3-D crustal structure of the extensional Granada Basin in the convergent boundary between the Eurasian and African plates

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Received 27 March 2001; accepted 10 September 2001

Abstract

Three-dimensional P and S wave velocity models of the crust under the Granada Basin in Southern Spain are obtained with a spatial resolution of 5 km in the horizontal direction and 2 to 4 km in depth. We used a total of 15407 P and 13704 S wave high-quality arrival times from 2889 local earthquakes recorded by both permanent seismic networks and portable stations deployed in the area. The computed P and S wave velocities were used to obtain three-dimensional distributions of Poisson’s ratio ($\nu$) and the porosity parameter ($V_p/C^2V_s$). The 3-D velocity images show strong lateral heterogeneities in the region. Significant velocity variations up to $\pm$7\% in P and S velocities are revealed in the crust below the Granada Basin. At shallow depth, high-velocity anomalies are generally associated with Mesozoic basement, while the low-velocity anomalies are related to the neogene sedimentary rocks. The south–southeastern part of the Granada Basin exhibits high $\nu$ values in the shallowest layers, which may be associated with saturated and unconsolidated sediments. In the same area, $V_p/V_s$ is high outside the basin, indicating low porosity of the mesozoic basement. A low-velocity zone at 18-km depth is found and interpreted as a weak–ductile crust transition that is related to the cut-off depth of the seismic activity. In the lower crust, at 34-km depth, a clear slow $V_p$ and $V_s$ anomalous zone may indicate variations in lithology and/or with the rigidity of the lower crust rocks. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: Crustal earthquakes; Low-velocity layer; P waves; S waves; Seismic tomography; Granada Basin

1. Introduction

The western Mediterranean experiences seriously associated with the interaction between the Eurasian and African plates. The spatial distribution of the earthquakes is in agreement with a well-defined plate boundary on the Atlantic side (Menard, 1965), where in its eastern segments, an ocean–ocean convergence between the two plates is taking place (Minster and Jordan, 1978; Argus et al., 1989, Sartori et al., 1994); earthquakes are scattered and focal mechanisms show a combination of thrust faulting and strike–slip motion (McKenzie, 1972; Fukao, 1973; Udías et al., 1976; Grimison and Chen, 1986). The plate boundary is less clear between the Iberian Peninsula and Morocco-Algeria, its seismicity is diffuse and the zone delimited by the seismicity reaches a maximum...
width of 300 km (Anderson and Jackson, 1987, Buforn et al., 1988). This region is characterized by a high rate of seismicity with many earthquakes of magnitude ≤ 5.0 (Fig. 1).

The Betic Cordillera in Southern Spain and the Rif in Morocco (Fig. 2) are among the most important Alpine orogene of the western Mediterranean, which were built up as the responses to the convergence between the Eurasian and African plates (Dewey, 1988). The Granada Basin is one of the most distinct intramontaneous basins of the Betic Cordillera, which is filled with sedimentary rocks lying over the External (EZ) and Internal Zones (IZ), which constitute the basement of Granada Basin (Fig. 2). The first features of the formation of the Granada Basin started in the Middle Miocene whose rocks crop out in the SE area; nevertheless, it is from the Late Miocene (Early Tortonian) to the present that it acquired its principal characteristics. The Tortonian rocks are marine and their outcrops reach up to 1500 m above sea level (Rodriguez-Fernandez, 1989), while the Turolian and subsequent rocks are continental. The sedimentary infill, like the metamorphic basement, is affected by upright open folds trending NE–SW and by normal-fault set with NW–SE strikes (Galindo-Zaldívar et al., 1999). The External Zones, which formed most of the southern and part of the eastern margin of the Iberian Massif, are mainly composed by limestones, dolostones, marls, sandstones and clays of Mesozoic and Cenozoic ages. Moreover, igneous rocks (ophiolites) and pillow-lava levels are intercalated into the Triassic rocks and the Jurassic and Cretaceous sedimentary series. The Internal Zones, situated farther to the east, consist of several tectonic units mostly composed by schists and marbles. The types of deformation occurring in the Internal and External Zones and their history are different, and it is only the great displacement (primarily during the Early Miocene) that created the present-day juxtaposition (Sanz de Galdeano, 1990).

