRUST

Several main features of the structure which reflect the rather complicated tectonic history seem to affect the first arrivals or phases which are refracted from the crystalline basement (P_g). Therefore, these phases will be discussed in more detail in this section.

CARTAGENA–CÁDIZ (FIGS 3 AND 7a)

Up to a distance of 65 km this profile traverses several small outcrops of the Nevado–Filabride and Alpujarride complexes and a few Neogene–Quaternary sediment basins of minor thickness. As can be seen from Figs 3 and 7(b) the arrival times of the P_g phase vary considerably along the profile with differences up to 0.3 s. The individual changes in arrival time cannot be correlated with the local sediment basins but occur also when stations are situated on outcrops of the older Triassic complexes. The average apparent velocity is 6.05 km s^{-1} between 30 and 65 km distance. The pronounced scatter in the P_g arrival times can be explained by the highly fractured substratum in this area which includes also the Triassic outcrops as postulated by Fontboté (1977).

At a distance of 65 km (Fig. 7a) a new phase can be observed with an apparent velocity of 6.2 km s^{-1} and a delay of about 0.3 s. In addition the spectral analysis of seismograms at this distance range shows a sudden change of the dominant frequencies from 5 to 3 Hz for the same shot (Banda 1979). This change is regarded as strong evidence for a lateral variation of the crustal structure in the area where the profile traverses the Alhama de Murcia fault system (Bousquet 1979). Further details about this fault system are given in the discussion later on.

The inversion of travel times up to a distance of 65 km lead to the velocity–depth distribution in Fig. 10 (dashed line). Because of the lack of data up to a distance of 15 km probably too high a velocity has been chosen for the uppermost kilometre as indicated by the travel-time curve in Fig. 7(a).

CARTAGENA–ADRA (FIG. 4)

The first three records on the profile Cartagena–Adra (Fig. 4) show the same scatter of the P_g phase as on the profile Cartagena–Cádiz (Fig. 3). This is not surprising because the same area was covered by the first part of the profile (see Figs 1 and 2).

ADRA–CARTAGENA (FIG. 5)

Beyond 40 km the P_g phase can be correlated with an apparent velocity of 6.05 km s^{-1} out to a distance of about 100 km. The observation points on the first 40 km of this profile were located on the Neogene sediments around Almería. This causes a clear delay of the P_g phase as compared to greater distances (Fig. 5).

ADRA–UBEDA (FIGS 6 AND 7b)

This is the only profile which crosses the Betic Cordillera from south to north perpendicular to the tectonic strike (Fig. 1). After traversing the Alpujarride and Nevado–Filabride com-

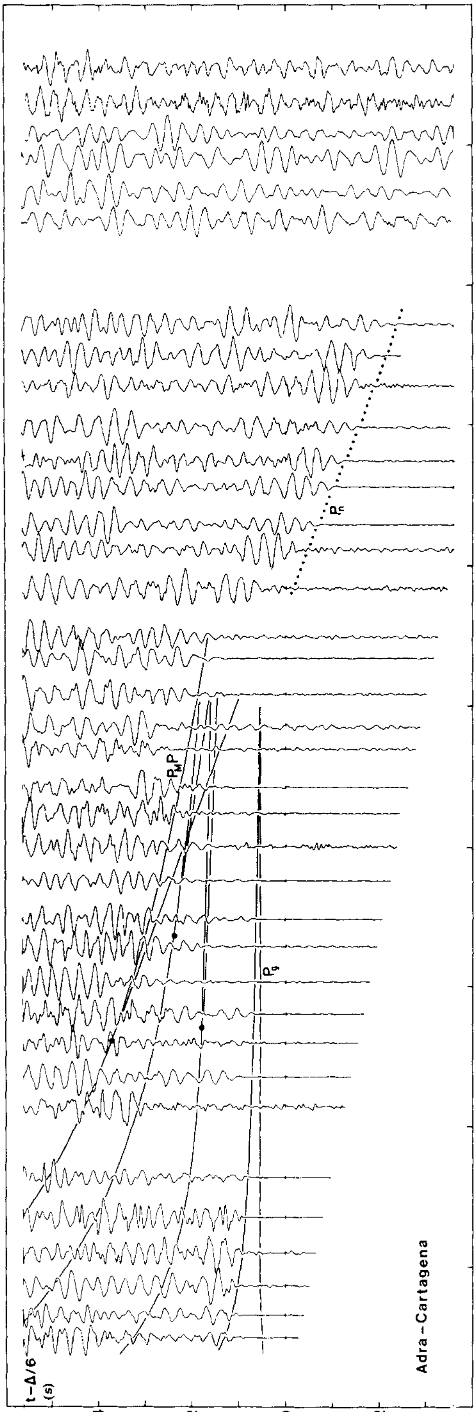

Figure 5. Record section Adra–Cartagena with travel-time correlations.

