Crustal deformation in the south-central Andes backarc terranes as viewed from regional broad-band seismic waveform modelling

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SUMMARY
The convergence between the Nazca and South America tectonic plates generates a seismically active backarc region near 31°S. Earthquake locations define the subhorizontal subducted oceanic Nazca plate at depths of 90–120 km. Another seismic region is located within the continental upper plate with events at depths <35 km. This seismicity is related to the Precordillera and Sierras Pampeanas and is responsible for the large earthquakes that have caused major human and economic losses in Argentina. South of 33°S, the intense shallow continental seismicity is more restricted to the main cordillera over a region where the subducted Nazca plate starts to incline more steeply, and there is an active volcanic arc. We operated a portable broad-band seismic network as part of the Chile–Argentina Geophysical Experiment (CHARGE) from 2000 December to 2002 May. We have studied crustal earthquakes that occurred in the back arc and under the main cordillera in the south-central Andes (29°S–36°S) recorded by the CHARGE network. We obtained the focal mechanisms and source depths for 27 (3.5 < Mw < 5.3) crustal earthquakes using a moment tensor inversion method. Our results indicate mainly reverse focal mechanism solutions in the region during the CHARGE recording period. 88 per cent of the earthquakes are located north of 33°S and at middle-to-lower crustal depths. The region around San Juan, located in the western Sierras Pampeanas, over the flat-slab segment is dominated by reverse and thrust fault-plane solutions located at an average source depth of 20 km. One moderate-sized earthquake (event 02-117) is very likely related to the northern part of the Precordillera and the Sierras Pampeanas terrane boundary. Another event located near Mendoza at a greater depth (~26 km) (event 02-005) could also be associated with the same ancient suture. We found strike-slip focal mechanisms in the eastern Sierras Pampeanas with shallower focal depths of ~5–7 km. Overall, the western part of the entire region is more seismically active than the eastern part. We postulate that this is related to the presence of different pre-Andean geological terranes. We also find evidence for different average crustal models for those terranes. Better-fitting synthetic seismograms result using a higher P-wave velocity, a smaller average S-wave velocity and a thicker crust for seismic ray paths travelling through the crust of the western Sierras Pampeanas (Vp = 6.2–6.4 km s\(^{-1}\), Vp/Vs > 1.80, th = 45–55 km) than those of the eastern Sierras Pampeanas (Vp = 6.0–6.2 km s\(^{-1}\), Vp/Vs < 1.70, th = 27–35 km). In addition, we observed an apparent distribution of reverse crustal earthquakes along the suture that connects those terranes. Finally, we estimated average P and T axes over the CHARGE period. The entire region showed P- and T-axis orientations of 275° and 90°, plunging 6° and 84°, respectively.

Key words: Andes backarc, continental crust, earthquake-source mechanism, seismotectonics, subduction zone.

1 INTRODUCTION
Crustal seismicity in Argentina and Chile between 29°S and 36°S is related to the Andean deformation (Fig. 1). The region is

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characterized by the convergence between the oceanic Nazca plate and the continental South America plate at a rate of 6.3 cm yr$^{-1}$ and an azimuth of 79.5$^\circ$ (Kendrick et al. 2003). Major along-strike changes in the Wadati-Benioff zone occur in this segment. The strike of the Nazca plate changes in dip from subhorizontal to more inclined in the vicinity of 33$^\circ$S (Barazangi & Isacks 1976; Cahill & Isacks 1992) (Fig. 1). Approximately 490 earthquakes of magnitude greater than 4.0 and focal depth <50 km have occurred in the last 20 yr in the main cordillera and backarc region between 29$^\circ$ and 36$^\circ$S. In addition, large crustal earthquakes have caused many deaths and great devastation near San Juan and Mendoza (Argentina) in 1861, 1894, 1944, 1952 and 1977 (Castellanos 1944; Grober 1944; INPRES 1977; Triep 1979; Volponi 1979; Bastias et al. 1993). There is also a record of one shallow $M_c = 6.9$ earthquake that occurred in the main cordillera causing damage in Chile in 1958 (Piderit 1961) (Fig. 2). Recently two crustal earthquakes with moment magnitudes of ~6.4 have occurred in the main Andean cordillera (PDE-NEIC location 34.93$^\circ$S and 70.39$^\circ$W, depth 16 km) and in Catamarca province (INPRES location 28.73$^\circ$S and 66.20$^\circ$W, depth 30 km) in 2004 August to September.