The main aim of this study is to analyze in detail the features of the crust below the Granada Basin and surrounding regions by using high-resolution seismic tomography. The availability of a dense seismic net-

Fig. 1. Seismicity of the South Spain, Morocco and West Argelia for the period 1983–2001 (magnitude ≥ 2.8) registered by Andalusian Seismic Network (Andalusian Institute of Geophysics, Granada University, Spain).
Fig. 2. Simplified geological map of Central Betic Cordilleras. Crosses, triangles, squares and dark circles denote the earthquakes, permanent seismic stations belongs to RSA, temporary seismic stations belongs to RSA and seismic stations belongs to National Geographic Institute, Ministry of Fomento and Royal Institute and Navy's Observatory at San Fernando, respectively. The insert map on the upper right side shows the location of the present study area.
work has allowed us to collect high-quality data from many earthquakes of small magnitudes in the area. The Granada Basin and the neighboring areas have the highest rate of microseismicity ($m_b \leq 5.5$) in the Iberian Peninsula (De Miguel et al., 1989). The seismic activity, on average, does not exceed 20 km in depth; the lower crust is essentially aseismic and the seismic activity is concentrated in the upper crust between 5 and 17 km. This lower cut-off in the seismic activity in the crust is interpreted as the limit brittle–ductile, limiting the thickness of the seismogeneic layer (Morales et al., 1997). These authors pointed out the coincidence of the lower cut-off in the seismic activity with the presence of an intracrust reflector detected by Banda et al. (1993) and Galindo-Zaldivar et al. (1997) and which would be layering the crust in two: a seismic part (brittle) and another aseismic (ductile), according to the differences observed in the style of the deformation between the upper and lower crust (Galindo-Zaldivar et al., 1997).

Fig. 3. (A and B) Trade-off curves between the variance of the solutions and arrival time residuals for inversions with irregular discontinuities (3D) and flat discontinuities (1D). The selected values (damping) for the final results are marked within the circle. (A) For P wave velocity. (B) For S wave velocity.
The present state of stress determined by focal mechanism solutions shows an extensional regime. Both the surface geological data and the focal mechanisms indicate a present-day regional NE–SW extension, with triaxial to prolate stress ellipsoids. However, the stress field is heterogeneous with both spatial and temporal variations, sometimes even acting simultaneously in adjacent areas. The most frequent changes consist of radial or NW–SE extension favored by the low axial ratio of the stress ellipsoids and NW–SE subhorizontal compression favored by the regional tectonic (Galindo-Zaldívar et al., 1999). The extension is normal to the regional convergence direction (NW–SE) (DeMets et al., 1990). Beneath this region, there are also deep earthquakes with focal depths down to 640 km although there is a lack of intermediate earthquakes (Buforn et al., 1997).

Numerous seismic tomographic studies have already been performed in this region: Blanco and Spakman (1993), Plomerová et al. (1993), Sallarés (1996), Gurría et al. (1997), Serrano et al. (1998) and Calvert et al. (2000) have found important velocity anomalies in this

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**P-wave tomography**

Fig. 4. Fractional P wave velocity perturbations (in percentage) at three depth layers for the case when the Moho and the mid-crust discontinuity are flat.
Fig. 5. (a) Fractional P wave velocity perturbations (in percentage) at the first six depth layers for the case when the Moho and the mid-crust discontinuity have lateral depth variations. The velocity perturbation is from the mean value of the inverted velocity at each layer. The depth of the layer is shown at the lower-left corner of each map. Blue and red colors denote fast and slow velocities, respectively. The velocity perturbation scale is shown on the right. Black lines show active faults. Red lines show the boundary between the Granada Basin (sedimentary rocks) and External–Internal Zones. (b) The same to the second six depth layers.
P-wave tomography

DEPTH: 18 KM

DEPTH: 22 KM

DEPTH: 26 KM

DEPTH: 30 KM

DEPTH: 34 KM

DEPTH: 38 KM

Fig. 5 (continued).
region. However, the low resolution of these results do not allow a detailed correlation between the tomographic results and shallow geological structures.