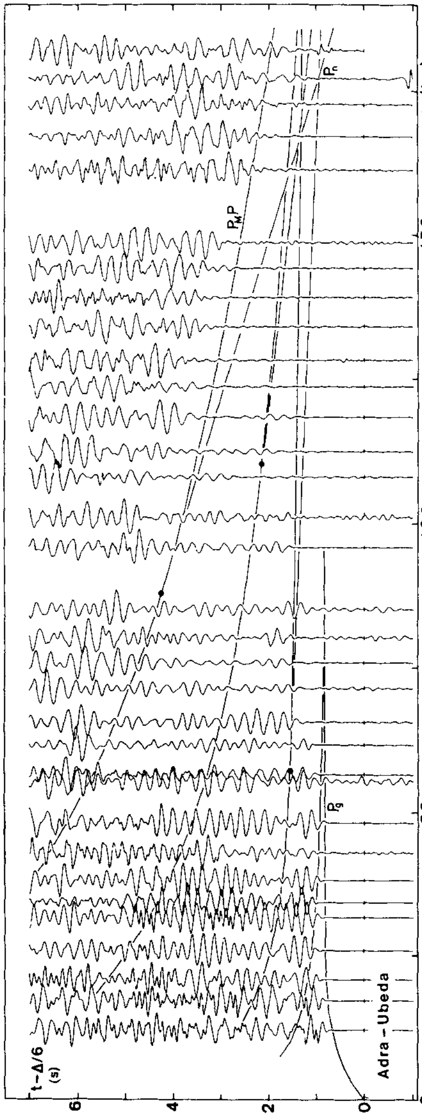

Figure 6. Record section Adra–Ubeda with travel-time correlations.

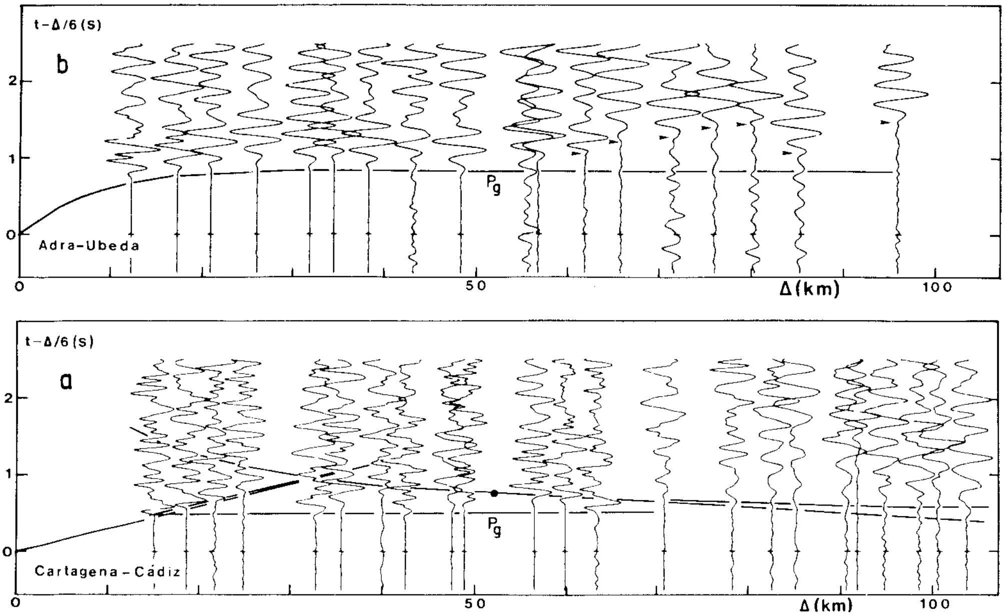

Figure 7. Record sections of the profiles (a) Cartagena–Cádiz and (b) Adra–Ubeda with travel-time correlations of upper crustal phases with an expanded time and distance scale. Note change of dominant frequencies around 65 km on profile Cartagena–Cádiz (a) and delay of P_g phase beyond 60 km distance on profile Adra–Ubeda (b).

