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The crustal structure of the Guayana Shield, Venezuela, from seismic refraction and gravity data

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Abstract

We present results from a seismic refraction experiment on the northern margin of the Guayana Shield performed during June 1998, along nine profiles of up to 320 km length, using the daily blasts of the Cerro Bolívar mines as energy source, as well as from gravimetric measurements. Clear Moho arrivals can be observed on the main E–W profile on the shield, whereas the profiles entering the Oriental Basin to the north are more noisy. The crustal thickness of the shield is unusually high with up to 46 km on the Archean segment in the west and 43 km on the Proterozoic segment in the east. A 20 km thick upper crust with P-wave velocities between 6.0 and 6.3 km/s can be separated from a lower crust with velocities ranging from 6.5 to 7.2 km/s. A lower crustal low velocity zone with a velocity reduction to 6.3 km/s is observed between 25 and 25 km depth. The average crustal velocity is 6.5 km/s. The changes in the Bouguer Anomaly, positive (30 mGal) in the west and negative (– 20 mGal) in the east, cannot be explained by the observed seismic crustal features alone. Lateral variations in the crust or in the upper mantle must be responsible for these observations.

Keywords: Seismic refraction; Crustal structure; Moho; Gravity; Guayana shield

1. Introduction

The Guayana Shield, which forms part of the Amazonian Craton in western South America, is an elevated plain consisting mainly of Precambrian rocks up to 3.6 billion years (Ga) and an average height of 1200 m with high plains (“Tepuys”) reaching up to 3000 m high. Gibbs and Barron (1993) indicate that this old shield region, or at least part of it, should have elevated during Mid-Tertiary, possibly as a consequence of a hot-spot or a thermal anomaly.

Due to its potential in minerals, there has been lots of geophysical and geological data collected in the Guayana Shield (e.g. Gibbs and Barron, 1993). The Venezuelan portion of the Guayana Shield is composed mainly of Archean and Proterozoic rocks (Mendoza, 1977), which can be divided into four provinces (Fig. 1). (a) Imataca, covering an area parallel to the Orinoco river, from the Caura Front in the west to the Orinoco delta in the east. It is composed mainly by Archean felsic to mafic gneisses,

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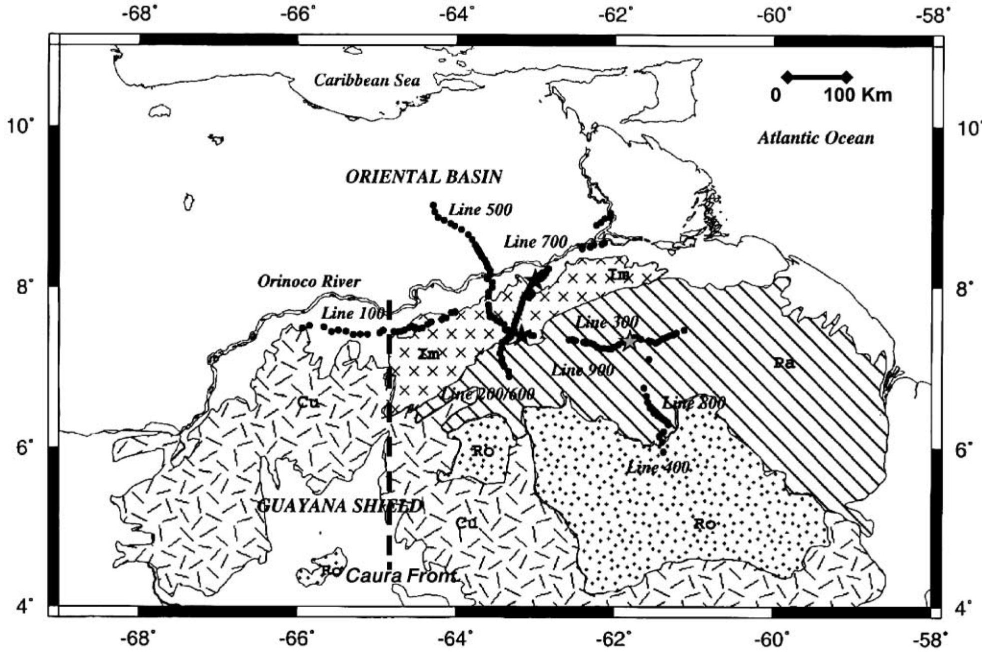


Fig. 1. Location map indicating the position of the seismic lines. The Caura Front divides the Imataca province from the Cuchivero province to the west. Stars = shot points; circles = recording points. Geological provinces of the Guayana Shield after Gibbs and Barron (1993). Im = Imataca, Pa = Pastora, Cu = Cuchivero, Ro = Roraima.

granulites and granitic intrusions accompanied with ferrugineous quartzites. The radiometric ages vary between 3.7 and 3.4 Ga (Montgomery et al., 1977).

(b) Pastora, located south of Imataca, represents a Lower Proterozoic greenstone belt with a sequence of acid and basic volcanic and sedimentary rocks of low

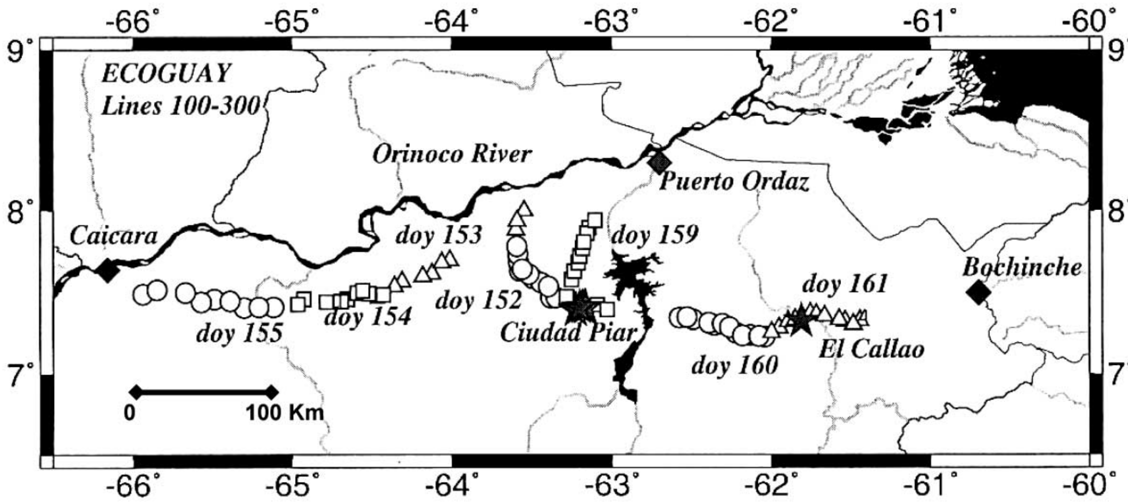


Fig. 2. Location map of record section line 100 (Ciudad Piar—west) and 300 (Ciudad Piar—east). The first 50 km of line 300 correspond to stations from line 700, and between 50 and 200 km signals are from line 300 (see Fig. 1). The sections are composed of records of 4 (doy = day of year, 152 – 155 for line 100) and 3 (doy 159 – 161 for line 300) different days, respectively, represented by different symbols for the recording sites.

grade metamorphism with ages between 2.3 and 1.9 Ga (Gibbs and Barron, 1983). (c) Cuchivero, reaching from the Orinoco river in the west to the Venezuela/Brazil border in the southeast, represents a set of Mid-Proterozoic acid plutonites as well as recrystallized and foliated volcanic belts of quartz–lateritic to rhyolitic composition, intruded by fine grained to coarse biotite–granites with ages ranging from 1.9 to 1.75 Ga (Gibbs and Barron, 1983). Cooling ages for the youngest (Parguaza) batholith of 1.2–1.5 Ga are reported (Mirón and Costanzo, 1997; Barrios, 1983; Martín, 1978). (d) Roraima, located between Pastora and Cuchivero, consists of Mid-Proterozoic sedimentary rocks (arenites, lutites and conglomerates), intensely intruded by diabasic sills and diques with ages between 1.8 and 1.7 Ga (Teixeira et al., 1989).