In local earthquake tomography, body-wave arrival times are used to estimate P and S wave velocities and from these, we deduce variations in lithology and physical properties of rocks. According to Christensen (1989), velocity usually increases with depth; however, in some regions, velocity reversals have been recognized. Comparisons of laboratory velocity data with seismically measured velocities are subject to many considerations because the influences of temperatures and pressures must be taken into account to infer mineralogy from seismic velocities. For crystalline rocks, the initial application of pressure affects the seismic velocity by reducing microporosity and within the upper part of the crust, large-scale fracturing may have an analogous effect. In deeper portions of the crust and in the upper mantle, however, it is likely that fractures and microcracks are no longer present because of the high confining pressures and metamorphic recrystallization. In sedimentary rocks, especially sandstone, porosity is much higher and does not completely close at elevated pressures and so will have a dominant influence on seismic velocities. For most common rock types, velocity decreases with increasing temperature. Velocities have been found to correlate well with mineralogy at pressure high enough to eliminate the influence of cracks. The lower density minerals, such as quartz, have lower velocities than the high-density pyroxenes, olivine and garnet. In general, metamorphic rocks of mineral composition similar to igneous rocks have identical velocities although this may be complicated by anisotropy (Christensen, 1989). A brief review of how rock properties relate to seismic velocity and attenuation can be found in Sander et al. (1995) and Lees and Wu (2000). Poisson’s ratio is directly related to \( V_p/V_s \), the ratio of compressional and shear-wave velocities and can be a useful indicator of lithology and pore fluid pressure. On average, Poisson’s ratio is 0.25 for Earth’s crust and upper mantle (Holbrook et al., 1988). Any variation from this value may indicate a change in the properties of the material. On the other hand, the product of compressional and shear-wave velocities, \( V_p \times V_s \) has been used to delineate porosity in sedimentary rocks (Iverson et al., 1989). It has been observed that lower \( V_p \times V_s \) indicates an increase of porosity, whereas \( V_p/V_s \), constant for specific lithology, does not change with porosity (Pickett, 1963; Tatham, 1992). \( V_p/V_s \) or \( \sigma \), is commonly used to delineate lithology, while the product \( V_p \times V_s \) can be used to identify variations in porosity for the shallow crustal rocks (Iverson et al., 1989; Tatham, 1992). In this work, we try to relate the three-dimensional variations in \( \sigma \) and \( V_p \times V_s \) to fluctuations in fluid content, porosity and changes in the lithologies of different zones of the Granada Basin.

2. Data selection

We used arrival time of P and S waves from local earthquakes recorded by the permanent and portable seismic stations of the Andalusian Seismic Network (RSA) that is operated by the Andalusian Institute of Geophysics (Granada University). Although the station coverage of this network is dense and adequate for locating local earthquakes, we have also used the data from other stations that belong to other institutions (i.e., National Geographic Institute, Ministry of Fomento and Royal Institute and Navy’s Observatory at San Fernando) in order to achieve a better ray coverage for the tomographic inversion. The data used were gathered from the events that occurred between 1983 and 1999. Since 1983 to 1988, the most of the data were recorded in analog formats, from 1988 onwards all the recordings are digital (except three stations belong to Ministry of Fomento). The seismic stations used are densely and uniformly distributed in the studied area (Fig. 2).