plexes the external zone is reached at a distance of about 70 km from the shotpoint near Adra in the northern part of the Guadix basin. Very clear P_g arrivals can be correlated to a distance of about 60 km which allow a more precise determination of the velocity–depth distribution in the upper crust of this area by using both travel times and amplitudes of these phases. In a first step velocity–depth distributions have been determined from first arrival times with and without topographic corrections using the Herglotz–Wiechert method which leads essentially to the same model. In the following, synthetic seismograms have been computed by the ray-theoretical method (Červený, Molotkov & Pšenčík 1977). The final model from amplitude interpretations is also in agreement with the one determined previously from travel times only, but gives a much more precise velocity–depth gradient. Fig. 9 compares observed and theoretical amplitudes with the scatter of the observed amplitudes indicated by the shaded area. The continuous line in Fig. 9 corresponds to the velocity–depth distribution for the uppermost crust under profile Adra–Ubeda as depicted in Fig. 10. A change of the mean velocity gradient by 0.01 s^{-1} would cause an amplitude pattern which lies considerably outside the shaded area in Fig. 9. The well-documented

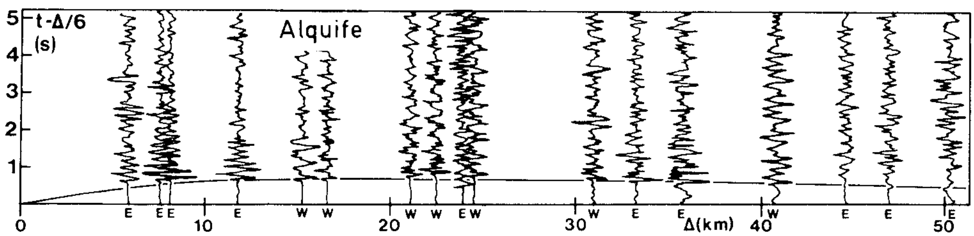

Figure 8. Record section Alquife with travel-time correlation. E, W denote observation points east and west of the shotpoint, respectively.

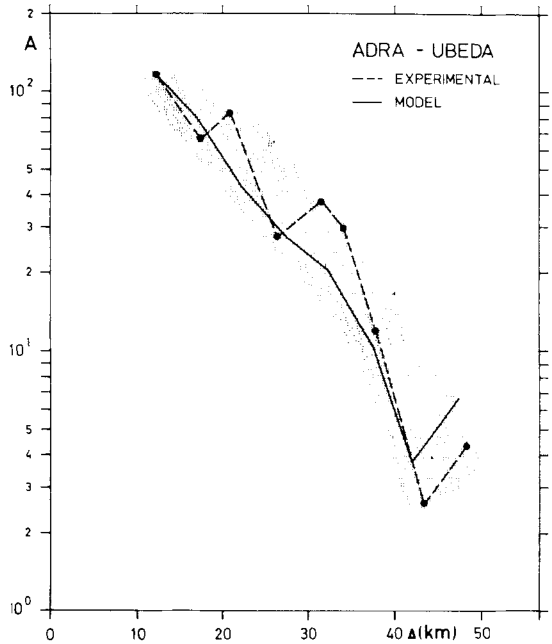

Figure 9. Amplitude–distance plot for P_g phase of the profile Adra–Ubeda. Dashed line denotes observed values, continuous line denotes the deduced model in Fig. 10.

model for this profile might indicate that all the deduced velocity–depth distributions under the other profiles may perhaps show velocities which are slightly too high. The convex shape of the amplitude distance curve of P_g phases in Fig. 9 is probably caused by the metamorphosed Permo-Triassic or older sedimentary layers along the profile Adra–Ubeda. In exposed crystalline basement the P_g amplitudes decrease normally with Δ^{-2} or less depending on the variation of velocity with depth.

Beyond a distance of 60 km, i.e. at about the distance where the profile enters the external zone the P_g arrivals are increasingly delayed as can be seen from Fig. 7(b). This delay of about 0.45 s is too high to be accounted for only by the Neogene sediments of the Guadix basin. A geological cross-section along this profile by Orozco (1978) shows a relatively thin cover of sediments overlying a deep-reaching fracture zone representing the contact between the internal and external zones.

ALQUIFE (FIG. 8)

Three explosions were recorded along an east–west profile to distances of 50 and 40 km, respectively. Most of the stations were located on the southern boundary of the Guadix sedimentary basin or directly on outcrops of the Nevado–Filabride complex with no pronounced structural variations along both directions (Fig. 1). Therefore, all the seismograms of these explosions were combined in the record section of Fig. 7. The location of the recording stations are denoted by capital letters for east and west, respectively. No significant difference between the two directions seem to exist in the travel times of P_g phases if one allows for some scatter. These data provided a control of the P_g-velocity distribution in the centre of the Betic Codillera as shown in Fig. 10. Whether the slightly higher P_g velocities under this profile compared to those under Cartagena–Cádiz and Adra–Ubeda are significant cannot be decided at this point.