Earlier seismic refraction studies in Venezuela have been carried out east of Maracaibo Lake in northwestern Venezuela in order to determine the crustal structure for more precise earthquake location in the region (Gajardo et al., 1986; Rivas et al., 1988). A complex crustal structure with velocities in the crust ranging between 2.8 and 5.4 km/s for the sedimentary cover, 6.1 km/s for the crystalline crust and 6.7 km for the deeper crust with a Moho depth varying between 33 and 43 km resulted, although arrivals from the deeper crust and Moho are poor. The principal objective of the ECOGUAY project is to contribute data on the crustal structure and thickness, in order to better understand the geological evolution of the craton in northeastern South America. We present results of a combined geophysical study integrating seismic refraction and gravimetric data.

2. Seismic data

2.1. Data acquisition

Field work was done within the scope of the ECOGUAY project (Estudios de la Estructura Cortical del Escudo de Guayana) in June 1998. We obtained a total of nine seismic refraction lines of up to 320 km length (Fig. 1), using the blasts of the iron mines operated by the company C.V.G Ferrominera Orinoco in Ciudad Piar ("Cerro Bolívar") as energy source with charges between 0.14 and 33 t explosives, as well as the Cantarrana quarry southwest of Puerto Ordaz and the La Unión and Colombia gold mines in El Callao for reversed profiles with charges between 0.1 and 10 t explosives (for shot list, see Schmitz et al., 1999). The seismic data show a good signal to noise ratio up to the end of the profiles for the Ciudad Piar shot points. The profiles that extended northwards into the Eastern Venezuelan Basin are more noisy due to the Quaternary sediments north of the Orinoco river. The records of the reversed profiles only show concise energy up to 70 km.

The seismic sections were obtained by "scrolling" with the available 13 seismological stations (11 PDAS-100 and 2 ORION, all equipped with Mark 3-D, 1-Hz seismometer and high precise quartz clocks, synchronized by GPS) along the profile. Thus, the seismic sections are composed by up to four different shots with distances of the recording points varying between 5 and 10 km.

2.2. Data processing

The origin times of the mine blasts were recorded with a digital seismograph installed close to the sources. As each profile is composed of a various mine blasts (see Fig. 2), exact determination of shot times was crucial for consistency of the seismic data along the profiles. In some cases, foreshots had been applied to the shot gathers, which generally extended some 300 m. The characteristics of each blast, which had a duration of up to 1.5 s with delays between holes and lines, have been analyzed in order to determine the exact time of the main seismic impulse. The seismic data, which were digitized in the field with 100 sps, were evaluated in the field-center using the PDIS program (Grunewald, personal communication) in order to check the quality of each shot point. The data were then processed using the Seismic Unix software package (Cohen and Stockwell, 1994) generating seismic sections with 6.0 and 3.46 km/s reduction velocity, applying bandpass filters with corner frequencies of 2 and 16 Hz. For P-waves as well as for S-waves, only vertical components are plotted, without static corrections.

After identifying the main crustal phases, 1-D modeling was realized using the GWBASIC routines (Giese, personal communication) based on the Herglotz–Wiechert inversion (Giese, 1976). The results of the 1-D modeling were used as input for the 2-D

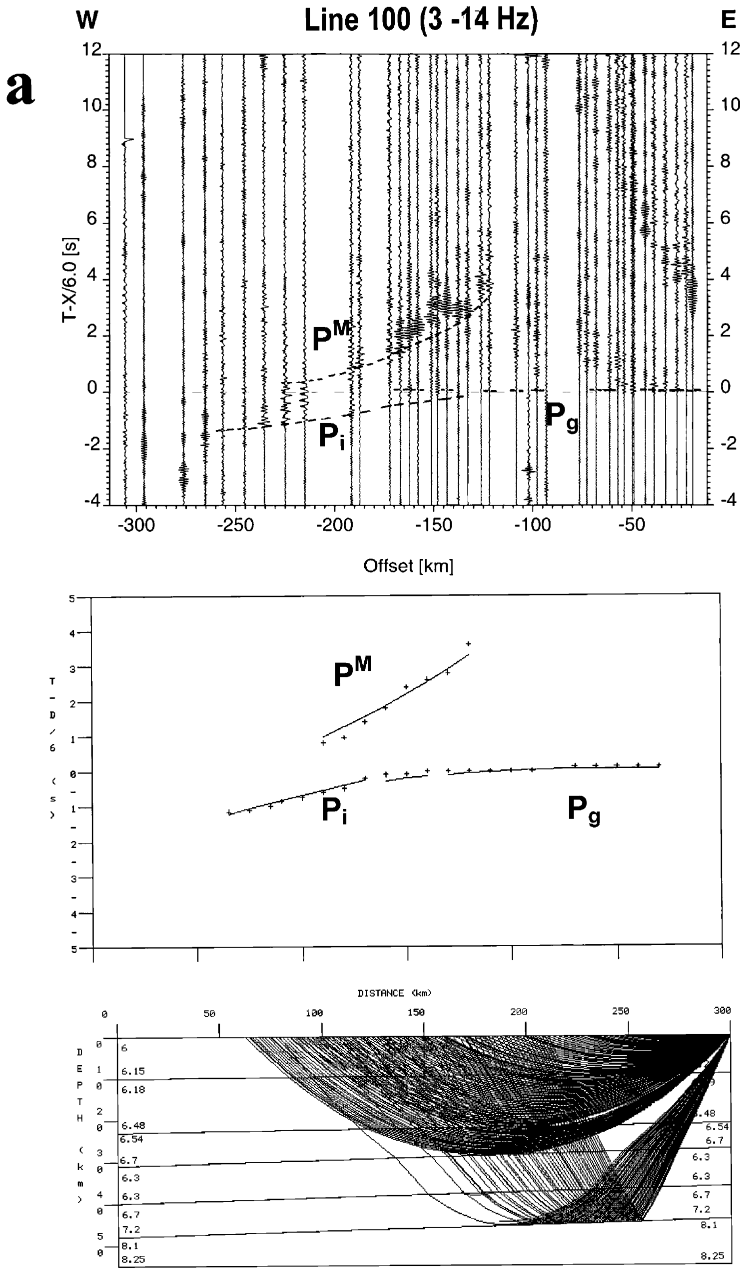


Fig. 3. Record section line 100 (Ciudad Piar—west) (a) and line 300 (Ciudad Piar—east) (b) with correlated travel time branches (top), correlated (dots) and calculated (lines) travel time branches (center) and calculated ray paths and velocity models (bottom). Basement arrivals (P_g), Moho reflections (P^M) from 44 to 46 km depth (line 100) and from 42 to 43 km depth (line 300) and first arrivals from an intracrustal phase (P_I) can be clearly identified on both sections. On the record section line 300 (Fig. 5b), reflections of an intracrustal phase (P^I) can be identified as well.