The backarc seismicity in the immediate eastern flank of the Andes since 1963 has been studied using teleseismic data and short-period seismic networks (Stauder 1973; Barazangi & Isacks 1976; INPRES 1977, 1985; Triep 1979, 1987; Chinn & Isacks 1983; Langer & Bollinger 1988; Pujol et al. 1991; Smalley & Isacks 1990; Regnier et al. 1992, 1994; Assumpção & Araujo 1993; Smalley et al. 1993; Langer & Hartzell 1996). More recently, the Harvard catalogue provided more than 30 focal mechanism solutions for the region since 1977 using the centroid moment tensor (CMT) method (Dziewonski et al. 1981; Dziewonski & Woodhouse 1983). A compilation of these seismic data, including all available focal mechanism solutions, can be found in Alvarado et al. (2004). However, continental seismicity that occurs in the easternmost part over the flat-slab segment and south of 33$^\circ$S of the Andean backarc is still poorly known.

Small- to moderate-sized crustal earthquakes are more abundant but difficult to study with global seismic data. Their study is important in understanding the seismotectonic processes and seismic hazards of the region. Unfortunately, these events are often not well recorded by global seismic networks. Regional broad-band portable seismic networks can help to characterize this seismicity by providing focal mechanism determinations and stress orientation estimates, improving source-depth estimates, and permitting tests of seismic velocity models for the region.

In this paper, we present a regional study of the moderate-sized crustal seismicity in the Andean backarc between 29$^\circ$S and 36$^\circ$S using broad-band seismic data collected during 18 months in the Chile–Argentina Geophysical Experiment (CHARGE). Complete three-component seismograms recorded at regional distances were used to invert for the seismic moment tensor of 27 crustal earthquakes. We also determined their focal depths. In addition, we explored average crustal seismic velocity structures for the region. These results are discussed in the context of previous geological and geophysical studies.

2 REGIONAL TECTONIC SETTING

Most of western South America between 29$^\circ$S and 36$^\circ$S is made up of pre-Carboniferous accreted terranes (Ramos et al. 1986, 2000;
Dalla Salda et al. (1992; Astini et al. 1995). Geological studies have documented how the major terrane sutures have experienced extensional and compressional deformation since the Jurassic (Ramos 1994; Rapela 2000; Ramos et al. 2002). In this seismic study we also explore what control these recognized geological terranes might have on present Andean deformation.

Fig. 2 shows the morphological and structural provinces and terrane boundaries of our study area. These provinces have developed since ~25 Ma in response to the subduction conditions that produced a progressive eastward migration of the orogenic front and the volcanic arc (Ramos 1988; Kay et al. 1991). The Principal Cordillera, which forms the main part of the Andes, is composed of Mesozoic marine deposits and the basement is involved in their deformation. The northern and southern parts are characterized by thick-skinned tectonics, while the central segment, with peaks with elevations higher than 6700 m, is characterized by thin-skinned tectonics (Ramos et al. 1996). The Frontal Cordillera includes Palaeozoic–Triassic magmatic rocks that behaved as a rigid block during the Andean deformation (Ramos et al. 1996). The Precordillera forms the foothills of the Andes between 29°S and 33°S. It is a fold and thrust belt developed on a Palaeozoic carbonate platform (Baldis et al. 1982). East of the Andes, the Sierras Pampeanas province is composed of a series of crystalline basement-cored Precambrian–Early Palaeozoic rocks uplifted and tilted by compression during the formation of the Andes and separated by broad and relatively undeformed basins (Jordan et al. 1983; Jordan 1995). These uplifts are considered a modern analogue to the Laramide uplifts of the western USA (Jordan & Allmendinger 1986). The western Sierras Pampeanas are composed principally of abundant crystalline calcites, amphiboles, basic and ultrabasic rocks, and scarce granitic bodies (Caminos 1979). In contrast, the eastern Sierras Pampeanas exhibit schists and gneisses of a mainly sedimentary origin, granitic rocks, some of which reach batholithic dimensions, and abundant migmatites and granulites (Kraemer et al. 1995; Varela et al. 2000). Proterozoic–Early Palaeozoic sutures separate the Río de la Plata, Pampia and Famatinia terranes in the eastern Sierras Pampeanas (Aceñaloza & Toselli 1976). The western Sierras Pampeanas contain rocks as old as 1.1 to 1.0 Ga (McDonough et al. 1993) that correlate in age with the Grenville orogeny. These rocks are part of the Cuyania terrane, which also includes the Precordillera (Dalla Salda et al. 1992; Ramos et al. 2000). The boundary between the Cuyania and Famatinia terranes is interpreted as a major Early Palaeozoic suture (Dalla Salda et al. 1992; Astini et al. 1995; Ramos et al. 2002). However, the origin and evolution of these terranes are still debated (Aceñaloza et al. 2002).