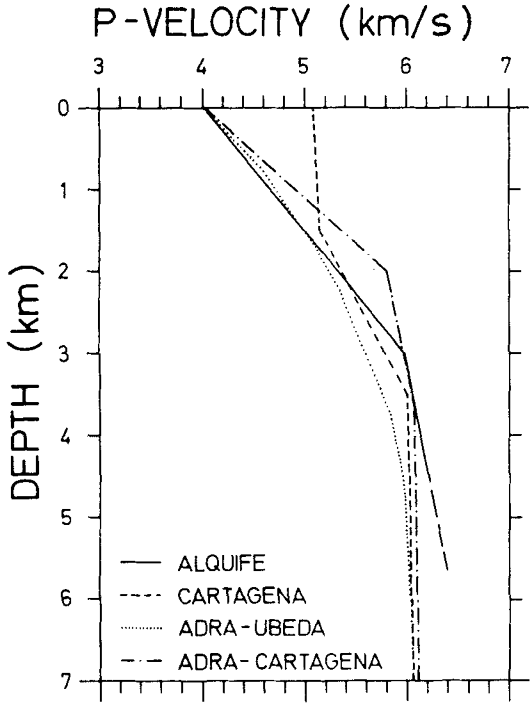

Figure 10. *P*-wave velocity distributions with depth in the crystalline basement for several profiles.

THE LOWER CRUST

The later phases which were correlated in the record sections of Figs 3–6 will be discussed together with the derived crustal structures in this chapter. Several inversion methods were applied to determine the crustal structure from the observations. In a first approximation plane horizontally layered models were assumed to evaluate the initial correlations on the individual record sections. The combination of adjacent or cross profiles leads to a considerable change of the correlations and to models with sloping interfaces. Synthetic seismograms were computed for those sections of the profiles where horizontal layering could be assumed. Finally a ray-tracing method was applied for areas with strong lateral variations (Červený, Langer & Pšenčík 1974).

A detailed comparison of the record sections in Figs 3–6 clearly shows the strong lateral variation of the crustal structure within short distances and thus only a rather generalized picture of the crustal structure can be derived. The method of shooting offshore at optimum water depth for a given amount of explosives in order to increase the efficiency (Wielandt 1975) decreases the peak frequency of the seismic signal to between 3 and 4 Hz with the charge sizes used in this experiment as compared to signals between 8 and 10 Hz from borehole shots. Therefore, the resolving power for structural details is considerably reduced. In general reflected and refracted phases from four distinct interfaces were correlated on the record sections of Figs 3, 5 and 6. These correspond to the crystalline basement which has been discussed already, two intermediate crustal interfaces and the crust–mantle boundary.

CARTAGENA–CÁDIZ

On the profile Cartagena–Cádiz (Fig. 3) the P_g phase dies out at a distance of about 65 km, i.e. beyond the Palomares fault (Fig. 1). A second phase with a lower dominant frequency can be correlated from first arrivals with a velocity of 6.3 km s^{-1} to a distance of 125 km when it disappears in the crossing P_n phase. A sudden increase in amplitudes about 0.5 s later indicates a third phase with an apparent velocity of 6.7 km s^{-1} corresponding to an intermediate layer in the lower crust. The computation of synthetic seismograms for this part of the profile has proven that a layer of 6.7 km s^{-1} is necessary in the lower crust in order to obtain the observed amplitudes. Very clear P_n arrivals together with the over-critically reflected phase P_MP are associated with the crust–mantle boundary indicating that this fourth interface is relatively sharp. The low apparent velocity of 7.7 km s^{-1} for the P_n wave from 80 to 180 km distance indicates a rather strong downdip of the Moho along this section of the profile, which is situated entirely in the internal zone. The sudden decay of the P_n phase beyond 180 km is most probably caused by the structure of the upper mantle as will be discussed later.

CARTAGENA–ADRA–CARTAGENA

A record section of completely different character was obtained from the same shotpoint along the south-eastern coast from Cartagena to Adra (Fig. 4). Besides the poorly documented P_g the most conspicuous phase with a velocity of 6.9 km s^{-1} can be correlated from first arrivals between distances of 80 and 150 km, a fact which is only observed on this profile and which is another evidence of lateral variations of the crustal structure. Beyond 150 km high-amplitude P_n arrivals are correlated to the end of the profile at 250 km, but there is no clear P_MP phase corresponding to it as in the previous profile. It is not possible to treat this section and the one from Adra–Cartagena as reversed profiles because of the position of the shotpoint Cartagena relative to the profile (Figs 1 and 2). Unfortunately, this record section has twice the station spacing of the one from Adra to Cartagena because one of the planned explosions for this profile had to be cancelled due to bad weather conditions.