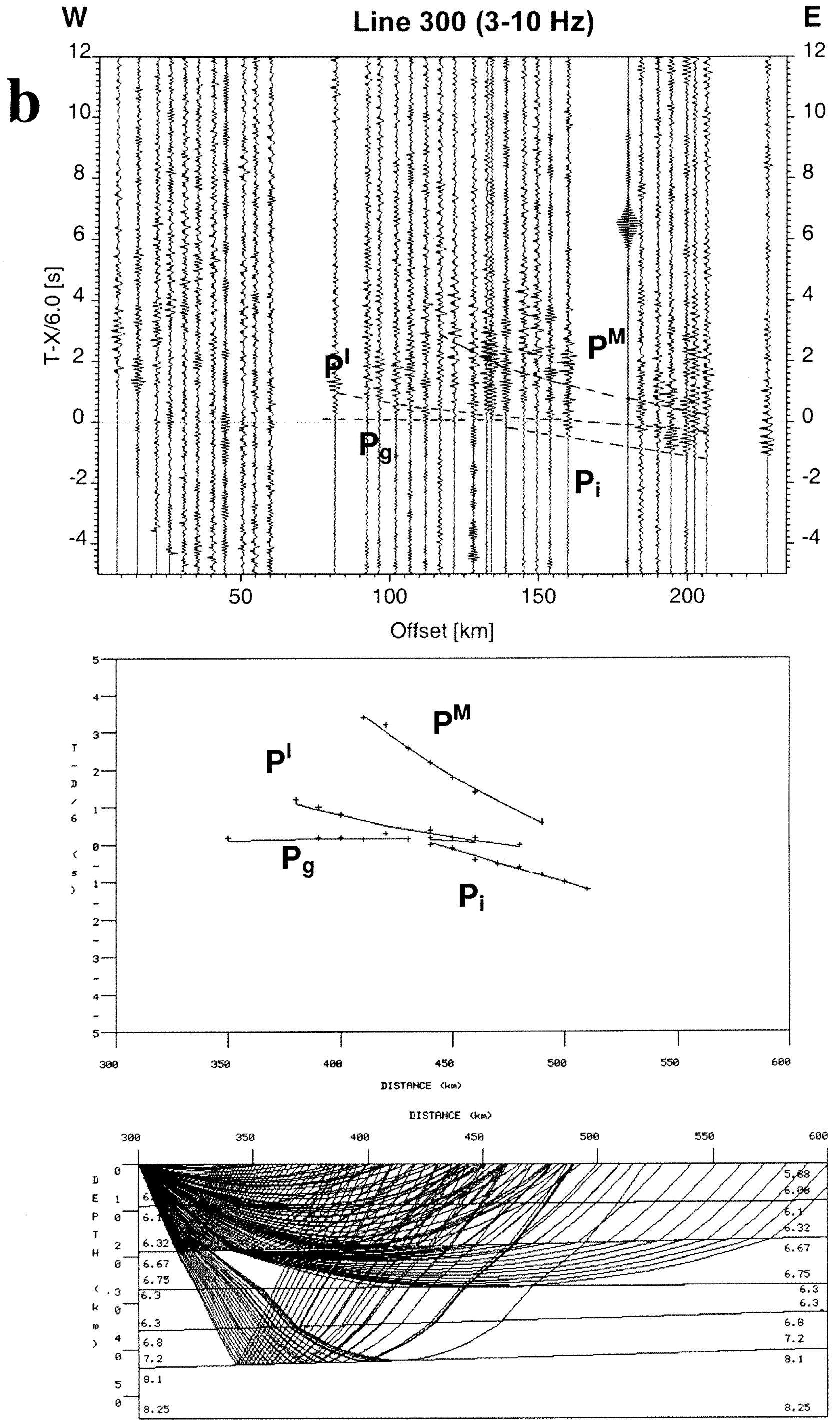


Fig. 3 (*continued*).

