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Estudio Sísmico de la Corteza Ibérica Norte 3.3: A seismic image of the Variscan crust in the hinterland of the NW Iberian Massif

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Abstract. An offshore vertical incidence reflection seismic study with simultaneous on-land wide-angle recording has been conducted, as part of the Estudio Sísmico de la Corteza Ibérica Norte (ESCIN) Project, in the NW Iberian Variscan Belt, an area deeply affected by Variscan and Alpine deformation episodes. Line ESCIN-3.3 was recorded across strike of the main Variscan structures, transecting two of the zones in which the NW Iberian Massif has been subdivided. This study intends to correlate the seismic features observed in the near-vertical reflection profile with the velocity model deduced from the wide-angle data, with the well-known geological structure of the Variscan upper crust, and with some other geophysical features described on the basis of offshore and onshore experiments carried out in the area. The main results can be summarized as follows: (1) identification of two important sedimentary basins in the shallow part of the marine seismic section, related with the Mesozoic extension that gave rise to the opening of the Bay of Biscay, (2) correlation of a number of subhorizontal and west dipping intermediate depth reflections with Variscan compressional structures, (3) interpretation of a local set of strong reflections at 6 s (two-way travel time) as lower crust emplaced at an anomalous shallow level (thought to image the northward prolongation of a body with high *P* wave velocity and high magnetic susceptibility, described on land underneath a late Variscan antiformal structure, the Lugo Dome), and (4) recognition of two controversial highly reflective subhorizontal bands, located at 7-9 and 11-12 s, respectively. The shallowest of them appears to have a lower crustal *P* wave velocity and to be separated from the deepest one, probably with similar characteristics, by mantle *P* wave velocity material. On this basis, a duplication of the lower continental crust is inferred. The deepest of the bands could either represent the remnants of a late Paleozoic crustal root or an Alpine underthrusting of the lower continental crust.

1. Introduction

The Estudio Sísmico de la Corteza Ibérica Norte (ESCIN) Project is a multichannel onshore/offshore seismic experiment carried out in north and NW Spain (Figure 1) aiming to improve the understanding of the upper Paleozoic history of this part of the Variscan Belt and its Mesozoic and Tertiary evolution.

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Detailed description of the project and first results are given by Pérez-Estaún et al. [1994, 1995], Pulgar et al. [1995, 1996], Martínez Catalán et al. [1995], and Alvarez-Marrón et al. [1996].

The Variscan Orogeny contributed in the late Paleozoic to the assembly of all the continental masses in the Pangea The NW Iberian part of this mountain belt has been reworked afterward as a consequence of the Mesozoic extension that led to the opening of the Bay of Biscay [*Lamboy and Dupeuble*, 1975; *Boillot et al.*, 1984] and the Cenozoic compression that triggered the Alpine orogeny. Evidences of this three major tectonic episodes are frequent everywhere from the Duero Basin to the continental shelf [*Boillot et al.*, 1984; *Alvarez-Marrón et al.*, 1996, 1997; *Pulgar et al.*, 1995, 1996; *Fernández Viejo*, 1997]. The profile described here, shot across strike of the Variscan structures, was meant to investigate mainly late Paleozoic features and designed as a marine line on the basis that such features, mapped on land, are assumed to continue toward the north in the continental shelf with the same direction.

The crust of the European Variscides and its postorogenic evolution is well known thanks to a number of seismic studies [e.g., *Meissner and DEKORP Research Group*, 1991; *Bois and ECORS Scientific Party*, 1991]. The lower crust and the Moho appear usually as flat features beneath the Variscan Belt, whereas the Tertiary belts may show crust-mantle imbrications often described as crustal roots. Since these orogens are the result of similar collisional processes, deep crustal roots must have existed at the end of the Variscan Orogeny. They must have been removed as a consequence of the crust/mantle transition reequilibration. Metamorphism and/or magmatism must have played a major role in the removal of these crustal roots [*Bois and ECORS Scientific Party*, 1991]. However, the conditions (mostly depth and temperature) necessary for the reequilibration of the crust might not take place in all the orogens (e.g., the Urals) or not in a global scale in the same orogen (e.g., reequilibrated Gulf of Lions versus intact Pyrenees). For that reason, deep Paleozoic, Proterozoic, and even Archean features are still preserved in some orogens [*BABEL Working Group*, 1991; *Knapp et al.*, 1996; *Carbonell et al.*, 1996; *Clowes et al.*, 1996].

This paper will describe the seismic features of the line ESCIN-3.3 and present an interpretation based on seismic data, surface geology, and the knowledge of the Variscan evolution of the NW Iberian Massif. Special attention will be paid to the controversial age of a likely lower crust duplication 100-130 km long.

2. Geological Setting and Geophysical Framework of NW Iberia

The NW Iberian Massif is composed of four zones that from east to west are the Cantabrian Zone (CZ), the West Asturian

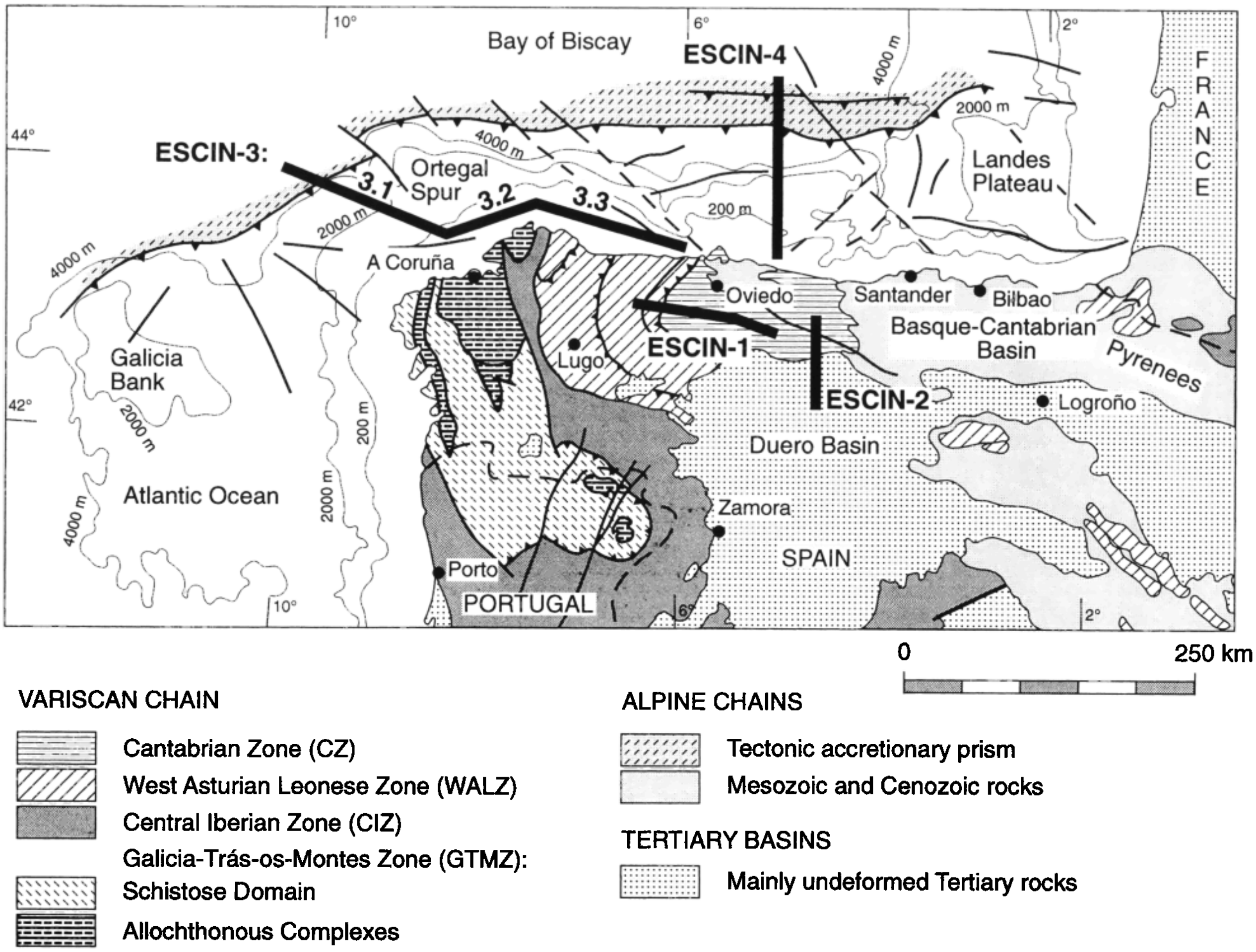


Figure 1. Location of the Estudio Sísmico de la Corteza Ibérica Norte (ESCIN) seismic profiles 1, 2, 3, and 4 in relation to the most important geologic features of NW Iberian Peninsula. Zones in the Variscan Belt are after *Julivert et al.* [1972] and *Farias et al.* [1987]. Offshore geology is after *Boillot and Malod* [1988]. Depths are in meters.

Leonese Zone (WALZ), the Central Iberian Zone (CIZ), and the Galicia-Trás-os-Montes Zone (GTMZ) (see geological map and cross section of WALZ, CIZ, and GTMZ in Figure 2). Two of them, WALZ and CIZ, have been sampled by the line ESCIN-3.3.

The CZ represents the foreland thrust and fold belt of the Variscan Belt and is characterized by a well-defined, thin-skinned tectonics [*Pérez-Estaún et al.*, 1994]. Regional metamorphism and internal deformation are very weak to nonexistent. The deformation is characterized by east directed thrusts and associated folds.