The profile Adra–Cartagena (Fig. 5) shows again four distinct phases to a distance of 120 km, when a strong lateral change in the crustal structure seems to occur which causes the sudden lack of energy and the appearance of strong P_n phases beyond 150 km. A high-amplitude second phase follows the P_g phase by 1.1 s. The extended overlap and the strong under-critical reflections from the bottom of a low-velocity zone lead to a velocity of 5.4 km s^{-1} for this material on top of the layer with 6.2 km s^{-1}. Prior to the P_MP phase again arrivals with a velocity of 6.7 km s^{-1} indicate a high-velocity layer in the lower crust as observed along the profile Cartagena–Cádiz. The P_MP phase itself is fairly clear, but the P_n arrivals observed beyond 150 km do not line up tangentially with it, which again indicates deep-reaching structural variations across the Palomares fault zone (Fig. 1).

Three steps in the interpretation have been taken in order to obtain crustal structures for the two profiles Cartagena–Adra and Adra–Cartagena. First, a model has been deduced for the profile Adra–Cartagena which fits both travel times and amplitudes by computing synthetic seismograms for the distance range from zero to 120 km distance. The deduced model is shown in the insert of Fig. 11 together with the observed and synthesized seismograms, which are in fairly good agreement with the experimental data. Close to Cartagena the layer with 6.3 km s^{-1} derived for profile Cartagena–Cádiz (Fig. 3) has been omitted on the profile towards Adra thus leaving a three-layered crust in this area east of the Palomares fault. Finally, P_n travel times have been computed for the two profiles by means of ray tracing using the previously deduced crustal structures above the Moho for the initial part of both profiles. The models shown in Figs 11 and 12 for the area between Adra and Cartagena are consistent with the computed P_n travel times as indicated with dots in Figs 4 and 5 and with the travel-time correlations at shorter distances.

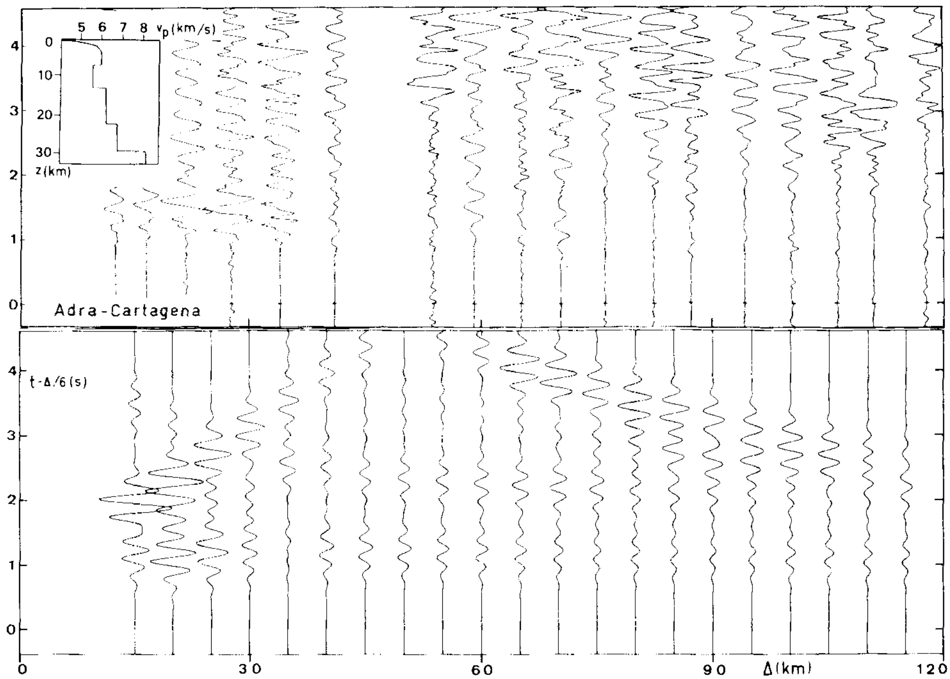

Figure 11. Record section of the profile Adra–Cartagena with expanded time and distance scale (upper part) and velocity–depth model inset from which the synthetic seismograms in the lower part have been computed.

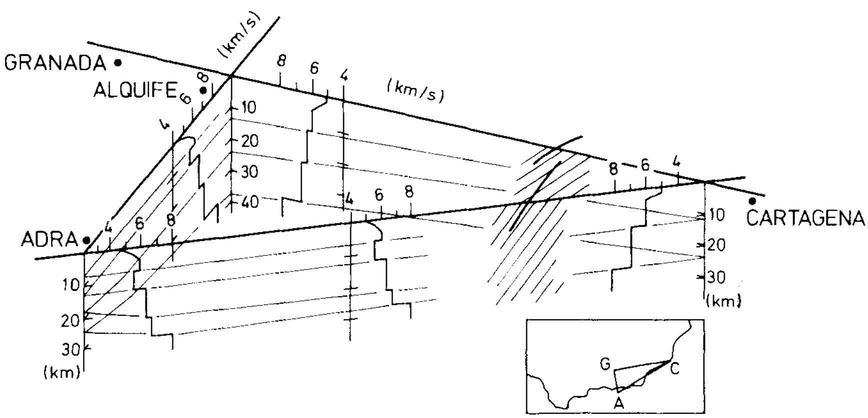

Figure 12. Three-dimensional model of the crustal structure under the eastern part of the Betic Cordillera. See change of crustal structure across the Palomares and Alhama de Murcia fault system (hatching). Faults are indicated by thick lines.