South of 33°S, a very active volcanic arc (Fig. 1) coincides with a decreasing age of the subducted plate and hence a higher geothermal gradient (Yañéz & Cembrano 2004). Crustal seismicity is very active in the foreland over the flat-slab segment (Smalley & Isacks 1990; Regnier et al. 1992; Smalley et al. 1993; Gutscher et al. 2000). The Precordillera and Sierras Pampeanas have produced large crustal earthquakes (Costa et al. 2001; INPRES 2005) with the most destructive having magnitude ~7.0 and occurring near San Juan in 1944 (Castellanos 1944; Groeber 1944) and 1977.
(INPRES 1977; Volponi 1979; Kadinsky-Cade 1985; Costa et al. 2000) (Fig. 2). In contrast, south of 33°S very shallow (depth <20 km) earthquakes are restricted to the high cordillera (Barrientos et al. 2004) (Figs 1 and 2).

3 DATA AND REGIONAL WAVEFORM MODELLING

The CHARGE network consisted of 22 seismic stations with 10 STS2 and 10 Guralp-esp sensors from IRIS-PASSCAL (USA), and two Guralp-40T sensors from the Instituto Nacional de Prevención Sísmica (INPRES) and the Universidad Nacional de San Juan (Argentina), respectively. They were deployed in two transects from the Chilean coast to central Argentina at about 30°S and 36°S with several stations in between (Fig. 1). Stations JUAN, USPA, HUER and LLAN were installed coincident with INPRES sites. The rest of the instruments were deployed inside a temporary vault located on hard-rock sites. The instruments operated continuously recording at 40 samples per second for 1.5 yr (2000 December–2002 May). In that period, we recorded 75 shallow earthquakes having body wave magnitudes $M\geq4.0$ including four events of $5.0<M<6.0$, with locations in the backarc and main Andean cordillera region (Fig. 1).

We performed a linear least-squares seismic moment tensor inversion (SMTI) for 27 crustal earthquakes that were well recorded using at least four CHARGE stations. We used available global data from station PEL (Geoscope) in Chile to study three events located near Mendoza. We also determined the best focal depth. We will refer to the events by the year and Julian day of occurrence. The SMTI technique (Randall et al. 1995) consists of modelling the complete three-component seismic-displacement records at regional distances to obtain the seismic moment tensor for a point source. We considered epicentral distances up to 727 km.

Accurate seismogram modelling depends upon reliable seismic locations and a seismic velocity structure. We used the epicentre information determined by the Argentine seismic network (INPRES) operating in the region. Data from more than 23 seismic stations that use short-period, broad-band and accelerometer sensors were used to constrain the hypocentres. The determination encompasses more than 10 phases with more than two $S$-wave arrival times. The earthquake locations were calculated using HYPO71 (Lee & Valdes 1985) and a simplified crust-upper mantle model. Horizontal epicentral errors are smaller than 10 km. Vertical hypocentre errors are in general larger, but we determined the focal depth using the complete broad-band regional waveforms in the SMTI.