We selected a set of events which are located between 36°48’N to 37°12’N and from 3°24’W to 4°03’W. These earthquakes were selected on the basis of minimum number of arrival times, at least 10, as well as the fact that they have a uniform spatial distribution in the study area. The 96% of the earthquakes selected have rms smaller than 0.2 s and the rms maximum for the hypocenter locations is 0.9 s. All the events are located by more than five stations. Finally, we selected a total of 2889 earthquakes, with hypocenters located using the method of Lienert et al. (1986). The mean error is 2 km for the latitude, 0.8 km for the longitude and 1 km for the depth provided by the location program. The depth of the selected earthquakes ranges from 0 to 40 km. The majority of
the events have focal depths shallower than 20 km below the Granada Basin; only the events located beneath the southwestern part of the basin are deeper. Fig. 2 shows the epicenter distribution of the selected earthquakes. A total of 15407 P arrivals and 13704 S arrivals were collected from these earthquakes in the study area of 97 × 52 km. For stations of the Andalusian Seismic Network, the accuracy of time picking of the P and S digital arrivals may be estimated in the most favorable cases as ±0.01 s. In the case of less impulsive arrivals and/or poor signal-to-noise ratio, the accuracy is degraded, but not more than 0.1 s. For the analog recordings (less than 4% of the total number of data used in this study), the mechanical characteristics of the drums and the velocity of development of the traces (2 mm/s) set a limit to the accuracy of about ±0.2 s. With regard to the picking up of the S arrivals, all seismic stations, except one, have vertical short period sensors.

3. Method and analysis

In this study, we have used the tomography method of Zhao et al. (1992). Although the conceptual approach of this method is derived from Aki and Lee
(1976), it has some additional features. The technique can deal with the complex geometry of seismic velocity discontinuities, such as the Moho or the subducting slab boundary, and it uses a 3-D ray tracing scheme to compute travel times and ray paths. For details of the method, see Zhao et al. (1992, 1994).

Our velocity model contains three layers: the upper crust, the lower crust and the upper mantle, which are bounded by the Moho and the mid-crust discontinuity. We set three-dimensional grid nets independently for every layer to express the three-dimensional velocity structure for layers, which are bounded by two adjacent discontinuities. Velocities at grid points are taken to be unknown parameters except those at the outermost grid. Velocities at the outermost grid are used just to interpolate velocities outside of the modeling space. A velocity at any point in the model is calculated by linearly interpolating the velocities at the grids surrounding that point (Zhao et al., 1992).

The P wave velocity ($V_p$) for the upper crust, lower crust and the uppermost mantle is 6.0, 6.8 and 8.0 km/s, respectively. $V_p$ in the upper mantle has a vertical gradient of 0.003 km/s/km. $V_p/V_s$ is set to be 1.7 in the initial model. We have constructed this initial velocity model for the shallow layers taking into consideration the results of Wadati Diagrams of 335 earthquakes occurred in the region (Morales et al., 1997) and

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**Checkerboard resolution test ($V_p$)**

![Checkerboard resolution test](image)

Fig. 7. Results of checkerboard resolution test for P (a) and S (b) waves. Black and white symbols denote slow and fast velocities, respectively. The depth of each layer is shown at the upper part of the map. The grid spacing is 5 km in the horizontal direction and 2–4 km in depth. The perturbation scale is shown on the right.
considering for all the layers the seismic profiles and gravimetric data from Galindo-Zaldívar et al. (1997, 1998) and the results for the Betic Cordillera from Banda et al. (1993).

In a preliminary inversion, we assumed that the mid-crust and the Moho discontinuities had constant depths at 17 and 38 km, respectively. Then, we tried to use a more realistic model with lateral depth changes in the Moho and mid-crust discontinuity. The Moho and mid-crust discontinuity geometries were constructed by referring to the seismic explosion profiles and gravimetric data (Galindo-Zaldívar et al., 1997, 1998). The mid-crust discontinuity depth ranges from 17 to 15 km; the Moho depth ranges from 38 to 36 km. After we compared the results for both of the cases, the most meaningful outcome is that the inversion with the curved discontinuities results in a final root-mean-square residual of 18% for $V_p$ and 12% for $V_s$ smaller than those with flat discontinuities (Fig. 3). This suggests that it is important to take into account the lateral depth variations of the seismic discontinuities in order to determine a detailed crustal structure. When the geometry of the discontinuities is taken into account, ray paths and travel times can be calculated more accurately (Zhao et al., 1992, 1994). Also, we found that the general patterns of velocity distributions in both cases are almost the same although there are some differences in amplitudes of the velocity anomalies for the upper and lower crust (Figs. 4 and 5).