In the WALZ, a hinterland zone immediately west of the CZ, the metamorphism and internal deformation become generalized and increase in grade and intensity toward the west. The structural style is characterized by east vergent overturned and recumbent folds cut and displaced by east directed thrusts. The most important geologic structures are the Mondoñedo Nappe [*Martínez Catalán*, 1985; *Bastida et al.*, 1986] and the Viveiro Fault, an extensional detachment which defines the limit between this zone and the CIZ, adjacent to the west [*Martínez Catalán et al.*, 1992]. The Mondoñedo Nappe occupies the western half of the WALZ and is characterized by a set of large recumbent folds bounded by a basal thrust. The folds, the basal thrust, and an associated ductile shear zone are folded by a pair of open, late subvertical folds. The western fold is an antiform known as the Lugo Dome, which is followed to the east by a wide synform. The autochthon of the nappe outcrops in the core of the antiform, forming the Xistral Tectonic Window (Figure 2).

The CIZ is represented in NW Spain by the narrow "Ollo de Sapo" Anticlinorium [*Parga Pondal et al.*, 1964]. This structure results from the interference between recumbent and late subvertical folds. Variscan deformation, metamorphism, and magmatism are comparable to those in the WALZ.

The GTMZ, the most internal of the zones in NW Spain, is characterized by a large stack of allochthonous units thrust over the CIZ and giving rise to several allochthonous complexes (Figure 1). These units include ophiolitic associations and early Variscan high-pressure metamorphic rocks, which denote their oceanic affinities and subduction-related processes, respectively [*Martínez Catalán et al.*, 1996]. Of these, only the easternmost units of the Cabo Ortegal Complex might have been sampled by the line ESCIN-3.3 (Figure 2).

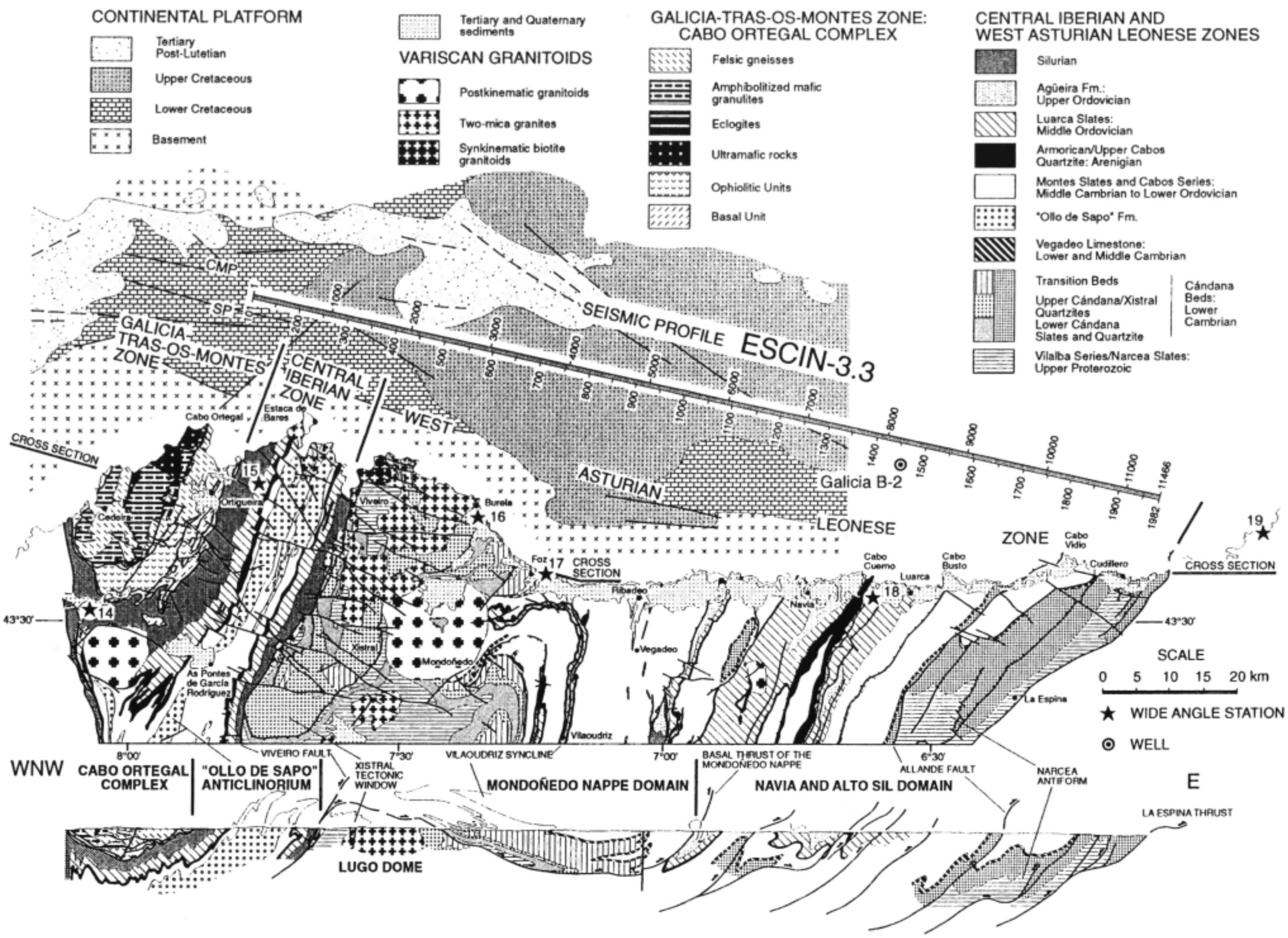


Figure 2. Geologic map and cross section of the northern part of the Iberian Massif in the area close to line ESCIN-3.3. The location of the line is also shown, together with that of the land stations and the Galicia B-2 marine drill [*Querol*, 1987]. Geology is after *Marcos* [1973], *Bastida et al.* [1982], *Martínez Catalán* [1985], and *Arenas* [1988]. Offshore geology is after *Lamboy and Dupeuble* [1975].

The northwestern part of the Iberian Peninsula underwent important post-Paleozoic tectonic activity. The opening of the Bay of Biscay between the Upper Jurassic and the Early Cretaceous [*Boillot et al.*, 1984] caused a counterclockwise rotation of the Iberian Peninsula with the subsequent creation of oceanic crust and the development of important sedimentary basins in the newly created passive margins [*Boillot et al.*, 1984; *Srivastava et al.*, 1990].

The northern Iberian margin became active as a consequence of the convergence between the Iberian and European plates during the Tertiary. The collision between Europe and Iberia triggered the raising of the Pyrenees and led to a shortening of 120-130 km [*Boillot and Capdevila*, 1977; *ECORS Pyrenees Team*, 1988; *Choukroune and ECORS Team*, 1989; *Roure et al.*, 1989; *Muñoz*, 1992]. The Cantabrian Mountains [*Alonso et al.*, 1995] are the westward prolongation of the Pyrenees. Northward imbrication of the Iberian crust beneath the European crust [*Suriñach et al.*, 1993; *Pulgar et al.*, 1996, *Fernández Viejo*, 1997] and partial southward subduction of the oceanic crust of the Bay of Biscay below the continental margin of Iberia [*Boillot et al.*, 1979; *Grimaud et al.*, 1982; *Boillot*, 1986] are the deep crust evidences of this Tertiary convergence.

Several commercial multichannel seismic profiles cover the Asturian and northern Galician margins, including a network of profiles striking N-S and E-W, acquired in 1977 by Empresa Nacional de Investigación y Exploración Petrolifera Sociedad Anónima (ENIEPSA) and named LC77. These lines cover the continental platform between Cabo Ortegal and Foz, reaching 5 s (two-way travel time (TWTT)), and give information on the Mesozoic to recent sedimentary cover and tectonics. These profiles have been used in this study, and references to them will be made eventually.

Another seismic experiment carried out in NW Iberia in 1982 recorded several refraction lines in the western part of the WALZ, in the northern part of the CIZ, and in the GTMZ [*Córdoba*, 1986]. Results reported by *Córdoba et al.* [1987, 1988], *Téllez* [1993], and *Téllez et al.* [1993] indicate the existence of a high-velocity (6.55-6.7 km/s) body located at shallow crustal levels (2.5-8 km) underneath the Lugo Dome. These high velocities might be the manifestation of a lower crustal layer situated in abnormally shallow depths. On this basis, *Vegas and Córdoba* [1988] speculated about the existence of a layer of granulites situated at only 8 km below the surface in the antiformal structure, carried there by a crustal-scale Variscan thrust.