ADRA–UBEDA

Except for the clear P_g phase and a strong P_MP phase the record section for the profile Adra–Ubeda does not show any conspicuous details (Fig. 6). This may be partly caused by the strong lateral variations extending from the south coast through the internal and external zone of the Betic Cordillera which is reached at a distance of 65 km from the shotpoint.

There is no clear indication for phases between P_g and P_MP. Therefore, we assumed a similar type of crust as for the profile Adra–Cartagena. Fig. 6 shows the theoretical travel-time curves for this model. The poorly developed intermediate phases could indicate a major change of crustal structure across the boundary between the internal and external zones. The choice of a true P_n velocity of 8.1 km s^{-1} leads to the same depth of the crust–mantle boundary at the point of intersection with the profile Cartagena–Cádiz east of Granada as can be seen in Fig. 12. Thirty kilometres north of Adra the depth to the Moho increases from 32 to 39 km under the intersection of the profiles.

CRUSTAL MODEL

Fig. 12 summarizes the different crustal models which have been derived for the eastern part of the Betic Cordillera from the presented data and supplementary observations west of Adra by Ansorge & Banda (1980). The strongly varying geologic structure in this area continues at depth with pronounced lateral variations of the crustal structure. The depths of individual interfaces as well as the *P*-wave velocities vary considerably.

The crust east of the Carboneras, Palomares and Alhama de Murcia fault system (Fig. 1) appears to be quite different from the crustal structure west of these faults. In the east the present data indicate a thin three-layered crust with an average crustal velocity of 6.3 km s^{-1} and a thickness of 23 km. The fault system seems to extend to the lower crust and broadens eventually. No detailed crustal information is available for this transition zone as indicated in Fig. 12.

West of the fault system the main part of the profiles lies within the internal Betic zone which is reflected in a more homogeneous type of crust varying only in depth. The data near the south coast lead to an upper-crustal low-velocity zone with a velocity of 5.4 km s^{-1}. Under the profile Cartagena–Cádiz across the centre of the Betic Cordillera the existence of a low-velocity layer in the upper crust could not be verified, because the profile crosses the above-mentioned fault system at the decisive distance range for the identification of such a structural feature. West of the Palomares fault the distance from the shotpoint is too far for the identification of details in the upper crust.

The high-velocity layer in the lower crust (6.6–6.7 km s^{-1}) increases in thickness towards the centre of the Betic Cordillera more rapidly than the upper-crustal layers. The depths to the Moho vary considerably from 23 km around Cartagena to 24 km under Adra and 39 km under the Central Betic Cordillera. The strong gradient of the Bouguer anomaly from south to north near Adra (Fig. 2) indicates a rather steep slope of the Moho close to the coast, which decreases further north towards the centre of the Betic Cordillera, in contrast to the oversimplified plane layering shown in Fig. 12. This has already been shown by Suriñach & Udías (1978) from a three-dimensional interpretation of the Bouguer gravity anomaly in this area.

The mean crustal *P*-wave velocity increases gradually from 6.0 km s^{-1} at the south coast to 6.2 km s^{-1} under the central part of the Betic Cordillera as compared to the relatively high value of 6.3 km s^{-1} east of the Palomares fault system.

THE UPPER MANTLE

With a total length of 440 km the main profile from Cartagena to Cádiz was also aimed at obtaining information about the structure of the uppermost mantle. The distance range from 100 to 350 km which shows the most conspicuous features pertaining to the upper mantle is reproduced in the upper part of Fig. 13 with a reduction velocity of 8 km s^{-1}.