We solved the inverse problem for 3996 (2015 for P and 1981 for S waves) velocity parameters at the grid nodes with hit counts (number of rays sampling a...
cell) greater than 10. Fig. 6 shows the distribution of ray paths at several layers. The errors for the velocity perturbations obtained are estimated to be < 1%. The damping parameter was selected based on an empirical approach (Eberhart-Phillips, 1986). A number of inversions were run with different damping values. Then the reduction in travel time residual is compared to the variance of the solutions and we draw a trade-off curve between them. The selected value of the damping parameter is the one which gives the optimal residual reduction and the solution variance (Fig. 3).

4. Hit counts and resolution test

Before describing the features of the obtained models, we first show the distribution of ray paths in the study area and evaluate the resolution of the tomographic image. Fig. 6 shows the distribution of hit counts at several layers. The density of ray path coverage varies throughout the study area. The coverage is very good down to a depth of 18 km in almost the whole area, with the highest density in the central part of the Granada Basin. From 22 to 38 km depth, the coverage is good in the western part of the study area and it is not good in the eastern part. Hence, reliable results are expected for the upper crust layers of the study area (from 2 to 14 km).

By the concept of resolution, we wish to know how the true structure is reconstructed in the calculated image. The most direct means of testing the resolution of the inverted solution is to first calculate the sets of travel time delays that resulted from tracing the actual ray set through a synthetic test structure, then invert

![S-wave tomography](image)

Fig. 8. The same as Fig. 5 but for S wave velocity structure. (a) Fractional S wave velocity perturbations (in percentage) at the first six depth layers for the case when the Moho and the mid-crust discontinuity have lateral depth variations. (b) The same to the second six depth layers.
those delays as though they are data and, finally, compare the synthetic inversion with the initial structure (Zhao et al., 1992). We used the checkerboard resolution test (see Humphreys and Clayton, 1988). Positive and negative perturbations are assigned to a 3-D grid nodes, which are arranged in the modeling space, the image of which is straightforward and easy to remember. We conducted a number of inversions by changing the grid spacing between grid nodes, which are set up in the study area. Through resolution analyses for the different grid spacing, we found that a grid spacing of 5 km in the horizontal direction and 2–4 km in depth is the minimum space for that the data give reasonable results (Fig. 7). The results of the checkerboard test are good down to a depth of 18 km; however, the results for depths from 22 to 38 km under eastern part of the basin are not very good. As expected, the resolution is good in areas where most of the earthquake occurred. The total number of grid nodes is $13 \times 23 \times 14$.

5. Results and discussion

The velocity at a grid point represents the best estimate for the volume surrounding the grid point. Similarly, the size and shape of velocity anomalies can provide reasonable estimates, but may not correspond to the exact boundaries of true velocity features (Eberhart-Phillips, 1986). Considering the good resolution in the shallow layers, we will mainly focus our discussion on features obtained in those layers. There is no volcanism and geothermal activity in the Granada Basin and surrounding areas, then anomalies in $V_p$ and $V_s$ may be related to changes in lithology, cracks and fluid distribution in the crust.
5.1. Layer: 2 km

One of the distinct features of this layer is the high P and S wave velocity anomaly under the eastern boundary of the Granada Basin (Figs. 5 and 8). This anomaly coincides with the outcrop of dolomies and marbles belonging to the Mesozoic basement (IZ). The anomaly also exhibits high \( V_p \times V_s \) anomaly in a small area (+15%), which may indicate low porosity, compacted and weakly fractured material here (Fig. 9).

In the south-central and south-southeast parts of Granada Basin, there is a high \( V_p \) zone that goes through neogene–quaternary sedimentary rocks (inside the basin) and marble–dolomies of the mesozoic basement (outside the basin, IZ). Its trend may be parallel to the local normal-fault set with NW–SE strikes. In this same zone, the results for \( V_s \) are different: inside the basin (sedimentary layers) shows low \( V_s \) (−6%) and outside the basin shows high \( V_s \) (+4%), which may correspond to low-crack density (mesozoic basement). The value of \( \sigma \) is high inside the basin (Fig. 10), which may be related to saturated and unconsolidated sediments. \( V_p \times V_s \) is high outside basin, indicating low porosity of the mesozoic basement.