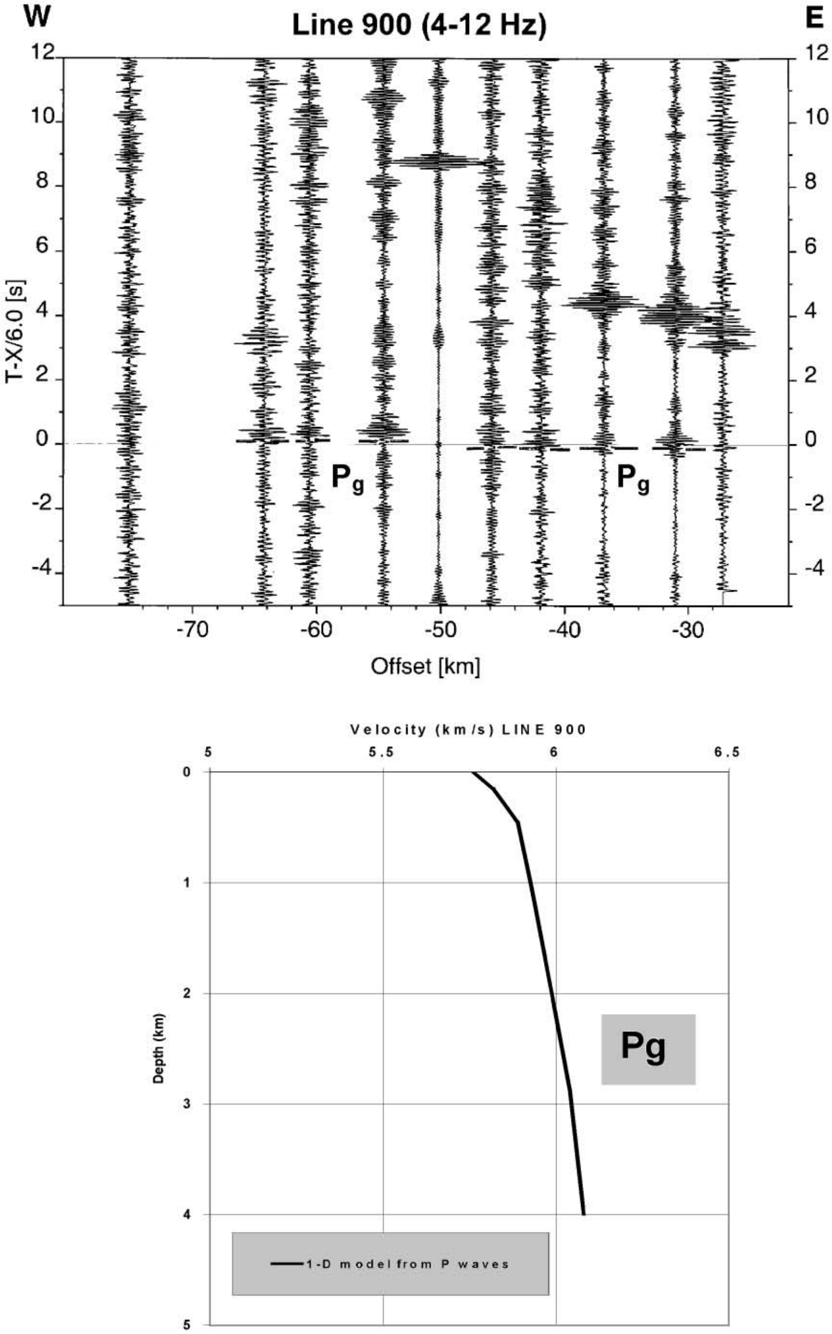
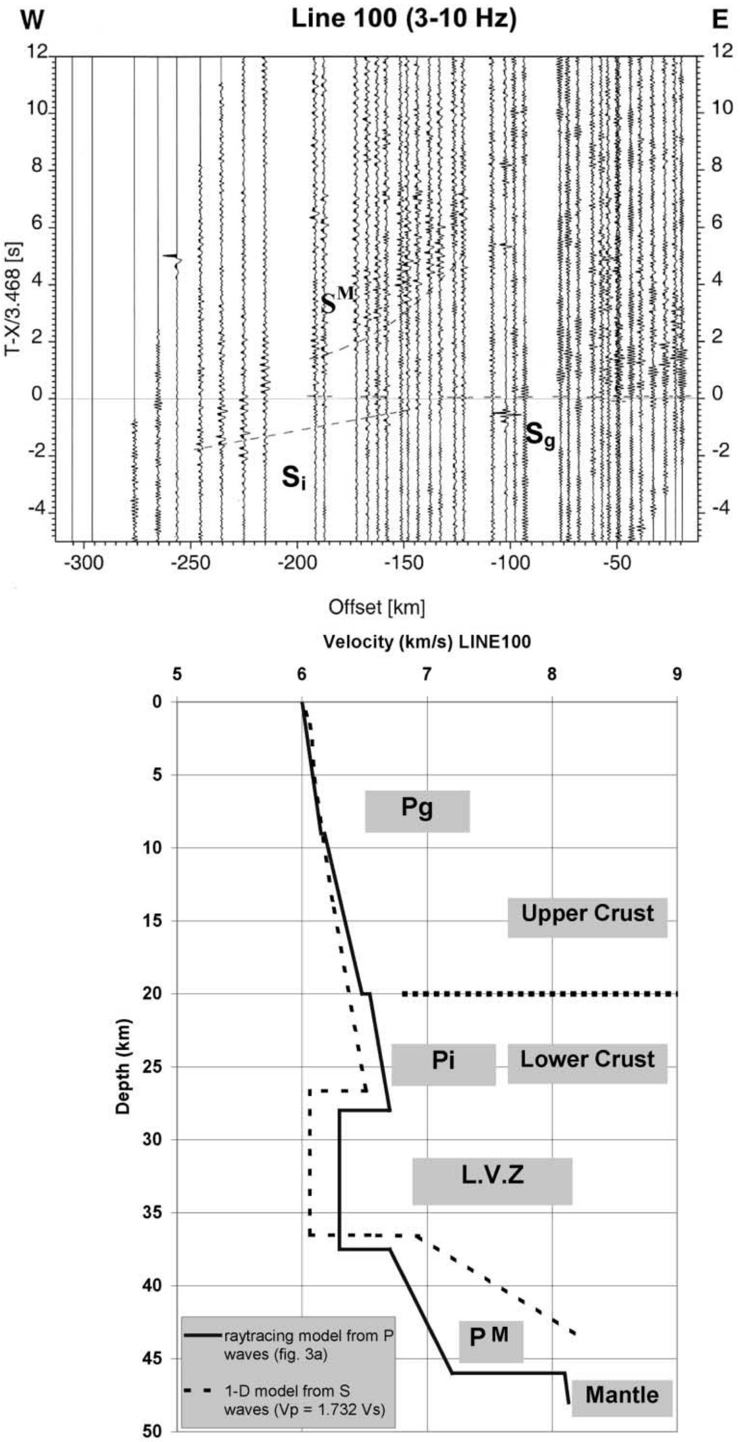


Fig. 4. Record section line 900 (El Callao—west) with correlated travel time branches and the velocity–depth function derived from 1-D modeling. Only the basement arrivals (P_g) could be identified, with the velocity increasing to 6.1 km/s at about 4 km depth.

a


W
Line 100 (3-10 Hz)
E
T-X/3.468 [s]
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S_g
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Velocity (km/s) LINE100
Depth (km)
Pg
Upper Crust
Pi
Lower Crust
L.V.Z
P M
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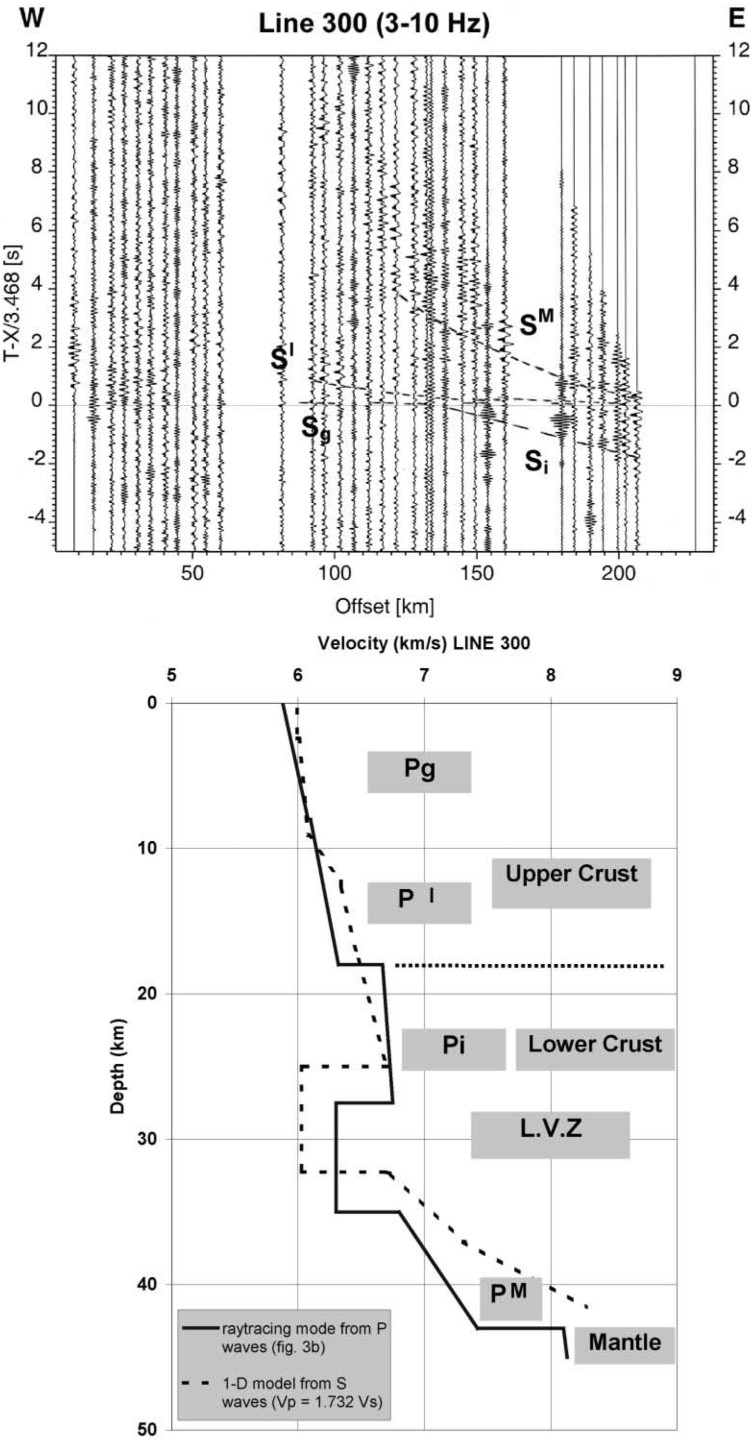
raytracing model from P waves (fig. 3a)
1-D model from S waves (Vp = 1.732 Vs)