In order to assess the effect of epicentral mislocations on our focal mechanisms, we performed several tests using the epicentre determinations. Before the inversion, the north-south and east-west seismic components at each CHARGE seismic station were rotated into the orthogonal radial and tangential components using the INPRES epicentre information. For each event, we produced plots of the horizontal particle motion at each station for a $\sim1.5$-s time window around the first $P$-wave arrival. For a correct epicentre location, there should be no energy on the tangential component, and horizontal particle motion should be maximum in the radial component. We found that horizontal hypocentral determination errors $\sim6$ km produced more than 90 per cent of the seismic energy concentrated in the radial direction with respect to the tangential direction. As another way to test the impact of epicentral mislocations, we determined the focal mechanisms with and without data from the closer CHARGE stations. Data from closer stations are more impacted by epicentral mislocation because the mislocation is a larger percentage of the total path length. Hence it introduces larger errors in the fitting of the data and consequently in the seismic moment tensor results. Finally, we compared synthetic and observed data aligned at the first $P$-wave arrival for each station. This reduces the dependence on location uncertainties, origin time and the chosen seismic velocity structure. Thus, from 75 earthquakes of magnitudes $M\geq4.0$ and depths <50 km that occurred during the CHARGE operation period (Fig. 1) we chose the 27 events with the best locations.

In determining the seismic moment tensor, we used two seismic velocity structures for the region, summarized in Table 1. These structures are based on receiver function analysis (Shearer 2002; Gilbert et al. 2005) and $Pn$ studies (Fromm et al. 2004) that also used the CHARGE data and forward modelling in this study. Model 1 represents the eastern Sierras Pampeanas, and Model 2 characterizes the western Sierras Pampeanas, Precordillera and Cordillera terranes. The main difference in the models is the $Vp/Vs$ ratio in the crust (especially $Vs$). Therefore, for cases involving ray paths in both terranes we combined both structures to obtain a better fit to the data.

We used a bandpass filter between 15 and 50 $s$ for larger earthquakes and 15–30 $s$ for smaller events to compare synthetic and observed data. As a result, we obtained the seismic moment tensor and corresponding best double-couple fault-plane solution for a point source that had the smallest amplitude misfit. We also estimated the compensated linear vector dipole (CLVD) component for each moment tensor. This parameter measures how different the source is from a pure double-couple solution. Thus $CLVD = 0$ per cent means that the moment tensor is 100 per cent a pure double-couple, producing a dislocation which in theory has a geometry identical to the faulting. A discussion about how non-double-couple components affect the solution using the same SMTI technique can be found in Mancilla et al. (2002). The amplitude-misfit normalized error is the sum of the square difference between synthetic and observed waveforms, divided by the sum of the squares of all observed seismogram amplitudes considered in the inversion. The seismic displacements are not linearly dependent on the focal depth. Therefore, we performed a grid search for the source depth, producing synthetic seismograms at a series of fixed hypocentral depths.

Fig. 3 shows the results for event 02-117 ($M_d = 5.3$) that occurred northeast of San Juan on 2002 April 27. We modelled this event at ten CHARGE stations for all three components of displacement using different trial depths. The stations used relative to the epicentre show good azimuthal coverage (Fig. 3c). The results of the focal depth versus normalized amplitude misfit, CLVD component and focal mechanism are shown in Fig. 3(a). The amplitude-misfit error curve shows a single minimum around the best depth (21 km). This is also coincident with a minimum for the CLVD (0 per cent). The focal mechanism solution remains robust for the depth range investigated with the inversions. Fig. 3(b) shows the fit of the data and synthetic for the best depth using a bandpass.