The W Granada Basin (EZ) exhibits high \( V_p \) (+6%) and \( V_s \) (+2%), generally consistent with the magnetic data (Ardizone et al., 1989). The positive magnetic anomaly and fast \( V_p \) anomaly can indicate the existence of a body formed by basic igneous rocks or metabasites, which belong to the Iberian Massif. The Poisson ratio shows higher values than those obtained at S–SE and E Granada Basin, indicating different lithologies on the two sides of the basin. However, the \( V_p \times V_s \) values are similar, indicating a low fracture density or highly compacted material on the two sides of the outside basin.

A low \( V_p \) anomaly is observed in the central and north Granada Basin, which is correlated with the neogene and quaternary rocks. \( V_s \) shows, in general, low values except in a small area, where we can see a high-velocity anomaly (Fig. 8). Also, in this small area, \( \sigma \) is very low (−25%), indicating rocks with a higher rigidity than the surroundings. This high \( V_s \) and low \( \sigma \) in this small area may be related to a thin sedimentary layer inside the basin.

Under the southwestern part of the Granada Basin, there are low \( V_p \) and \( V_s \) anomalies and we obtain the smallest values of the porosity parameter (−14.8%) of the whole study area, indicating a high-porosity
material. This velocity anomaly shows a general N–S trend. This anomaly (about 150 km²) corresponds to the southern part of a great extension of Jurassic carbonated rocks (Sierra Gorda) and alluvial materials belonging to “Polje de Zafarraya”, formed by sand levels. This aquifer is one of the most important features in this region (MAGNA, 1979).

A P and S low-velocity zone is imaged outside the NW Granada Basin and the porosity parameter is very low. The major part of this anomaly corresponds to one of the thick continental areas filled of the Granada Basin (MAGNA, 1988). It is formed by marls, silts, limestones, conglomerates, sands and silt clays.

5.2. Layers from 4 to 14 km

The eastern part of the Granada Basin continues showing high $V_p$ (+5%) and high $V_s$ (+5%) at 4-km depth, which disappear at 8-km depth. The porosity parameter also shows high values (+11%) down to a depth of 6 km, indicating low porosity in this area.

At a depth of 4 km, the high $V_p$ and $V_s$ anomaly in the southeastern part of the basin decreases to smaller values. This anomaly again appears at a depth of 10 km, showing high values of $V_p$ (+5%) and $V_s$ (+5%). Under the southeastern part of the basin, Poisson ratio is high (+22%) down to a depth of 4 km, indicating low rigidity.

The western zone of the Granada Basin exhibits high values of $V_p$ and $V_s$ down to a depth of 10 km, while the porosity parameter is high down to a depth of 4 km. From a depth of 6 to 10 km, the area west of the basin shows high $V_p$ (+5%) and lower values of $V_s$.

It is important to point out that from 6 to 10 km depth outside the N–NW Granada Basin, the $V_p$ model shows a high-velocity body with a NS trend. At the

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**Fig. 10.** Percent perturbation of Poisson’s ratio at (a) the first six depth layers and (b) the second six layers.
surface, this area shows outcrops of igneous rocks (Subbetic units), such as ophiolites, which are normally found located together with the Triassic rocks of keuper facies and pillow-lava levels, intercalated into the Jurassic and Cretaceous sedimentary series. One of the most intense positive aeromagnetic anomalies of the Betic Cordillera is located in this area (Ardizone et al., 1989). The maximum in the total field surface anomaly map reaches up to 120 nT and its place is coincident with the positive $V_p$ anomaly obtained in this study. According to Bohoyo et al. (2000), this large extent of the magnetic anomaly maximum is probably the consequence of a large body of basic igneous rocks. They investigated this area with new magnetic and gravity data and their results indicate that the high aeromagnetic anomaly values and the small difference between the land and aeromagnetic anomaly values indicate that it has a deep origin.

Under the central Granada Basin, we can observe a NE–SW preferential orientation of the boundary between high and low velocity, the southeastern part of the basin shows higher velocity, in general, than that of the northwest part. This boundary can be in relation to the NE–SW-trending contact between the EZ and IZ.