b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Depth (km)
Pg
Upper Crust
P I
Pi
Lower Crust
L.V.Z
P M
Mantle
raytracing mode from P waves (fig. 3b)
1-D model from S waves (Vp = 1.732 Vs)

modeling, which was done using the ray tracing program RAYINVR (Zelt, 1992).

2.3. Interpretation

The best seismic data could be obtained along the west–east profile located entirely on the Precambrian formations of the shield, ranging from Caicara in the west to Bochinche in the east with the shot point (Ciudad Piar) in the center (Fig. 2). In the seismic section of profile 100, recorded to the west, first arrivals can be distinguished with a good signal/noise ratio up to 250 km distance (Fig. 3a). The first arrivals of the basement (P_g) are observed at a reduced time of 0 s up to 175 km distance. P-velocities increase from 6.0 km/s at the surface to 6.4 km/s at about 20 km depth. An intracrustal phase (P_i), observed between 250 and 140 km with early arrivals (− 2 to 0 s) and a velocity of up to 6.7 km/s, extends down to 30 km depth, followed by a low-velocity zone with a velocity reduction to about 6.3 km/s. The reflections from the upper mantle (P^M) are observed at 190–120 km distance. The crust–mantle boundary (Moho discontinuity) is calculated to 44–46 km depth with an upper mantle velocity of 8.1 km/s and an average crustal velocity of 6.5 km/s.

From the same shot point recorded towards the east (line 300: Ciudad Piar-east, see Figs. 2 and 3b), first arrivals can be identified up to 225 km distance. The P_g arrivals are observed up to 140 km distance with a velocity of 6.3 km/s, followed by an intracrustal reflection (P^I) between 210 and 80 km distance (0–1 s), resulting in a depth of 18 km. First arrivals between 140 and 210 km distance (0 to − 1 s) are observed from an intracrustal phase at about 18–27 km depth (P_i). As well as along line 100, this phase is followed by a low-velocity zone in the lower crust. The Moho reflection is observed between 210 and 120 km distance at 0–2.5 s, calculated to a depth of 41–43 km. Average crustal velocity is 6.5 km/s. From the reversed shot point El Callao—west, records could be obtained up to 80 km distance only (Fig. 4). The velocity structure of the uppermost crust could be determined down to 4 km depth with a velocity of 6.1 km/s.

In the record sections of lines 100 and 300 with a reduction velocity of 3.46 km/s, the S_g, S_i and the S^M phases, as well as a S^I phase on line 300, could be clearly identified (Fig. 5). The most striking feature in the S-wave sections is the intracrustal phase S_i, which represents the top of the lower crust. A pronounced low-velocity zone results for the lower crust, which is mainly due to the high velocities obtained for the intracrustal phase S_i.

For the 1-D and 2-D model calculations displayed in Fig. 5, the total crustal thickness varies between 43 and 46 km for line 100 and 41–43 km for line 300. Furthermore, estimations of the overall crustal thickness along both seismic lines were done using the formula for maximum depth calculation (Giese, 1976). Considering the upper mantle velocity to vary between 7.9 and 8.2 km/s, crustal thickness of line 100 varies between 42 and 46 km (critical distance of P^M reflection between 120 and 130 km), and for line 300 crustal thickness varies between 40 and 43 km (critical distance 120–125 km). The average crustal velocity for these observations varies between 6.4 and 6.6 km/s.

The signal-to-noise ratio decreases significantly towards the north due to the attenuation of the seismic energy caused by the sediments of the Oriental Basin, north of the Orinoco river and possible variations in the seismic energy released by the different mine blasts. On line 500 (Ciudad Piar—northwest), the P_g may be observed up to 175 km with a slight delay of the first arrivals at distances of more than 140 km due to the sedimentary cover (Fig. 6). The resulting velocity at 20 km depth is 6.0 km/s. There are some indications for a P^M phase between 170 and 100 km distance, which result in a crustal thickness of about 43 km. There are no indications for an intracrustal phase in this record section, as observed further south on the lines 100 and 300 (Fig. 3).

The signal-to-noise ratio at line 700 is significantly better, but unfortunately the distance range 95–145 km, where the critical reflections from the Moho are

Fig. 5. (previous pages) Record sections line 100 (Ciudad Piar—west) (a) and line 300 (Ciudad Piar—east) (b) with S-wave travel times from 1-D model calculations (top) and the velocity–depth functions derived from 1-D (S-waves) and 2-D (P-waves) model calculations (bottom). The function derived from the S-wave 1-D models are plotted based on the correlation $V_P = 1.732 \times V_S$. Total crustal thickness varies between 43 and 46 km for line 100 and between 41 and 43 km for line 300.