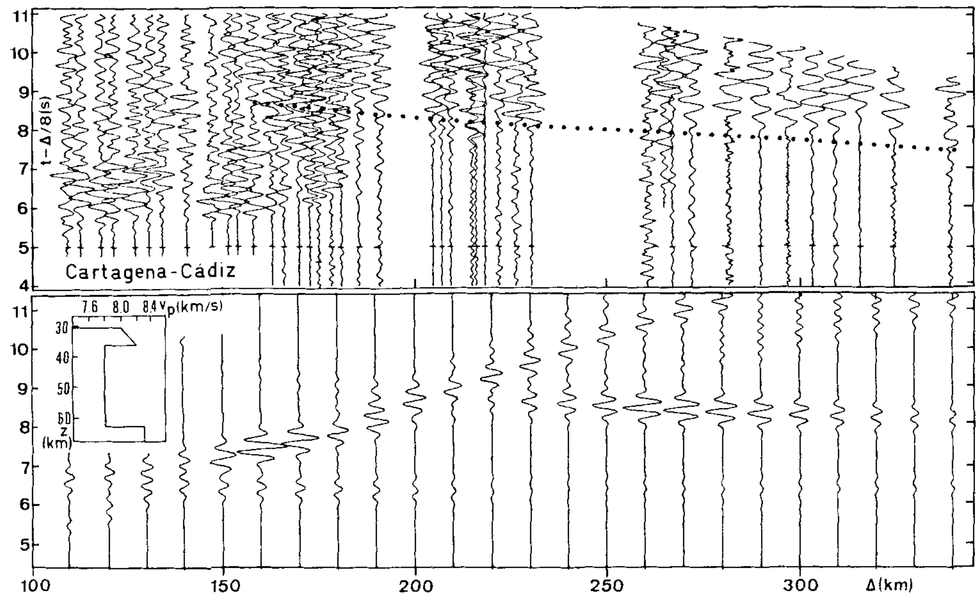

Figure 13. Record section of the profile Cartagena–Cádiz with reduction velocity of 8 km s^{-1} for the phases refracted and reflected from the uppermost mantle (upper part) and *P*-wave velocity–depth distribution in the uppermost mantle (insert) from which the synthetic seismograms in the lower part have been computed.

High-amplitude P_n phases as first arrivals between 125 and 180 km indicate a rather strong velocity gradient beneath the crust–mantle boundary (see also Fig. 3). The low apparent velocity of 7.7 km s^{-1} reflects the increasing depth of this boundary towards the centre of the Betic Cordillera. Phases from the intermediate crustal layers and the over-critically reflected P_MP phase from the Moho (Fig. 3) cover the reduced travel-time interval from 5 to 7.5 s up to a distance of 200 km in Fig. 13. A phase arriving about 1 s later and starting at about 160 km is attributed to an interface in the upper mantle below the Moho. The observed travel times and high amplitudes of this phase require a pronounced low-velocity zone with a strong velocity contrast at its lower boundary. The major features of this structure are shown by the insert in the lower part of Fig. 13 which is characterized by a 6 km thick high-velocity layer at the crust–mantle boundary with a strong gradient from 8.0 to 8.2 km s^{-1}, followed by 26 km with a low velocity of 7.8 km s^{-1}. At a depth of 62 km the velocity increases to 8.3 km s^{-1}. The synthetic seismograms in the lower part of Fig. 13 have been computed for this model of the uppermost mantle assuming a mean crustal model along the profile Cartagena–Cádiz. When comparing the synthetic seismograms with the observations one has to take into account that a horizontally layered model had to be used for these computations which, of course, is an over-simplification of the actual structure. Nevertheless, the main amplitude characteristics are well represented.

For a more realistic model with sloping interfaces and lateral variations of crustal structure travel times have been computed with the ray tracing technique for the reflected arrivals from the lower boundary of the low-velocity zone. The good fit of these computed travel times with the observations can be seen by the dots in the upper part of Fig. 13. Since the crustal structure is not known over the entire length of the profile, we assumed the velocity–depth profile given in Fig. 12 for the eastern part of the long-range profile. According to the Bouguer anomaly (Fig. 2) the Moho was raised to 35 km about 100 km west of Granada. A horizontal layering of the upper mantle structure shows the best agreement with the observations. Lateral variations of the crustal structure along the profile may well change details of the schematic model for the uppermost mantle in Fig. 13, which gives only the main characteristics, i.e. a thin lid with a normal P_n velocity over a thick low-velocity zone of about 26 km and a velocity of only 7.8 km s^{-1}. A pronounced increase in velocity occurs at a depth between 60 and 65 km.

Discussion and conclusions

The eastern part of the Betic Cordillera consists of two differently structured crustal blocks. To the east of the Palomares–Alhama de Murcia fault system (Fig. 1) the continental crust close to the south Balearic basin has a thickness of only 23 km with a high average P-wave velocity of 6.3 km s^{-1}, caused by a thick high-velocity layer of 6.9 km s^{-1} in the lower crust between depths of 12 and 23 km. The structural transition from the continental crust to the oceanic crust of the south Balearic basin is still unknown.