Aller et al. [1994] modeled a prominent feature of the aeromagnetic map of Spain [*Ardizone et al.*, 1989], the so-called Eastern Galicia Magnetic Anomaly (EGMA) that accounts for

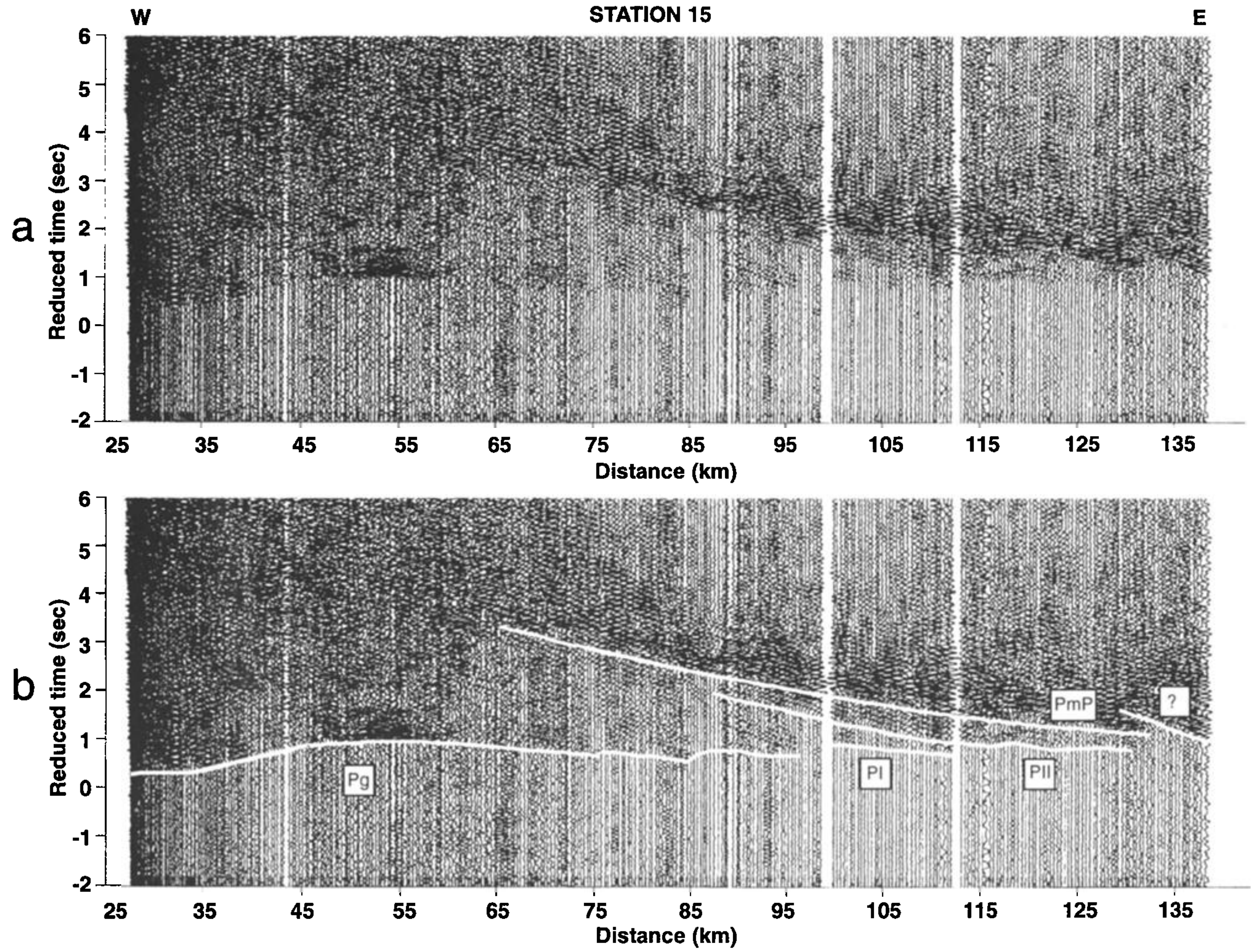


Figure 3. (a) Wide-angle record section for land station 15. (b) Interpreted record section.

magnetic intensities up to 190 nT and is situated along the above mentioned Lugo Dome. Their models show that this magnetic anomaly can be produced by a middle to lower crustal layer a few kilometers thick, with a magnetic susceptibility of approximately 0.045 (in SI units) carried to shallow crustal levels by a thrust with ramp and flat geometry. The thrust should be rooted in the lower crust, and therefore these authors consider this highly magnetic body to be composed of mafic and/or ultramafic material of lower crustal provenance. This structure is in close agreement with the surface geology, since the Lugo Dome and the Viveiro Fault can be interpreted in relation with the displacement along this deep-rooted Variscan thrust.

3. The Wide-Angle Reflection Data

3.1. Acquisition and Correlation

The line ESCIN-3.3 was shot in February 1993 onboard the M/V *Seisquest*, contracted by Schlumberger-Geco-Prakla. It has a length of 141 km running with a N102°E direction at an average distance to the coast of around 20 km. Vertical incidence and wide-angle data were recorded, the latter by six autonomous, three-component Mars 88 Lennartz stations, located in Figure 2 by the stars numbered 14 to 19. The location of stations, not in line with the profile (except station 19), constitutes a handicap to fit all the record sections with a unique identical model.

The phase correlation was carried out in record sections (receiver gathers with a reduced velocity of 6 km/s), corrected for bathymetry with normalized amplitudes, and band-pass filtered (3-12 Hz). Most of the stations show very clear arrivals that have been correlated with four main phases. The maximum recording distance corresponds to station 14 (160 km). However, this record section shows a very low signal to noise ratio which hampers the correlation of phases. The rest of the stations registered arrivals in a smaller range of distances (from 25 km up to 135 km), insufficient to identify phases coming from very deep reflectors. We will only show the two more representative record sections, corresponding to stations 15 and 18 although the arrivals from all six stations have been considered in the construction of the final model. Station's 15 record section is shown in Figure 3 and the derived model for this station appears in Figure 4. Station's 18 record section is shown in Figure 5 and the derived model in Figure 6.

In all cases, the only refracted wave train identified as first arrival is the *Pg* (Figures 3 and 5). It is observed in the different

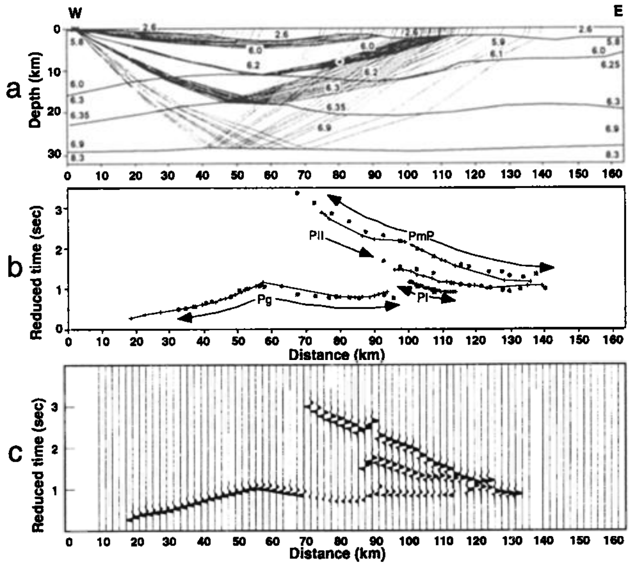


Figure 4. (a) Ray tracing for station 15. (b) Observed (asterisks) and model (crosses) arrival times for the same station and correlated phases. (c) Synthetic seismogram for the model.

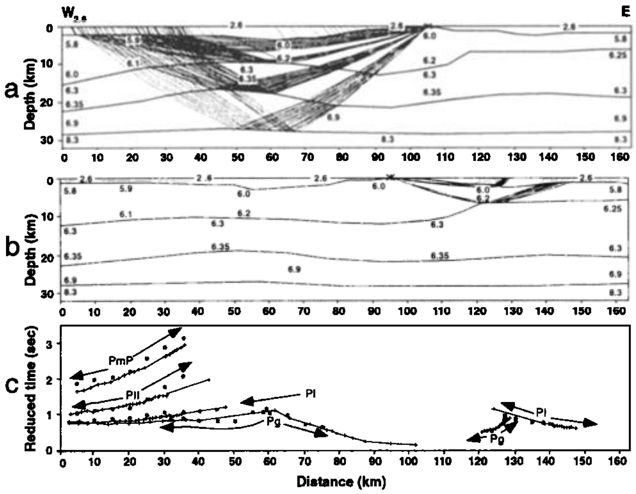


Figure 6. (a) Ray tracing for station 18, western part. (b) Ray tracing for station 18, eastern part. (c) Observed (asterisks) and model (crosses) arrival times for the same station and correlated phases.