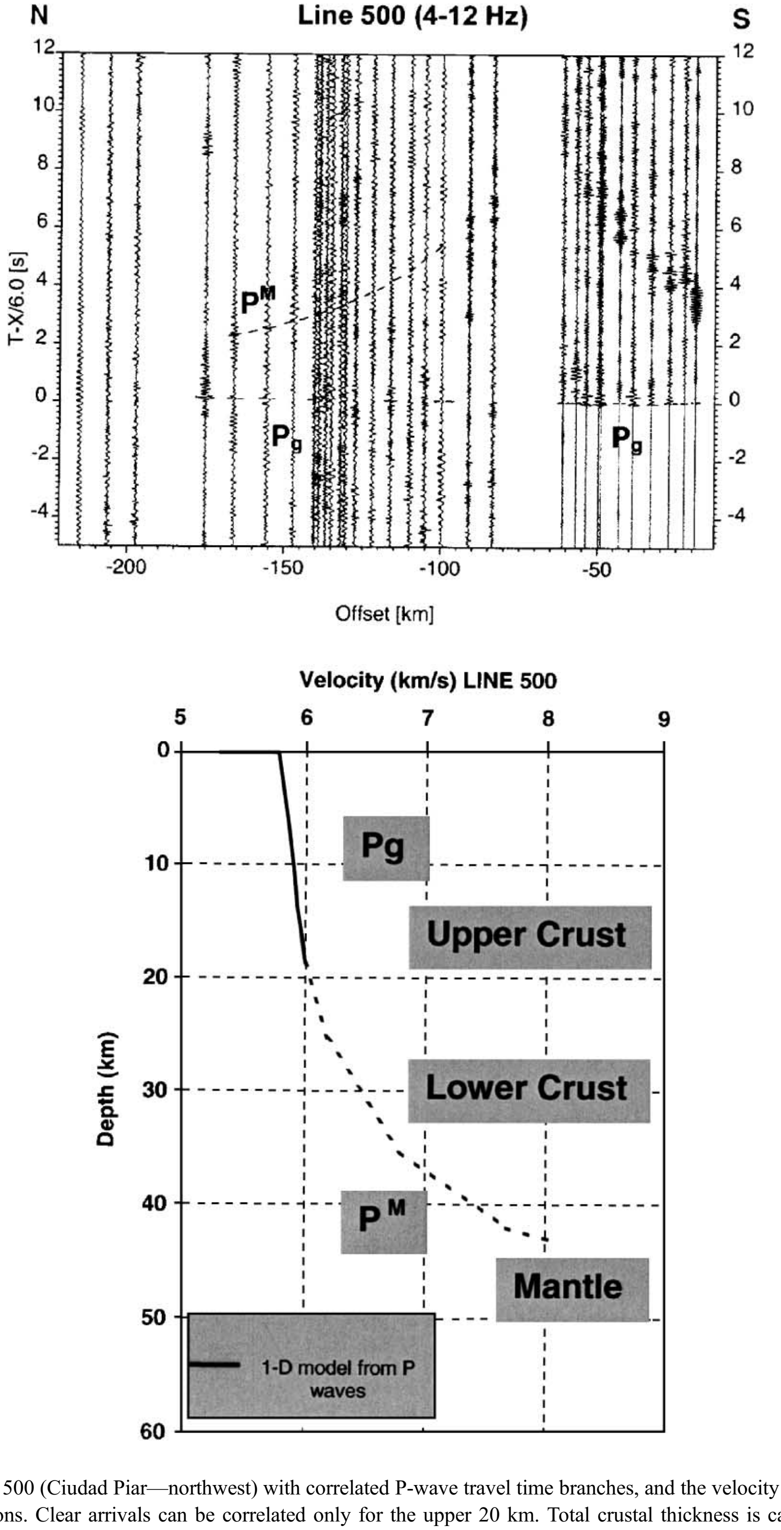

Fig. 6. Record section line 500 (Ciudad Piar—northwest) with correlated P-wave travel time branches, and the velocity–depth function derived from 1-D model calculations. Clear arrivals can be correlated only for the upper 20 km. Total crustal thickness is calculated from weak P^M arrivals to 43 km depth.

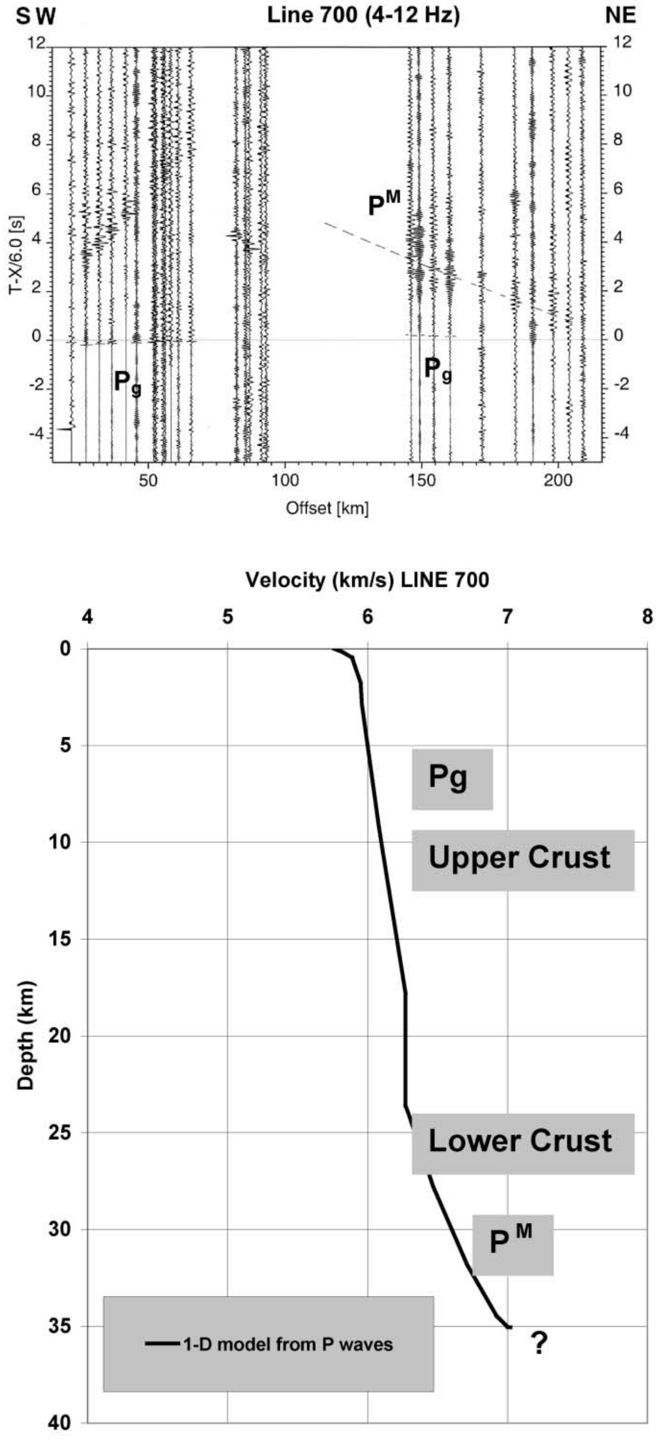

Fig. 7. Record section line 700 (Ciudad Piar—northeast) with correlated P-wave travel time branches, and the velocity–depth function derived from 1-D model calculations. The total crustal thickness cannot be calculated, as records are missing in the critical distance of P^M arrivals.

expected, is not covered by seismic signals (Fig. 7). The P_g can be observed up to 160 km with a delay of 0.2 s and the P^M phase between 200 and 145 km distance, without possibility to determine the critical distance and therefore the crustal thickness. Down to about 25 km depth, the velocity does not exceed 6.3 km/s and a velocity of 7.0 km/s is reached at 35 km depth. As on line 500 (Fig. 6), there are no indications for an intracrustal phase in this record section.