| Table 1. Crustal velocity structures used in the SMTI. |
|---|---|---|---|
| Thickness (km) | $V_p$ (km s$^{-1}$) | $V_p/V_s$ | Density (g cm$^{-3}$) |
| Model 1 | 10 | 5.88 | 1.70 | 2.70 |
| | 35 | 6.20 | 1.70 | 2.85 |
| Half-space | 8.15 | 1.80 | 3.30 |
| Model 2 | 45 | 6.20 | 1.80 | 2.85 |
| Half-space | 8.15 | 1.80 | 3.30 |
filter between 15 and 50 s. The SMTI solution is $M_0 = 5.81 \times 10^{16}$ N m; $M_w = 5.1$; depth = 21 km; fault plane 1: strike = 34°, dip = 40°, rake = 98°; fault plane 2: strike = 204°, dip = 50°, rake = 83°. We used the Aki & Richards (1980) convention for each fault-plane solution, which gives the azimuth of the fault, the dip of the fault considering the block located at the right when looking at the azimuthal direction, and the rake, which is the angle of displacement measured in the fault plane from the azimuthal direction. The Harvard-CMT solution for this event ($M_0 = 8.02 \times 10^{16}$ N m; $M_w = 5.2$; depth = 49.3 km; fault plane 1: strike = 34°, dip = 41°, rake = 31°; fault plane 2: strike = 227°, dip = 71°, rake = 127°) has a similar magnitude with a strike-slip component for the focal mechanism and a deeper source for the centroid located south of the INPRES location. Even though our focal mechanism is different from the CMT solution, the $P$-axis orientation does not significantly change within 8° (Table 2). Modelling higher frequencies (as using the regional SMTI technique) than those considered using the CMT inversion method should give us a better estimate of the depth of the event. There are only two CMT solutions in the region during the CHARGE network operation period that we can compare with (events 02-117 and 01-285). Overall, both solutions have small differences (their focal mechanisms are mainly reverse or mainly strike-slip solutions), although the CMT depths tend to be larger (Table 2).

Fig. 4 shows another example for event 02-107 ($M_D = 4.4$) located north of event 02-117. We modelled this earthquake using the closer stations and data filtered with a bandpass between 15 and 30 s. We present these results using Model 1 and Model 2 separately. The focal mechanism solutions do not change significantly around the best depth. However, a better fit is obtained using the seismic velocity structure described in Model 2 (Fig. 4a). This reflects the fact that most of the seismic ray paths travelled in the western Sierras Pampeanas for the stations considered in the SMTI (Fig. 4b). We show in Fig. 4(c) the synthetic and observed data match for the best depth (25 km) that produced the solution $M_0 = 2.01 \times 10^{15}$ N m; $M_w = 4.2$; depth = 25 km; fault plane 1: strike = 31°, dip = 31°, rake = 124°; fault plane 2: strike = 173°, dip = 65°, rake = 72°.

We also present the SMTI results for event 02-005 ($M_D = 4.8$) with an epicentral location near the Barrancas anticline, the structure on which the 1985 ($M_w = 5.9$) earthquake is thought to have occurred as discussed in Section 5.3 (Figs 2 and 5). The best results show the minimum in the amplitude-misfit error curve as a function of the source depth occurs for a minimum in the CLVD curve at 26 km depth (Fig. 5a). 10 CHARGE stations were used in this inversion (Fig. 5b). Station AREN recorded the largest amplitudes of the three-component seismic displacements used. In order to test how much this station dominates the SMTI we ran the inversion taking out data from AREN. We found similar results ($M_0 = 1.18 \times 10^{16}$ N m; $M_w = 4.7$; depth = 26 km; fault plane 1: strike = 61°, dip = 22°, rake = 175°; fault plane 2: strike = 156°, dip = 88°, rake = 68°). Fig. 5(c) shows the synthetic-observed data comparison for the best depth (26 km).

4 SENSITIVITY TO CRUSTAL MODELS

The seismic modelling for the source depends to second order on the seismic velocity structure used to calculate the Green’s functions (Sileny 2004). In order to test the sensitivity of the solution to the chosen crustal model we selected two earthquakes (events 01-138b and 02-117) with epicentral locations in the eastern and western Sierras Pampeanas, respectively. For each seismic event, we performed a grid search for the crustal thickness, $P$-wave crust velocity, and crustal $V_p/V_s$ ratio using the SMTI method described previously.