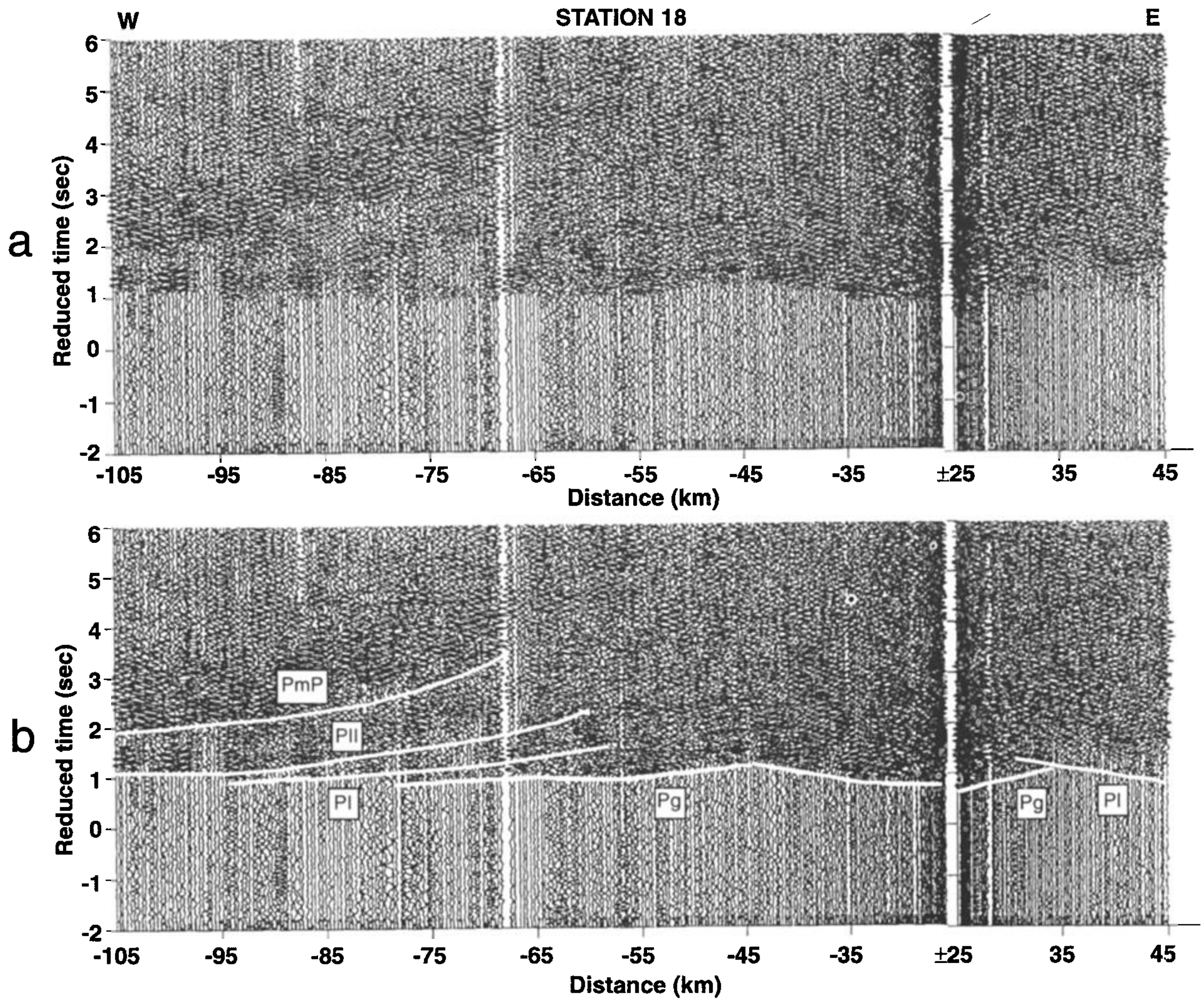


Figure 5. (a) Wide-angle record section for land station 18. (b) Interpreted record section.

stations in distances ranging from 25 to 90 km, with an average apparent velocity of 6 km/s that locally exhibits important variations. These arrivals are interpreted as the refracted phase of a basement, with velocities around 6 km/s. The delays of up to 0.7 s of reduced time observed in this phase as well as the changes in its apparent velocity (5.1 km/s to 6.5 km/s) are likely related to the remarkable variations in thickness of the sedimentary layer (Figures 4a and 6a).

Several wide-angle reflection arrivals can be identified. Shortly delayed from *Pg*, there is a usually low-amplitude *PI* phase. After 0.2-0.7 s, a new phase, *PII*, with higher amplitudes appears. Both *P*I and *P*II phases are interpreted as intracrustal wide-angle reflections. Finally, a high-amplitude *PmP* train wave, the energy-dominant phase after 55-65 km, can be correlated over long distances until the end of most of the record sections. This phase is interpreted as the wide-angle reflection in the crust/mantle boundary. The transition between *PII* and *PmP* appears to be highly reverberative.

3.2. Velocity/Depth Model

Forward modeling techniques have been used to develop the final velocity model. We have used two different programs: MacRay [*Luetgert*, 1992] and Rayamp [*Spence et al*, 1984], the second of which permits the modeling of amplitudes in order to test the velocity contrasts. The ray tracing and arrival times for stations 15 and 18 are shown (Figures 4 and 6) as well as the synthetic seismogram for station 15 (Figure 4c).

Only one of the stations (19, Figure 2) was in line with the profile. The rest of them were located off the line. To handle this kind of configuration, we modeled separately the eastern and western parts of every record section. The range of distances considered in each model was taken from the distance between every shot point and both ends of the line. As the strike of the Variscan structures is perpendicular to the profile and therefore is not supposed to change in a ~N-S direction, the implied error is not very important and decreases toward the end of the line. Besides, the 5 s deep lines ENIEPSA LC77 helped to keep an extra control on the Alpine structures. The models obtained for all the stations look alike and represent the same velocity/depth configuration as the one obtained for station 19.

As discussed in section 3.1, we interpret the *Pg* phase as the refraction in a basement with an average velocity of 6 km/s. Overlying this basement, sedimentary rocks with an average velocity of 2.6 km/s and variable thickness explain the apparent velocity variations observed in the *Pg* phase. Two sedimentary basins have been included in the models. Their existence is supported by the vertical incidence data (see below). The western basin is well constrained and reaches a depth of around 3 km. The eastern one, less clearly defined in both data sets, shows slight variations in the models, depending on the azimuthal coverage of the station that recorded the arrivals.

An interface located at a depth between 7 and 12 km is responsible for the *PI* wide-angle arrivals. According to the model, this boundary implies a low-velocity contrast of 0.1-0.2 km/s that therefore justifies the low amplitudes that the *PI* phase usually exhibits. Only in the record section of station 17, can anomalously high-amplitude *PI* arrivals be identified [*Ayarza*, 1995] and seem related to some energy-focusing mechanism originated by the local concave shape of the interface in the reflection zone. The evolution of this interface toward the east is constrained by *Pg* arrivals that are correlated over shorter distance ranges for stations 18 (eastern part) and 19 and by a *PI* phase that appears earlier in time and distance for those two stations. This indicates a decrease in depth toward the east of the above mentioned interface, reaching 6-7 km in the easternmost part of the profile. From wide-angle data, the meaning of this interface is unclear. It could represent a detachment observed in the vertical incidence data set (see sections 4.2.2 and 4.3.2) which does not imply an important velocity contrast.

Another interface with a higher velocity contrast (6.3-6.9 km/s) is responsible for the higher-amplitude *PII* reflected arrivals. This boundary shows in the models an average depth of around 20 km, getting slightly shallower (17 km) below the area that would represent the northward prolongation of the Lugo Dome. The high velocity of the lower layer (6.9 km/s), often found in the lower parts of the continental crust of this and neighboring areas, [*Córdoba*, 1986; *Téllez*, 1993; *Pulgar et al.*, 1996; *Fernández Viejo*, 1997], suggests that the interface represents the limit between the middle and lower crust.

The deepest modeled interface is responsible for the *PmP* arrivals recorded by all the stations. Located at a depth of about 29 km, according to the models, it represents the Mohorovicic discontinuity. The high amplitude of the arrivals and the short critical distances at which they appear are the result of a rather high contrast on the *P* wave velocities (6.9-8.3 km/s). The existence of converted phases *PmS* [*Ayarza*, 1995] supports the first-order discontinuity character of this interface. Even though smaller velocities for the lower layer (8.1-8.2 km/s) do not match the short critical distances of the *PmP* phase and imply a significant decrease of its amplitude, constraints based on a *Pn* refracted phase would be necessary to properly define the mantle velocity and its gradient.

The reverberative and complex nature of the transition between *PII* and *PmP* may indicate the layered character of the lower crust [*Ayarza*, 1995] as discussed by *Wenzel et al.* [1987]. The layered signature of the lowermost crust is well depicted by the vertical incidence data (see Plate 1 and section 4.2.3).

The models developed for the different stations, supported by the information obtained from the vertical incidence data (see section 4, Figures 7 and 8 and Plate 1), have been integrated in a final velocity model which will be shown with a geological interpretation of the vertical incidence data in Figure 9 . The fact that most of the stations are not in line with the profile adds a certain imprecision to this final velocity/depth distribution, but it is considered a reasonably good representation of the deep structure of the crust.

It is worth mentioning that multiple refraction/wide-angle experiments have been carried out in this part of the Iberian Massif, and most of them resulted in similar velocity/depth models [*Córdoba*, 1986; *Córdoba et al.*, 1988; *Téllez*, 1993; *Pulgar et al.*, 1996; *Fernández Viejo*, 1997]. High mantle velocities (8.2-8.3 km/s) seem to be quite common in this area [*Córdoba et al.*, 1987, 1988], and the depth of the modeled interfaces coincides with that found in other experiments for the NW Spanish Variscan crust [*Córdoba et al.*, 1988; *Téllez*, 1993]. Lateral variations of velocity (6.7-7.1 and 8.1-8.3) have been modeled in the lower parts of the crust and in the mantle by *Fernández Viejo* [1997] over much longer lines (≥ 200 km). The reduced lateral extension of our record sections (≤ 25-135 km)