The fault zone which is crossed by the profile Cartagena–Cádiz has been located in detail by Julivert *et al.* (1972), Bousquet & Monténat (1974), Bousquet, Dumas & Monténat (1975), Gauyau *et al.* (1977) and Bousquet (1979). This system of sinistral faults (Carboneras fault, Palomares fault, Alhama de Murcia fault, Fig. 1) extends from east of Almería north-eastwards close to Alicante. The entire block east of the fault system has been moved to the north-east first by extension and subsequent north–south compression before the Tortonian. After a short period of tension a new compressional phase began in the Quaternary (Bousquet & Philip 1976). On the other hand considerable recent microseismic activity has been observed around the Alhama de Murcia fault and a recent earthquake in this area of magnitude 4.2 showed a clear dip-slip focal mechanism (Mezcua 1979). These somewhat contradictory observations together with the thin crust east of the fault system are closely related to the northward moving African plate and the development of the western Mediterranean basins which include areas under compression as well as areas with rifting (Banda *et al.* 1980).

West of the fault system where all the profiles are still lying in the internal zone of the Betic Cordillera a much bigger crustal block of a relatively homogeneous structure seems to exist. A reasonably well-documented low-velocity layer in the upper crust with a velocity of 5.4 km s^{-1} has been derived from the profiles near the southern coast. Whether this feature still exists further north under the central part of the Cordillera cannot be determined from the presently available data.

Clear delays of the P_g phases along the profile Adra–Ubeda to the north of the boundary between the internal and external zones support changes in the structure of the upper crust. This seems to be in agreement with the interpretation of Kampschuur & Rondeel (1975) who identify this boundary as the zone of megashears which were active during the opening of the Atlantic Ocean before the convergence of the African and European plates. Andrieux *et al.* (1971) have identified this boundary as the suture zone between the Iberian plate and their postulated Alboran subplate.

The comparison of deduced P_g-velocity distributions with depth in Fig. 10 shows that a value of 6 km s^{-1} is reached between depths of 3 and 5 km. With regard to the almost constant increase of velocity with depth under the profile Adra–Ubeda we suggest that the thick metamorphic Nevado–Filabride complex (Fontboté 1977) continues to the depth where the low-velocity layer is found (Fig. 12). Mueller (1977) has proposed such a structure in a new model for the continental crust where the upper crystalline basement of metamorphic rocks overlies a zone of granitic intrusions.

In contrast to the preliminary interpretation of some of these data by the Working Group for Deep Seismic Sounding in Spain 1974–1975 (1977), which proposed a high-velocity layer in the lower crust of 7.14 km s^{-1}, the highest velocity in the crust does not exceed 6.7 km s^{-1} in this area. This in turn leads to lower average P-wave velocities of 6.0 km s^{-1} close to the south coast, and 6.2 km s^{-1} at the intersection of the profiles Cartagena–Cádiz and Adra–Ubeda (Figs 1 and 12).

It has to be pointed out that well-developed P_n phases are only observed on those profiles which lie within the internal zone. The generation of the P_n phase depends largely on the velocity gradient below the crust–mantle boundary, which implies from the previous observation that there are still structural differences between the internal and external zones reaching into the uppermost mantle.

The long-range observations along the profile Cartagena–Cádiz lead to a thin layer with a normal P_n velocity of 8.1 km s^{-1} followed by a pronounced zone of lower P-wave velocity of 7.8 km s^{-1} down to a depth of about 63 km. This is in agreement with the structure derived by Payo (1972) and Payo & Ruiz de la Parte (1974) from the dispersion of surface waves for southern Spain. The layer of low P-wave velocity in the upper mantle may continue southward under the Alboran Sea where P_n velocities between 7.5 and 7.9 km s^{-1} have been observed (Working Group for Deep Seismic Sounding in the Alboran Sea 1974, 1978; Hatzfeld 1978). This means that the already thin layer with a P-wave velocity of 8.1 km s^{-1} has disappeared completely combined with a continuing rise of the crust–mantle boundary southward to a depth of 17 km under the Alboran Sea. Nothing can be said about the orientation of a possible subduction zone, but the unusual low P-wave velocity in the uppermost mantle and the rather diffuse seismicity east of the Strait of Gibraltar (Udías *et al.* 1976) exclude a simple subduction of the African plate dipping under the Iberian Peninsula. The main findings of this study are: (1) The existence of blocks with clearly different crustal structure within the Betic Cordillera, (2) the strong variation of crustal thickness, (3) deep-reaching structural differences between the internal and external zones, and (4) the pronounced decrease of velocity in the uppermost mantle. These results in addition to earlier investigations in the Alboran Sea and northern Africa have to be incorporated in any geodynamic model for the evolution of the arc of Gibraltar.

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