First, we constructed the Green’s functions for event 01-138b with magnitude $M_w = 4.3$ located near Córdoba (Table 2 and Fig. 6c) for a set of simple average models consisting of a crustal layer over a half-space. The models had a crustal thickness varying from...
Table 2. Parameters of intraplate earthquakes of the Andean backarc between 29°–36°S and 64°–71°W. Origin time, epicentral location (error ≤ 6 km), and duration magnitude $M_D$ are from the local INPRES catalogue. Magnitude $M_s$ is from the NEIC catalogue. Modelled depth (error ≤ 2.5 km), focal mechanism, seismic moment, moment magnitude $M_w$ (with errors given by the seismogram fitting shown in column rms), and the last three columns are from this study.

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**Figure 4.** Results of the SMTI for event 02-107. (a) Curves with focal mechanisms showing the amplitude-misfit errors between synthetic and observed seismograms versus hypocentral depth for two velocity structure models (Table 1). The minimum is observed for a CLVD = 7 per cent (dashed line) at a source depth of 25 km using Model 2 (western Sierras Pampeanas). (b) Location map of the stations used in the SMTI showing ray paths from the epicentre to the CHARGE stations. (c) Synthetic and observed waveforms for the best fit (25 km) using a bandpass filter of 15–30 s. Epicentral distances and azimuth of the stations used are shown below the station name.

25 to 65 km in steps of 5 km, a $P$-wave crustal velocity varying from 5.8 to 6.6 km $s^{-1}$ in steps of 0.2 km $s^{-1}$, and a $V_p/V_s$ ratio fixed at 1.70. We maintained the $P$-wave velocity at 8.15 km $s^{-1}$ for the half-space with a $V_p/V_s$ ratio of 1.80. We ran an inversion matching synthetic and observed seismograms using a series of source depths for each crustal model. We used a bandpass filter between 15 and 30 s. Fig. 6(a) shows the results for the minimum normalized misfit errors obtained for each model. Accepting misfit errors with rms <0.25 constrains the crustal thickness to be <45 km and $P$-wave...
crustal velocities to be \( \sim 6.0\)–6.2 km s\(^{-1}\). The misfit increases rapidly (>75 per cent) using a high crustal \( V_p \) and/or a thickened crust (Figs 6a, d and e).

The second step consisted of fixing \( V_p \) at 6.0 km s\(^{-1}\) and initiating a grid search around crustal thickness and \( V_p/V_s \) ratio using the SMTI for the same earthquake. This means that only the \( S \)-wave velocity and crustal thickness varied. We ran an inversion for each model of crustal thickness varying the thickness from 25 to 65 km in steps of 5 km and varying \( V_p/V_s \) ratio varying from 1.60 to 1.85 in steps of 0.05. Again, changes in \( V_p/V_s \) ratio correspond to \( S \)-wave crustal velocity variations only. No modifications were made in the half-space described above. Fig. 6(b) displays the results of the inversions obtained using each seismic model. A crustal thickness of \(<44 \text{ km} \) and \( V_p/V_s \) values of 1.65–1.70 produce misfit errors with rms \(<0.25\).

Therefore, models using a low crustal \( V_p \) of 6.0–6.2 km s\(^{-1}\), a low \( V_p/V_s \) ratio of 1.65–1.70, and a crustal thickness corresponding to a normal crust (30 km), or at least no thicker than 45 km, predict the best results. Nevertheless, the seismic ray paths for the earthquake considered in this grid search (event 01-138b) recorded at the CHARGE seismic stations used in the inversion test travelled mostly through the eastern Sierras Pampeanas (Figs 2 and 6c). The focal mechanism solutions did not change significantly around the best depth (Fig. 6c) and remained stable for the entire range of parameters tested.

The same grid-search process was carried out for event 02-117 with magnitude \( M_w = 5.1 \) and epicentre location in the western Sierras Pampeanas (Table 2 and Fig. 3). For this test, we only considered stations located in the western part of the CHARGE network region (compare Fig. 7c to Fig. 3c).