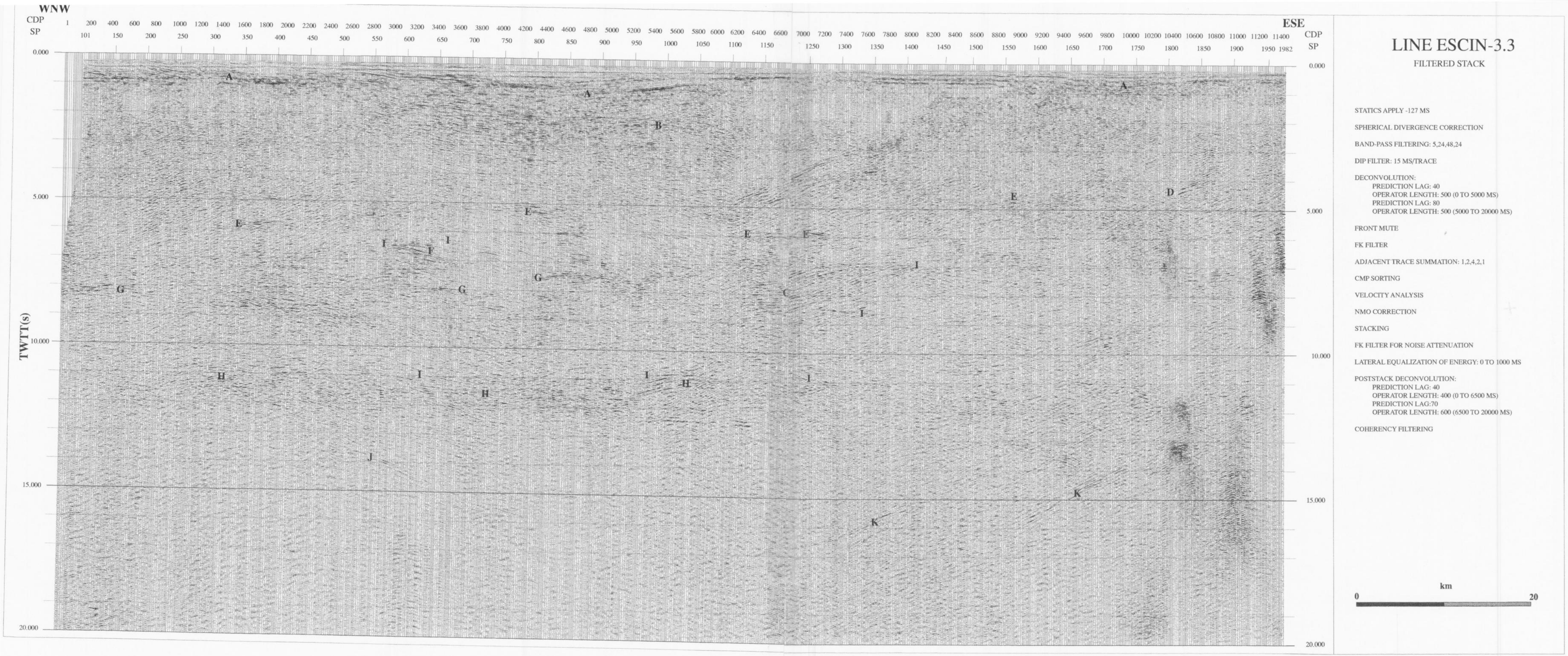


Plate 1 Stack section of the

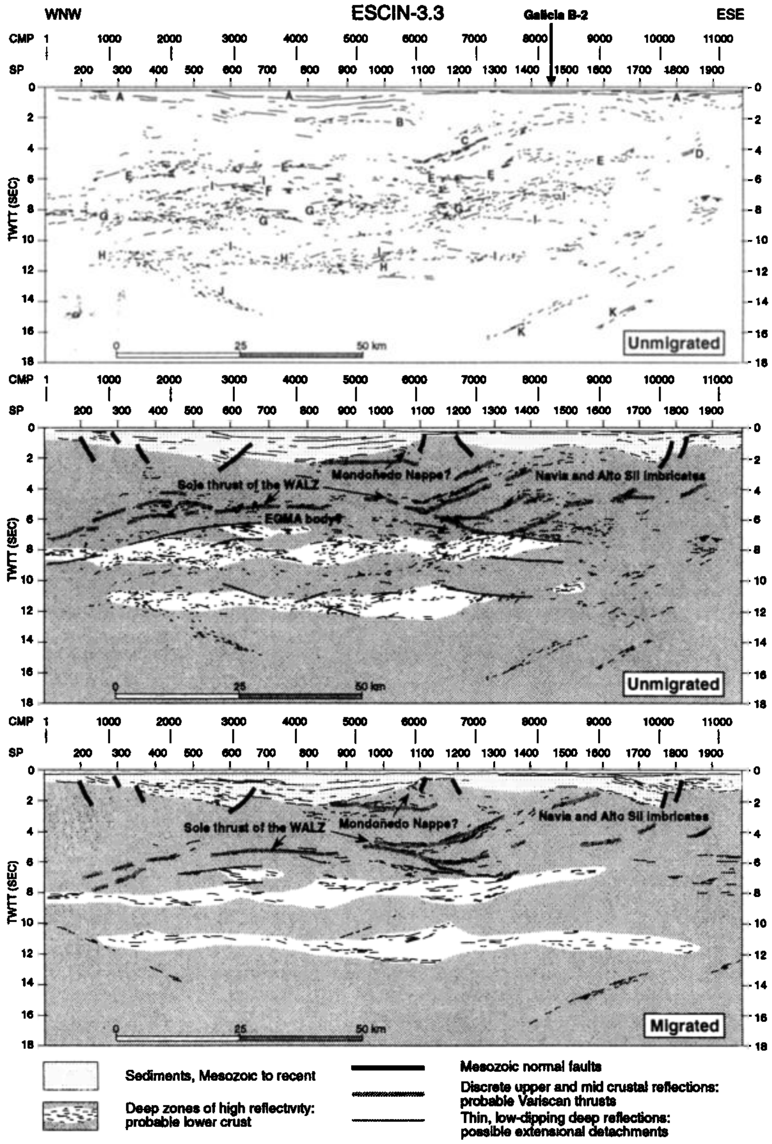
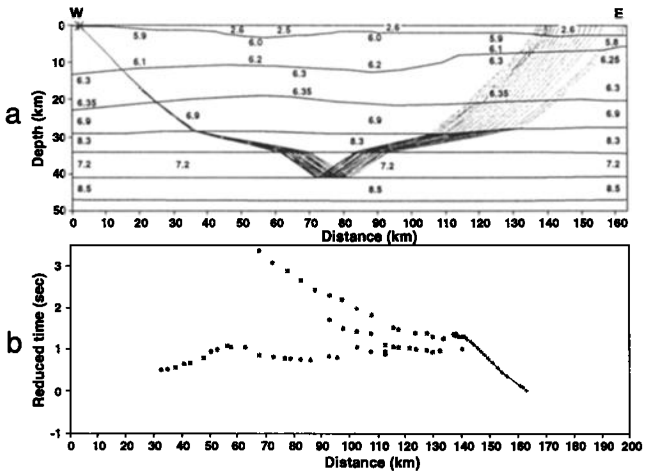


Figure 7. (a) Line drawing of the stack section with indication of the main reflections described. The location of the Galicia B-2 marine drill is shown. (b) Interpreted line drawing from the unmigrated stack. (c) Interpreted line drawing from the migrated stack.



together with the complexity of the data hampers the detailed modeling of lateral variations of velocity and therefore adds a certain imprecision to our results.

4. The Vertical Incidence Data

4.1. Acquisition and Processing

The acquisition parameters for the near vertical incidence profile are displayed in Table 1.

Figure 8. (a) Ray tracing for the velocity/depth model developed for station 15, with the addition of a lower crustal layer between 34 and 41 km. (b) Observed arrivals for station 15 (asterisks) and for the second crust/mantle interface modeled at 41 km (crosses).

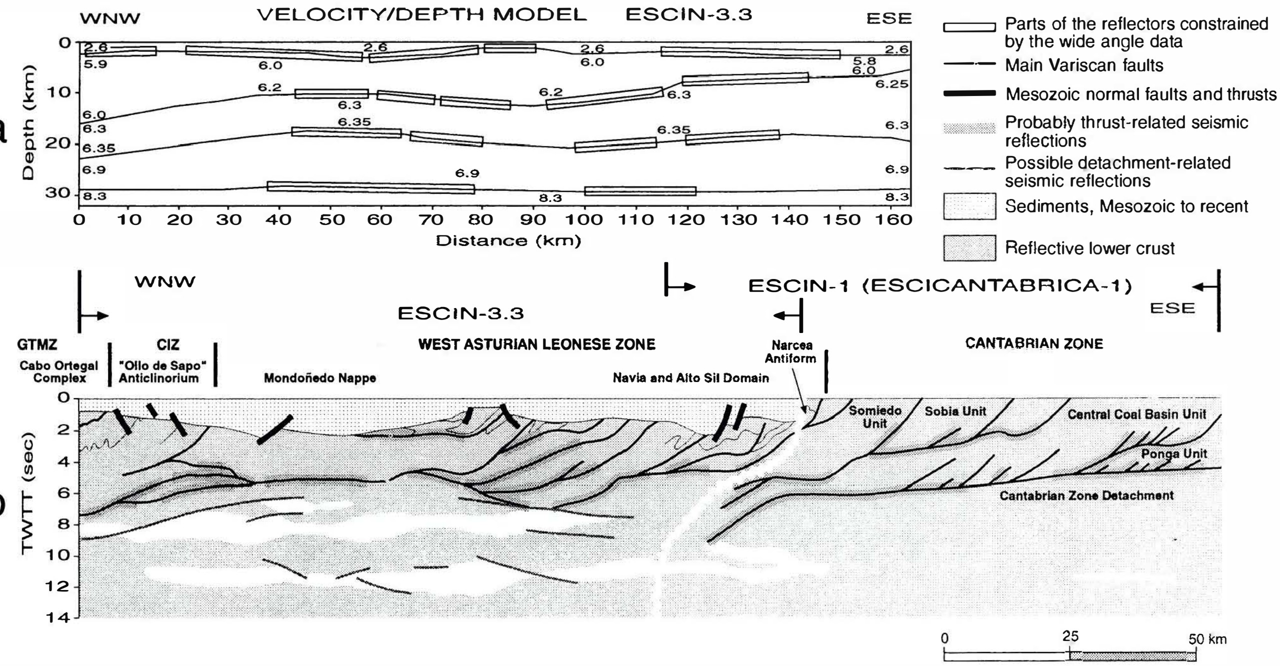


Figure 9. (a) *P* wave velocity/depth model for the crustal section sampled by line ESCIN-3.3 deduced after study of wide angle reflected and refracted arrivals recorded by the land stations. In rectangular boxes, are the parts directly constrained by the wide-angle data. The rest of the model has been extrapolated using the vertical incidence data. (b) Interpretation of seismic lines ESCIN-1 [*Pérez-Estaún et al.*, 1994] and ESCIN-3.3 (this study), displayed together to show a long geotraverse of this part of the Variscan Belt.