3. Crustal model

The structural information obtained during the ECOGUAY project is summarized along a profile covering the northern part of the Guayana Shield (Fig. 8). Along the 600 km long west–east profile from Caicara (at the Orinoco river) in the west to Bochinche at the Venezuela–Guyana border, an upper crust with velocities between 6.0 and 6.3 km/s and a depth of about 20 km can be distinguished from a lower crust with P-wave velocities of 6.5–7.2 km/s down to the Moho, which is slightly inclined from 42 km depth in the east to 46 km depth in the west. The lower crust is characterized by a high velocity layer on the top, clearly identified in the record sections of lines 100 and 300 with a velocity of 6.5–6.7 km/s, underlain by a 10 km thick low-velocity zone and then again an increase in velocity down to the Moho. There are no indications for anomalous upper mantle velocities, which are considered to range between 8.1 and 8.2 km/s. The average crustal velocity is calculated to about 6.5 km/s. No significant changes in crustal velocities can be derived for the Arquean Imataca province in the west and the Proterozoic Pastora province in the east. Nevertheless, total crustal thickness varies slightly between both units. The Caura front, proposed by Mendoza (1977) as a suture zone between Precambrian plates (see Fig. 1), could not be identified in the seismic data, as this zone was not directly covered by vertices of the seismic rays.

In north–south direction towards the center of the Oriental Basin in the north, the crustal structure is only poorly constrained by the information from lines 500 and 700. The upper crust with an average velocity of 6.2 km/s in the shield region is supposed to thin out towards the north. Here, sedimentary thickness in the Oriental Basin increases to about 6 km at El Tigre (González de Juana et al., 1980) with Tertiary and Cretaceous sediments. Paleozoic sediments (Feo Codecido et al., 1984) could be responsible for a delay in the P_g arrivals north of Orinoco river (Figs.

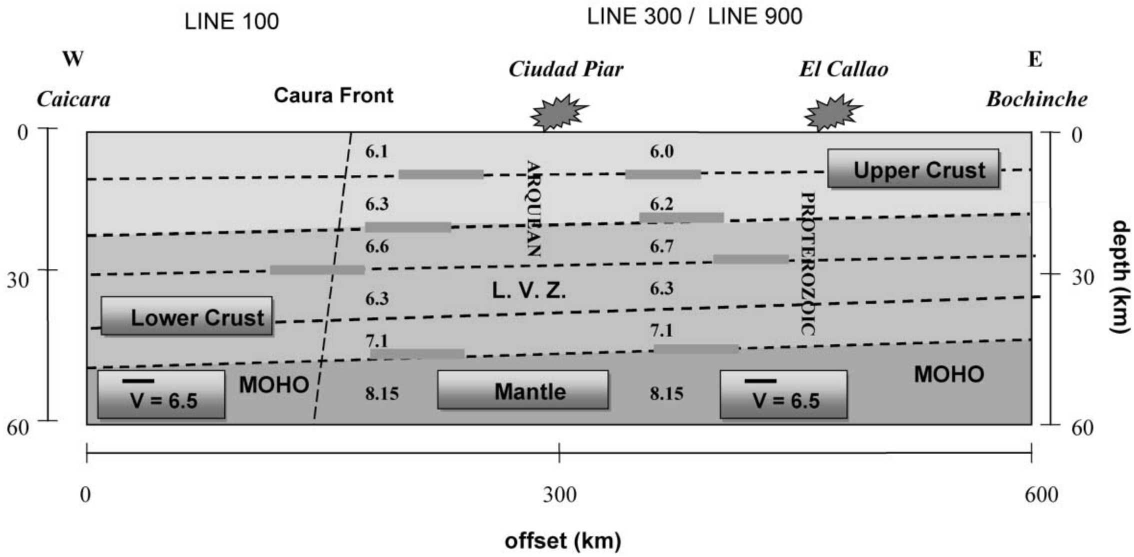

Fig. 8. E–W profile on the northern edge of the Guayana Shield with the structural information derived from the seismic lines 100, 300 and 900. Average P-wave velocity of crustal units are indicated. Areas covered by vertices of seismic rays are marked with bold lines. The Moho is characterized by a sharp boundary with a slight inclination from 42-km depth in the east to 46 km in the west.

6 and 7) with respect to the region further south. The average crustal velocity decreases towards the north. The calculated crustal thickness is about 43 km. An inclination towards the north is assumed due to the sedimentary loads, but cannot be confirmed by the existing seismic data.

4. Gravimetric observations

About 300 new gravimetric data were obtained in order to complement the existing gravity data base at Simon Bolívar University (Venezuela) (Graterol, 1993). Processing has been done following the guidelines given by Dehlinger (1978) and Introcaso et al. (1992). The Bouguer anomaly shows variations from 30 mGal in the west to − 20 mGal in the east along the main east–west profile (Figs. 9 and 10). In the Oriental Basin in the north of the study region, the Bouguer anomaly drops from 30 mGal in the west to − 190 mGal in the east, north of the Orinoco delta (Martín, 1978).

The observed Bouguer anomaly along the east–west profile (Fig. 10) seems to be in contradiction with the simplified crustal structure derived from the seismic refraction results (see Fig. 8). The conversion from velocity to density was done using the empirical velocity–density curves of Ludwig et al. (1970). A clear discrepancy between observed and calculated Bouguer anomaly along the Guayana Shield, which