5.3. Layer: 18 km

One interesting feature is the presence of a low-velocity zone at mid-crustal depths beneath the Granada Basin. This slow $V_p$ and $V_s$ appear at a depth of 18 km and strong changes of seismic velocity are visible from a depth of 14 to 18 km (Fig. 11). The low-velocity zone is in agreement with the depth range of crustal seismicity (Morales et al., 1997), which might imply that it is a crustal decoupling zone at the base of the
seismogenetic zone. A strong high $V_p$ and $V_s$ anomaly is also detected in the southeastern boundary of the basin. This anomaly (Fig. 10) coincides with the earthquake swarm activity located in the zone (Saccorotti et al., in press) and could be associated with an asperity.

These results together with the highest values of $\sigma$ in two small areas, NW Loja and N Sierra Alhama, may be indicating the weak sections of the seismogenic crust due to overpressurized, fluid-filled, fractured rock matrices, similar to those detected in the source area of the 1995 Kobe earthquake in Japan (Zhao et al., 1996).

5.4. Layers from 22 to 38 km

The low $V_p$ and $V_s$ values continue under the Granada Basin down to a depth of 26 km. Very slow $V_p (-5\%)$, slow $V_s (-3\%)$ and low $\sigma$ are obtained in the NW of the study area at a depth of 34 km. The low resolution in the lower crust is considered to be due to the lack of earthquakes in these layers and the resolution for S wave velocity structure is inferior to that for P wave. For this reason, we prefer not to do interpretation of the results although these values may indicate variations of the lithology or a high rigidity of the crustal material as compared with the surrounding areas.

6. Conclusions

High-resolution P and S wave tomographic images are determined for the Granada Basin. The results show strong lateral heterogeneities in the crust.
In the depth range 2 to 4 km, the SE part of the Granada Basin shows a strong high $V_p$ zone (+6%). Its trend is parallel to the local normal-fault set with NW–SE strikes. We obtained low $V_s$ (−6%) inside the basin (sedimentary layers) and high $V_s$ (+4%) outside the basin (IZ, mesozoic basement), which may correspond to the difference in crack density. The value of $\sigma$ is high inside the basin, which may show saturated and unconsolidated sediments. $V_p \times V_s$ is high outside the basin, indicating a low porosity for the mesozoic basement. The eastern part of the Granada Basin shows high $V_p$ (+6%) and $V_s$ (+7%), low $\sigma$, in accordance with the presence of Mesozoic basement (IZ). A strong high anomaly of $V_p \times V_s$ in a small area (+15%) indicates a material of very low porosity, compacted and weakly cracked. The western part of the Granada Basin (Sierra Gorda, EZ) shows high $V_p$ (+6%) and $V_s$ (+2%), generally consistent with magnetic data. The positive magnetic anomaly and fast $V_p$ anomaly can indicate the existence of a body probably comprising basic igneous rocks or metabasites, which belong to the Iberian Massif. $\sigma$ shows higher values than those obtained in the southeast and eastern, indicating different lithologies in the two sides of the basin. However, the $V_p \times V_s$ values are similar, indicating a low fracture density or highly compacted material.

In the depth range of 6 to 14 km, the southeastern part of the basin shows high $V_p$ (+5%) and high $V_s$ (+5%). In the depth range of 18 to 38 km, slow $V_p$ and $V_s$ appear at a depth of 18 km under the south part of the basin. The maximum depth of seismicity is 14–17-km depth. This low-velocity layer may be related to a highly fractured and fluid-filled zone. $\sigma$ is highest in two small areas: NW Loja and N Sierra Alhama. It may indicate the weak sections of the seismogenic crust due to overpressurized, fluid-filled, fractured rock matrices.

Acknowledgements

This work has been supported by the Comision Interministerial de Ciencia y Tecnologia project AMB99-0795-C02-01 and REN2001-2418-C04-04 (Spain). The first author (I. Serrano) thanks the University of Granada (Spain) and Ministerio de Ciencia y Tecnologia for two postdoctoral fellowships at Ehime University (Japan). The authors would also like to thank Claudio Chiarabba and an anonymous reviewer for their very constructive reviews and suggestions.

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