The quality of the vertical incidence data has been partly conditioned by the proximity of the line to the coast (20 km) and the shallow water depth in the area (80-150 m). The first version of the stack exhibited an energetic low-velocity (1300 m/s), low-frequency noise (8-13 Hz) sometimes dipping to the high offsets and sometimes the opposite, characteristic of marine data, as it is related to the movement of the streamer. Reprocessing of the data was undertaken to increase the signal/noise ratio. The processing flow is shown in Table 2. The unmigrated stack is shown in Plate 1, and a line drawing of the stack, with and without interpretation, plus an interpreted line drawing from the migrated stack are depicted in Figure 7.

4.2. Description and Interpretation

The line ESCIN-3.3 was shot across strike of the two easternmost hinterland zones of the Iberian Massif (WALZ and CIZ) and subparallel to the Alpine structures, that is, in a privileged position to image the Variscan crustal structures. Line ESCIN-1, aimed to study the Variscan foreland [*Pérez-Estaún et al.*, 1994, 1995], and lines ESCIN-2 and ESCIN-4 [*Pulgar et al.*, 1995, 1996], focused on the Alpine structures, will be referred to in this paper for interpretation purposes.

The stack section shows many highly reflective and coherent events at different depths. What follows is a description of the main groups of reflections observed, considering four intervals which refer to four distinct ranges of depths: shallow, intermediate, deep, and very deep reflections. The most important reflectivity features will be described and related with the wide-angle model already presented.

4.2.1. Shallow reflections. Included in this group are the events appearing in the upper 2-3 s (A in Plate 1 and Figure 7a). Starting at the western part of the line, a group of subhorizontal to east dipping reflections (A) can be followed up to Common Mid Point (CMP) 4400, where they change their inclination, defining a synformal structure. Nearby CMP 6400, they become subhorizontal again and narrower, and to the east, their reflectivity and continuity decrease.

Table 1. Acquisition Parameters for Line ESCIN-3.3

Parameter	Value
Air guns volume	5490 cu.in (90 l)
Pressure	2000 p.s.i.(13.8 MPa)
Offset	240 m
Shot interval	75 m
Receiver interval	360 groups spaced 12.5 m*
Fold	30
Record length	20 s
Sample rate	4 ms, resampled to 8 ms

*posterior adjacent trace summation implies 180 groups with a 25 m trace interval.

Table 2. Processing Sequence for Line ESCIN-3.3

Process	Value
Prestack Processing	
Statics corrections	-127 ms
Spherical divergence correction	
Bandpass filtering	5, 24, 48, 24
Dip filtering: 15 ms/trace	
Deconvolution	
Prediction lag	40 ms
Operator length	500 ms (0 to 5000 ms)
Prediction lag	80 ms
Operator length	500 ms (5000 to 20000 ms)
Frontal mute	
Frequency-wave number (FK) filtering	
Adjacent trace summation	1, 2, 4, 2, 1
Common Mid Point (CMP) sorting	
Velocity analysis	
Normal Move Out (NMO) correction	
Stacking	
Poststack Processing	
Frequency-wave number (FK) filter for noise attenuation	
Lateral equalization of energy	0 to 1000 ms
Poststack deconvolution	
Prediction lag	40 ms
Operator length	400 ms (0 to 6500 ms)
Prediction lag	70 ms
Operator length	600 ms (6500 to 20000 ms)
Lateral coherency filtering	
Frequency-wave number (FK) migration	(5 km/s)

This group of reflections seems to be the seismic image of two sedimentary basins, one in the western half of the line and the other in its eastern part. It is difficult to determine the maximum thickness of sediments in the basins because of the large number of multiples. In the western basin, a group of subhorizontal reflections (B) seem to be truncated by the bottom of the sedimentary basin at a depth of 2.3 s, that is, about 3 km (average velocity of 2.6 km/s deduced for the sedimentary basin from wide-angle data on the basis of the *Pg* phase, see section 3.2). Therefore that will be considered the maximum depth of this basin. The limits of the eastern basin are unclear.

4.2.2. Intermediate depth reflections. This group is composed of reflections identified in the time range of 2-3 to 6 s. The shallowest of them are the above-cited reflections (B) appearing at CMP 4600 at about 2.3 s (Plate 1 and Figure 7a). They are east dipping, low-energy events that disappear around CMP 6150 at 2.5 s. At greater depths (4-5 s) and around CMP 5600, there is a group of east dipping reflections that close to CMP 6200 are substituted by a group of very energetic west dipping events (C in Plate 1 and Figure 7a). The latter can be followed until CMP 8200, where they reach 2 s. Several packages of less energetic and less continuous west dipping reflections can be found to the east around CMPs 7700, 9600, and 10,600 (D) between 3 and 5 s.

At greater depths, the west dipping reflections (C) merge with a group of subhorizontal reflections (E). The same type of subhorizontal reflections are found around 6 s in the western and central parts of the line and slightly shallower (5 s) in the eastern part, where they represent the lower limit of the west dipping reflections (D in Plate 1 and Figure 7a).

Given the depths of these reflections, an origin in the Variscan basement can be envisaged. Their dips as a whole match that of the structures identified to the south, on land, so that a relation with the fold and thrust tectonics that characterize the WALZ seems straightforward. The C-D reflections can be either the limbs of the overturned and recumbent folds or, most probably, the thrust faults and associated ductile shear zones and the E group, a sole thrust in which they are rooted [*Pérez-Estaún et al.*, 1991].

4.2.3. Deep reflections. At 6.5-7 s and between CMPs 3000 and 4000, there is a group of very energetic and convex reflections, with limited lateral extent (F). This package overlies deeper series of subhorizontal energetic arrivals (7-9 s) with similar reflectivity, but they are fairly more continuous and widespread (G). According to their high reflectivity and to the velocities that seem to appear in the wide-angle models (6.9 km/s) for equivalent depths, they could all arise from lower crustal material. The base of the G group of reflections is comparable to the crust-mantle boundary modeled at a depth of 29 km (Figure 9a).

A still deeper band of subhorizontal reflections (H) with a reflectivity and a lateral continuity similar to those described above (G) occurs at 11-12 s. According to its characteristics, we consider the possibility that it corresponds also to lower crustal material. Arrivals from this possible deep crustal layer have not been identified in the wide-angle record sections. However, if a second layer of this type is introduced in the wide-angle model at the appropriate depth (34-41 km), the crust-mantle wide-angle reflection could be observed only in the case of station 15 and after distances of 135-140 km (Figure 8) when a mantle velocity of 8.5 km/s is considered. Lower values of this velocity would result in critical distances far beyond the boundaries of this and any other of our receiver gathers. Therefore the distances at which we could identify an hypothetical *PmP* from the deepest crust/mantle discontinuity would probably exceed that of our wide-angle records. In the case of station 15, some arrivals observed after the *PmP* correlation at the far end of the record section (question mark in Figure 3b) could correspond to a hypothetical second crust-mantle boundary with the modeled characteristics of velocity and depth. Anyhow, the limited lateral extent of this arrival derives in a very imprecise correlation and does not allow a reliable modeling.

The low-reflectivity narrow area (1 s) between the two above mentioned reflective bands (G and H) should include upper mantle material, because of the *P* wave velocity (8.3 km/s) deduced from the high amplitudes and short critical distances obtained for the *PmP* phase in the wide-angle data. The limited thickness of this band may be the reason why no refracted phase *Pn* from the mantle is observed.

Therefore we are dealing with what seems to be a threefold repetition of the lower crust (F, G, and H). The bottom of the deepest reflective band (H), equivalent to a second crust-mantle boundary, appears at around 40 km, an anomalous depth for the Mohorovicic discontinuity of Variscan crust unaffected by the Tertiary compression, which commonly has been reequilibrated and shows this interface at around 30 km [*Meissner*, 1986]. It is this part of the present lithosphere what will be referred to as a crustal root.

A feature that may help in the interpretation of the age and relationships of the reflective lower crustal bands is the presence of weak and narrow reflections, slightly dipping to the east or west (I), which crosscut the individual reflections of the three strongly reflective bands (F, G, and H). In particular, the shallowest of them (F) is truncated by one of these reflections in such a way that its upper part seems to have been removed out of the section of the profile (Plate 1 and Figure 7). The same stands for the intermediate band (G) between CMPs 7000 and 8000 at 7 s and for the deepest one (H) between CMPs 6300 and 8000 at 11-12 s.