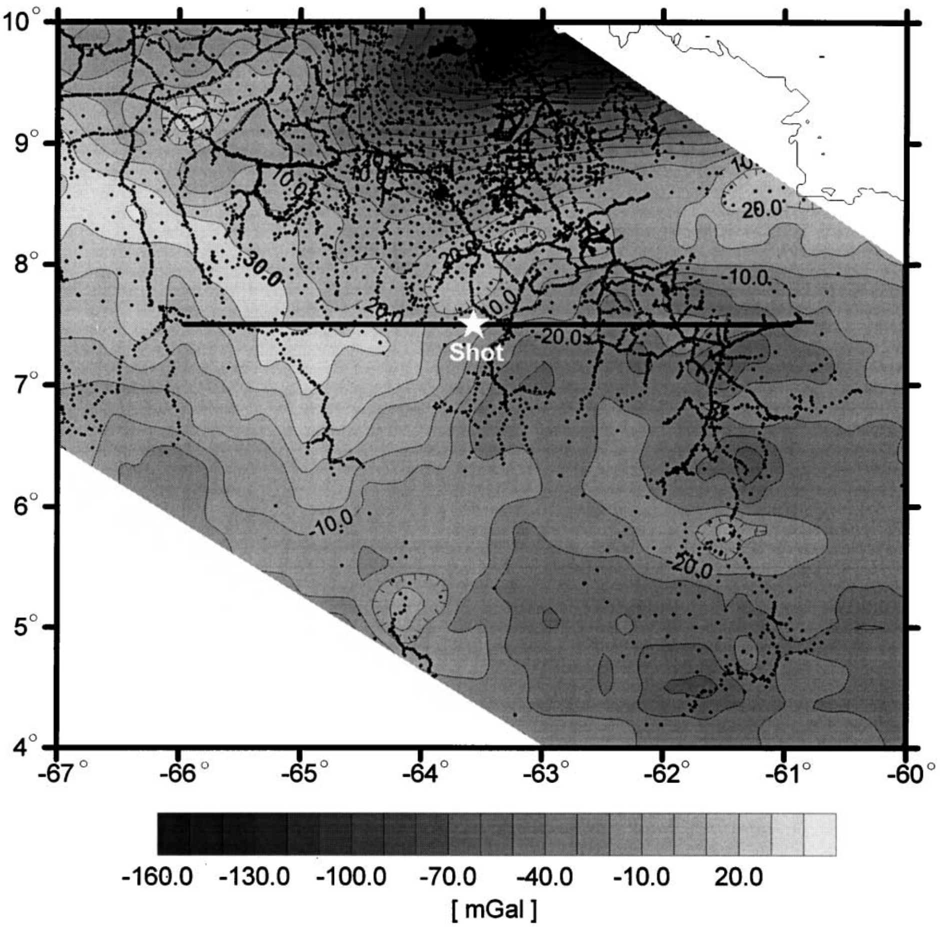

Fig. 9. Bouguer anomaly map with the location of the profile of Fig. 10 (black line) and the shot point Ciudad Piar (star) from seismic lines 100 and 300 (see Fig. 1 for location). The Bouguer anomaly varies between − 20 mGal in the east and about 30 mGal in the west.

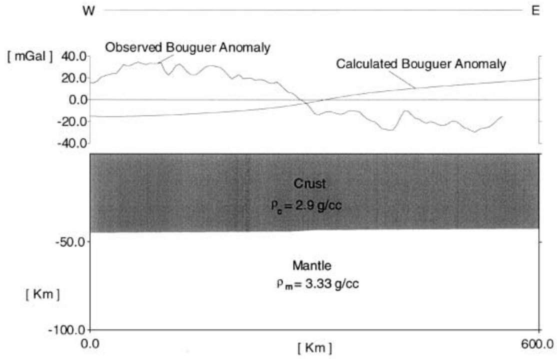

Fig. 10. Model calculations for the density structure along an east–west profile (Fig. 9) covering seismic lines 100 and 300 as derived from the velocity model (Fig. 8), with a simplified crustal structure. Observed and calculated Bouguer anomalies show an opposite behaviour for the structure derived from seismic refraction. Lateral density variations in the crust as well as lateral density variations in the mantle should be responsible for the difference in Bouguer anomaly between the eastern and the western part of the profile (Fig. 9).

increases in crustal thickness from east to west, is observed (Fig. 10).

Lateral variations at different depth levels must be responsible for the observed Bouguer anomaly, which might reflect contributions from crustal as well as from mantle levels. The prominent NW–SE striking positive anomaly (30 mGal) in the northwestern part of the study region (Fig. 9) shows a continuation towards NE with a less pronounced positive anomaly (10–20 mGal). These anomalies might be generated from crustal level, but the seismic data cannot resolve these structures.

Lateral variations in densities might be expected between a more dense Arquean crust (Imataca Province) to the west and the Proterozoic crust (Pastora Formation) to the east (Fig. 1). The Proterozoic crust in the east may have followed thermal processes different and more recent to the ones occurred to the Arquean crust in the west (Gibbs and Barron, 1993), which may have weakened the Proterozoic crust.

A more regional contribution might be generated from mantle level, indicated by a regional gravity low of − 20 to − 30 mGal in the southern part of the study region, more pronounced to the east (Fig. 9). One of the possible options to explain the general trend might be relicts of subducted oceanic crust at depths of about 120–150 km in the western region of the Guayana Shield (Sobolev et al., 1998). The processes responsible for the generation of this basaltic environment could have thickened the crust in that region and modified the upper mantle. Durrheim and Mooney (1994) suggest for Archean cratons a less dense mantle than the surrounding asthenosphere during lithospheric evolution. The above mentioned features (at crustal and mantle scale) might explain the discrepancy between the observed Bouguer anomaly and the simplified velocity structure along the E–W profile (Fig 10).

5. Discussion and conclusions

In order to determine the crustal structure of the northern part of the Guayana Shield, 1-D traveltime inversion as well as forward raytracing modeling have been applied to the wide-angle seismic sections obtained during the ECOGUAY project. In the shield region, upper crustal velocity varies between 6.0 and 6.3 km/s down to 20-km depth. Lower crustal velocities range between 6.5 and 7.0 km/s with a total crustal thickness of 42–46 km, slightly inclined to the west, and an average velocity of 6.5 km/s. The northwestern part of the Guayana Shield is composed of Arquean crust (Imataca Province), while the Cuchivero Formation and Pastora Formation, both Proterozoic, are exposed to the south and to the east, respectively (Fig. 1). The aborted triple junction of the Takutu graben southeast of the Guayana Shield (Gibbs and Barron, 1993) could have been generated by a mantle plume during Paleozoic time. The thinner crust to the east could be a relict of that time and the thicker crust towards the west might be influenced by the sedimentary load of the Oriental Basin. Velocities decrease towards the Oriental Basin north of the cratonic region.

The obtained crustal thickness is significantly higher than the ones reported for a similar tectonic environment as the Archean Kaapvaal Craton with 35–40 km (Durrheim and Green, 1992) or the Precambrian Churchill and Superior Groups with 40–43 km (Delandro and Moon, 1982). Pronounced seismic roots were discovered recently for Precambrian shield regions, as the Torngat Orogen (Funck and Louden, 1999) and the Baltic Shield (BABEL Working Group, 1993), with 50 and 60 km depth, respectively. The average crustal thickness for shield regions, derived from seismic refraction studies, is reported to about 41 km with an average velocity of 6.5 km/s (Mooney et al., 1998). Whereas the crustal thickness obtained for the Guayana Shield is on the upper edge of the worldwide observed data, the velocity fits well to the observed ranges. The thickness of the upper crust, without sedimentary cover, is slightly higher than the worldwide average. A distinction between middle and lower crust is difficult because the observed intracrustal phase from 20–30-km depth (see Fig. 5) does not allow the direct observation of deeper crustal phases. The underlying low velocity zone in the depth range of about 25 to 35 km might be explained by differentiation processes in the crust during the Arquean and Proterozoic magmatic cycles. Mafic gneisses in the Imataca province and the amphibolite facies greenstone belts in the Pastora province (Cox et al., 1993) might be the surface expression of the high velocity crustal phases. The anorthositic Nain Plutonic Suite in Labrador, Canada, deeply exposed by erosion of some 10 km, shows a similar feature with high seismic velocities up to 6.8 km/s at about 10 km depth underlain by lower seismic velocities (Funck et al., 2000). The buoyant rise of anorthositic magmas to midcrustal levels may have created the high velocity bodies with underlying low velocity zones in the Guayana Shield, as proposed by Funck et al. (2000) for the Nain Plutonic Suite.

The observed lateral uniformity of the crust with a slight inclination to the west is in contradiction to the Bouguer anomaly, which shows positive values of about 30 mGal in the western part of the study region and negative values of about -20 mGal in the east (Figs. 9 and 10). Lateral variations within the crust and the upper mantle should be responsible for this misfit. Candidates to explain these lateral variations are density variations between the Arquean and the Proterozoic crust as well as relicts of subducted oceanic crust beneath the western part of the Guayana Shield or density differences between Archean and Proterozoic asthenosphere.

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