Accordingly, the weak and narrow reflections (I) appear as low-angle detachments. However, the strongly reflective packages do not seem to be repeated by the presumed subhorizontal faults, a situation that would be expected if they were of the thrust type, which are additive structures. Conversely, the lack of these packages above the faults can be expected from an extensional fault moving in a direction at high angle to the plane of the profile, because normal faults are subtractive. Therefore we interpret such reflections as normal detachments affecting a reflective lower crust. Similar types of reflections have often been found at deep crustal levels of extended areas although their interpretations do not seem to be unique. *Blundell et al.* [1989] see them as anastomosing shear zones that accommodate the crustal extension at lower crust depths, whereas *Bois* [1992] argues that they are the remnants of Paleozoic or older structures. The first of the interpretations seems to be more in agreement with our data.

In the migrated section, these detachments are very poorly identified, and the deepest among them disappear, probably because of its scarce energy and the migration effects (Figure 7c).

4.2.4. Very deep reflections. This group consists of a series of inclined reflections reaching depths greater than 12 s. Around CMP 1500 and from 11.5 s downward, there is a group of reflections dipping to the east that reaches 15 s around CMP 3800 (J). Another two groups of inclined reflections, dipping to the west, appear in the easternmost part of the line at depths between 11 and 17 s (K). None of these bands disappears after migration.

The interpretation of these deep reflections is troublesome, and the possibility that they represent artifacts has been considered. They could be side echoes coming from faults with azimuths roughly parallel to the coast and slightly oblique to the profile (Figure 2), as the ones described by *Lamboy and Dupeuble* [1975]. However, the ENIEPSA LC77 profiles transverse several of these faults with angles similar to that of ESCIN-3.3, and they never produce reflections comparable in intensity or continuity to the events described here. Therefore they seem to have a deep source. They could be either branches of diffraction hyperbolae or represent dipping reflectors, though in this case, their strike might be oblique to that of the seismic line and represent lateral reflections coming from outside the plane of the profile.

5. Geological Correlation and Implications

This section intends to establish the correspondence between the data obtained from the profile and well-known geological events, pointing out the more important contributions of this seismic study to the geological knowledge of the area.

5.1. Sedimentary Cover

The two sedimentary basins identified in the uppermost 2-3 s should be related with the opening of the Bay of Biscay. The westernmost of them is the best defined and does not show evidence of any compressional tectonic structures, probably because the orientation of the line is not appropriate to show the roughly E-W Alpine structures. In the N-S ENIEPSA LC77 profiles, inverted extensional faults and a few folds can be seen in this basin. These structures are related to the Alpine convergence.

5.2. Variscan Structure of the Basement

A correlation of the intermediate depth reflections with structures identified on land 12 to 32 km to the south of the seismic line can be established. The Basal Thrust of the Mondoñedo Nappe defines a very open synform in the coast with an axial plunge of around 10°S (Figure 2). If the plunge and direction of this structure remain constant in the continental platform to the north, the flat bottom of the synform might have been imaged by the profile, roughly between CMPs 4000 and 6000. The shallow reflections (B) that appear truncated by the Mesozoic sediments in this part of the line might clearly correspond to this particular structure (Figure 7).

Owing to their position, the west dipping reflections (C and D) may correspond to imbricated thrust sheets of the WALZ (compare Figures 2 and 9b). The fact that all these events seem to terminate against or merge downward with a subhorizontal group of reflections (E) suggests that the latter might represent the sole thrust of the WALZ. The existence of such a thrust, which would represent the floor detachment of that zone, was proposed by *Pérez-Estaún et al.* [1991] on the basis of structural fieldwork in the area.

Important structural elements such as the Viveiro Fault, the "Ollo de Sapo" Anticlinorium, and the Cabo Ortegal Complex (Figure 2) have not been identified in this profile. As a matter of fact, steeply dipping structures like the Viveiro Fault or the limbs of the anticlinorium are very seldom imaged through standard processing of vertical incidence seismics. The Cabo Ortegal Complex shows a general synformal structure but has a complicated internal geometry. Its position on land suggests that it might have been partly imaged by the western edge of the profile. Its apparent absence may be explained either by the complexity of its internal structure or by the possibility that the synform closes or narrows to the north.

5.3. Structure of the Lower Crust

The upper package of strong, convex deep reflections (F) occurs in the place where the high-velocity (6.9 km/s) crustal layer reaches the shallower position in the wide-angle models: the offshore prolongation of the Lugo Dome. It is in this structure where *Córdoba* [1986], *Córdoba et al.* [1987], and *Téllez* [1993] located a high *P* wave velocity body (6.55 km/s) at a depth between 2.5 and 8 km, interpreted as a lower crustal layer at higher crustal levels. Also in the Lugo Dome in the most septentrional area of the Mondoñedo Nappe, *Aller et al.* [1994] found that the body with high magnetic susceptibility causing the EGMA was located at its minimum depth (2.5 km), deepening down to 12.5-15 km toward the south.

Those seismic and magnetic anomalies can be considered as different manifestations of the same lower crustal slice, located at relatively high crustal levels. The northward prolongation of this body is poorly constrained, because the magnetic anomaly maps have sampled only as far north as the shoreline. However, important geologic structures mapped in the WALZ have their image in the seismic line. It is then straightforward to associate the upper package of deep reflections (F) to this shallow body of lower crustal material. The depth at which it appears in the line ESCIN-3.3 is not anomalous if we consider the fast southward deepening of this body in the *Aller et al.* [1994] models. Besides, this slice is located right below a 2.3 s thick Mesozoic sedimentary basin and, consequently, at only 3.8 s below the top of the Variscan basement, that is, around 12 km.

The previous considerations suggest that the highest of the lower crust repetitions imaged in the profile is of Variscan age, as it is related to a Variscan structure, the Lugo Dome. Estimations about the age of the lowest of the repetitions (reflective band H) are not so straightforward. Some features exhibited in the profile may help to constrain the possibilities, but interpreting geological features in a seismic line which is parallel to the Alpine structures implies missing the effects of the Tertiary compression. This issue will be discussed in section 6.

5.4. Deep Alpine Tectonics

The Pyrenean deformation in the area seems to be concentrated in a band, roughly oriented E-W, bounded to the south by the northward imbrication of the Iberian crust beneath the Cantabrian crust [*Pulgar et al.*, 1996; *Fernández Viejo*, 1997] and to the north by the southward subduction of the oceanic crust of the Bay of Biscay below the continental margin of Iberia [*Boillot et al.*, 1979; *Grimaud et al.*, 1982; *Boillot*, 1986].

It is tempting to link the very deep reflections (J and K) with the above mentioned crustal imbrication and subduction. Besides, the deep east dipping (J) features at the western end of the line seem to crosscut the deepest band of the duplicated lower crust in the migrated section (Figure 7c). However, the frontal thrust representing the imbrication of the Duero Basin beneath the Cantabrian Mountains has an E-W orientation [*Pulgar et al.*, 1996], and the same holds for the tectonic accretionary prism related to the partial closure of the Bay of Biscay (Figure 1). Given the weak obliquity between these structures and the profile, this would imply that if the so-called very deep reflections were an image of the mentioned Alpine features, they should represent mainly lateral arrivals from dipping reflectors located off the seismic line.

Nevertheless, it is also possible that the effects of the Alpine compression in the Iberian crust are represented by the deepest of the repetitions of lower crust. The ongoing Tertiary compression propagated from east to west and triggered the partial subduction of the oceanic crust of the Bay of Biscay. As this subduction could not progress, the compressive stresses provoked indentation of continental fragments [*Fernández Viejo*, 1997] in a more or less E-W direction, that is, parallel to the profile. The lowest of the reflective bands seen in the profile could represent one of these imbrications (Figure 10c).

6. Constraints on the Origin and Age of the Crustal Root

The most controversial and striking feature of line ESCIN-3.3 is the likely duplication shown by the lower crust along most of the profile, what we have called a crustal root. In this section, we will discuss the age and likely origin of this feature. It is worth mentioning again that the direction of the line along the strike of the Alpine structures permits a good identification of Variscan structures but hampers that of the Alpine events. Besides, the facts that the band of high reflectivity (H) is basically parallel to the subhorizontal band appearing between 7 and 9 s (G, Plate 1 and Figure 7a) and that no crosscutting relationships can be established between them hinder a straightforward interpretation. The lines ESCIN-2 and ESCIN-4 (Figure 1) [*Fernández Viejo*, 1997], shot with the aim of imaging the influence of the Alpine cycle in the Variscan crust, show an important imbrication of lower crust, which is Tertiary in age. However, both lines are located toward the east, in the Duero Basin and CZ. As this part of the Iberian Peninsula is affected by NW-SE late Variscan strike-slip faults, the development of the Alpine structures in a E-W direction might have been influenced by the previous existence of these faults, and its continuation to the west of the CZ is not known. A N-S profile through the WALZ and/or CIZ would shed some light on this controversy, as an Alpine slab would be imaged as a north dipping feature [see *Choukroune and ECORS Team*, 1989; *Roure et al.*, 1989; *Muñoz*, 1992; *Pulgar et al.*, 1996; *Fernández Viejo*, 1997] and a Variscan one would appear subhorizontal.

Several facts should be considered in order to justify the possible existence of a Variscan crustal root in this particular area of the Iberian Massif.

First, the established correlation of the shallower reflective band (F) with the high *P* wave velocity and high magnetic

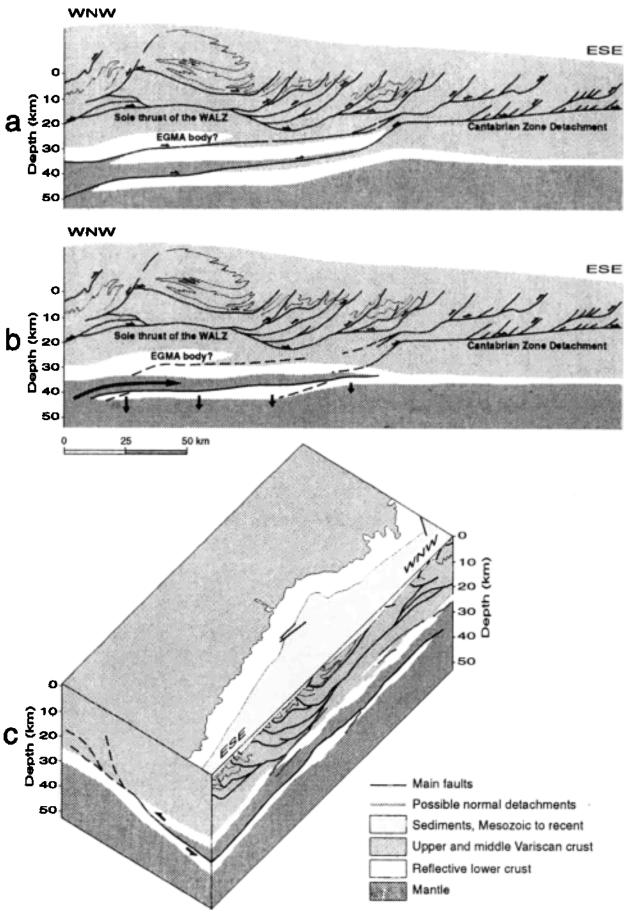


Figure 10. Three different hypothesis for the origin of the identified crustal root: (a) preserved Variscan underthrusting of part of the Cantabrian Zone Basement, (b) arrested delamination affecting a thickened Variscan crust, and (c) Alpine imbrication of the Iberian continental crust to the north.

susceptibility body (EGMA), emplaced during the Variscan thrust tectonics, implies that a reflective lower crust already existed and was repeated at least once during the Variscan deformation.

Second, the age of the reflectivity of the presently duplicated lower crust has to be Variscan. The weak and narrow deep reflections (I) interpreted as normal detachments support this interpretation. Two extensional episodes can account for these structures: the synorogenic to postorogenic Variscan extensional collapse and the Mesozoic extension linked to the opening of the Bay of Biscay. Several Variscan extensional detachments have been described in neighboring areas of the basement [*Escuder Viruete et al.*, 1994; *Díez Balda et al.*, 1995, *Martínez Catalán et al.*, 1996], and many of them show a tectonic direction of transport parallel to the trend of the belt. This would imply a N-S direction of extension in the area of the profile, the same expected for the Mesozoic tectonics. Consequently, both extensional episodes may explain the observed truncation relationships. If the detachments are Variscan, the reflectivity is probably also Variscan, though earlier. If they are Mesozoic, the age of the reflectivity is not constrained, but a Variscan age is still possible.

Third, an evaluation has to be made of the width of the duplication. In the profile ESCIN-1 (Figure 1) carried out on land in the foreland thrust belt (CZ), *Pérez-Estaún et al.* [1994] argue the existence of two main thrusts: one of them running at a depth of 3-4 s and a deeper one, between 4 and 6 s, named the Cantabrian Zone Detachment. A feasible interpretation of the ESCIN-1 profile has been drawn in Figures 9b and 10a together with the interpreted ESCIN-3.3. Both profiles are not continuous, but they are complementary in the construction of a transversal section of the Variscan Belt, and it is useful to display them together. Note that both profiles overlap to some extent because both have covered the Narcea Antiform.

In the line ESCIN-1, the Cantabrian Zone Detachment continues to the west through a ramp, reaching 10 s below the Narcea Antiform (Figure 9b). It might continue farther to the west between the two lower crustal layers observed in the profile ESCIN-3.3 and be responsible for its duplication (Figure 10a). For this relationship to be geologically possible, the shortening undergone by the CZ in the area affected by the CZ Detachment should be at least of the same order as the length of the repetition of lower crust observed in line ESCIN-3.3, that is, around 130 km. The total shortening in these thrust sheets has not been calculated, but in the Ponga Unit and in the front of the Central Coal Basin, it amounts to approximately 93 km [*Alvarez-Marrón*, 1995]. It is quite possible that an additional shortening of at least 30-40 km had occurred to the east of the Ponga Unit and in minor imbrications inside the Central Coal Basin. Accordingly, the 130 km of duplicated lower crust could represent continental crust initially situated to the east of its present position, probably below the present Somiedo and Sobia Units, which would have been subducted to the west as a consequence of the Variscan shortening (Figure 10a).

Fourth, if a Variscan age is going to be inferred, we would need to account for the reasons why this part of the crust has been preserved. The existence of a Variscan root would be in conflict with widely accepted models of crustal evolution, according to which late to postorogenic crustal reequilibration processes lead to the elimination of any root and thick crust and to the development of a new Mohorovicic discontinuity [*Meissner*, 1989; *Bois and ECORS Scientific Party*, 1991; *Nelson*, 1992]. As a matter of fact, the profile ESCIN-3.2, immediately to the west, does not exhibit any repetition in the lower crust [*Ayarza*, 1995], and the Moho occurs between 9 and 10 s [*Alvarez-Marrón et al.*, 1996], roughly equivalent to 27-30 km.

Most of the WALZ apparently never reached an extraordinary crustal thickness, and the temperatures were never high enough to give rise to widespread melting processes and a total reequilibration of the crust with the establishment of a new crust-mantle interface. The shortage of granites in this zone, by comparison with the more internal CIZ, might be evidence for the lack of generalized synorogenic to postorogenic melting processes in the WALZ. However, some reworking of the lower crust seems to have taken place, as suggested by the occurrence in this area of postorogenic granitoids of lower crustal provenance with mantle participation [*Capdevila et al.*, 1973]. The possibility arises that its duplication was caused by a delamination process, which in this zone would have been incipient or aborted after a slight sinking of the lowermost crust and mantle lithosphere (Figure 10b).

The alternative possibility is that of an Alpine imbrication. Even though Archean to Paleozoic features are still preserved in moderately stable crusts [*Thouvenot et al.*, 1995; *Carbonell et al.*, 1996; *Knapp et al.*, 1996; *Echtler et al.*, 1996, *Clowes et al.*, 1996, *BABEL Working Group*, 1991], the case under consideration has had a complicated post-Variscan evolution.

If we consider how the Iberian lower crust has been affected in neighboring areas by the Alpine compression [*Fernández Viejo*, 1997], an Alpine root seems to be another likely hypothesis. The specific orientation of the line ESCIN-3.3 does not allow us to study Alpine reworking. However, the influence of the Alpine deformation in the Variscan crust appears to be far more important than previously thought. All the N-S ESCIN lines show how the crust has been imbricated as part of the Alpine cycle. The subduction of the Iberia underneath Europe, imaged by Etude Continentale et Océánique par Reflection et Refraction Sismique (ECORS) in the Pyrenees [*Choukroune and ECORS Team*, 1989], may be prolonged at least as far west as the Cantabrian Mountains, where the lower continental crust of the Duero Basin has been detached from the rest and imbricated underneath the Cantabrian crust [*Pulgar* et al., 1995; *Fernández Viejo*, 1997]. The downgoing slab reaches a depth of 50 km by the shoreline [*Fernández Viejo*, 1997]. A western equivalent of this slab may occur at a depth between 35 and 40 km in the area imaged by ESCIN-3.3 (Figure 10c).

7. Conclusions

The marine seismic line ESCIN-3.3 has provided wide-angle and vertical incidence data to study the crust in the NW Iberian Massif. On the basis of the first set of data, a velocity/depth model has been constructed for this part of the crust in the northern Iberian platform, which is similar to previous models based on on-land nearby refraction experiments regarding to number of layers and their velocities. The interfaces defined on the basis of wide-angle data show a close relationship with the reflectivity intervals found in the vertical incidence data (Figure 9).

Apart from the identification of two sedimentary basins in the shallow part of the seismic section, the main results concern the Variscan structure of the crust. Many of the reflections in the vertical incidence profile have been related to known Variscan geologic structures. A picture of the tectonic imbrication, already well characterized in the upper and middle parts of the crust from on-land geology, has been obtained by vertical incidence data. Two important results are the supporting evidence for the existence of a sole thrust to the West Asturian Leonese Zone running through the whole profile (Figures 7 and 9) and the interpretation of a local set of strong reflections at 6 s (TWTT) as a lower crustal slice emplaced at an anomalous shallow level during the Variscan compression.

Also important is the identification of a likely duplication of the lower crust in an area where post-Variscan reequilibration might not have been generalized but which has been reworked by the Alpine compression. The orientation of the profile subparallel to the Alpine structures hampers the identification of the latter, and no conclusive evidence has been found concerning the age of the described crustal root. Three different hypotheses have been considered: a preserved crustal-scale Variscan underthrusting, an arrested delamination affecting a thick Variscan crust, and an Alpine imbrication (Figure 10).

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