We started testing crustal models with \( P \)-wave crustal velocity varying from 5.80 to 6.60 km s\(^{-1}\) in steps of 0.2 km s\(^{-1}\), \( V_p/V_s \) = 1.80, and thickness varying from 25 to 65 km in steps of 5 km. We maintained the mantle parameters \( V_p = 8.15 \text{ km s}^{-1} \) and \( V_p/V_s \) = 1.80. We ran an inversion for each model and a set of fixed source depths using a bandpass filter of 15–50 s. Fig. 7(a) shows a map of the SMTI minimum amplitude-misfit error results for each crustal model. The best fit (>75 per cent) occurs around a crustal \( V_p = 6.40 \text{ km s}^{-1} \) and a thickness of 45–52 km.

Fig. 7(b) shows SMTI results for a grid search around \( V_p/V_s \) and crustal thickness for the same event. We tested models fixing the crustal \( P \)-wave velocity at 6.20 km s\(^{-1}\) and varying the crustal \( S \)-wave velocity to obtain \( V_p/V_s \) ratios in the range 1.60–1.85 in steps of 0.05. The crustal thickness varied from 25 to 65 km in steps of 5 km. The mantle parameters were the same as before.

In contrast to the test results for the event in the eastern Sierras Pampeanas (Fig. 6b), the best results for event 02-117 occur when considering a crustal structure with \( V_p/V_s \) ratio greater than 1.80 and thickness between 40 and 57 km. Fig. 7(d) shows examples of the amplitude-misfit error and CLVD versus source depth for a set of 45-km-thick crust models. We conclude that the best results are observed using a higher \( V_p/V_s \) ratio (1.80–1.85) if the ray paths travel in the western Sierras Pampeanas geological terranes.

Fig. 8 shows seismic ray paths for variations in \( S \)-wave velocity used in the SMTI of the 27 earthquakes here studied. The \( P \)-wave velocity that we used in all inversions was practically the same for both models (Table 1). However, based on the SMTI results (Figs 6a, d, e and 7a), it seems that \( V_p \) is smaller in the eastern than in the western part of the region. Fig. 8 shows that most of the ray paths are sampling the western geological terranes. However, even with less sampling in the eastern Sierras Pampeanas the results indicate that Model 1 is more appropriate for this region. The last two columns in Table 2 show the total number of seismic components and the crustal model used in the SMTI of each event, respectively.

We did not modify the structure for the upper mantle in this grid-search analysis. Previous studies in the Altiplano using the same technique and band-period ranges of 15–50 s and 15–100 s demonstrated that the SMTI technique is less sensitive to changes in the upper mantle seismic velocities (Swenson et al. 2000).

Our results are in agreement with the \( V_p/V_s \) estimates based on receiver function analysis (Gilbert et al. 2005) and crustal thickness...
Figure 6. Grid search results obtained for event 01-138b using the SMTI technique. (a) Synthetic and observed data misfit errors for different crustal $V_p$ and crustal thickness combinations. We maintained crustal $V_p/V_s = 1.70$, mantle $V_p = 8.15$ km s$^{-1}$ and $V_p/V_s = 1.80$. Dots represent different crustal model parameters used in (e). (b) Synthetic and observed data misfit errors for different crustal $V_p/V_s$ ratio and crustal thickness pairs maintaining crustal $V_p = 6.0$ km s$^{-1}$. The mantle parameters were the same as in (a). (c) CHARGE stations used in the inversions with respect to the event location. (d) Synthetic and observed seismogram fit for the best depth using the model with crustal $V_p = 6.00$ km s$^{-1}$ and thickness = 40 km. (e) Amplitude-misfit errors versus hypocentral depth for different crustal $V_p$ model parameters showed as dots in (a) for a 40-km-thick crust. Dashed lines are the CLVD component in the solution. Note how the focal mechanism solution does not significantly change around the best depth ($\sim 5$ km).

from Pn studies (Fromm et al. 2004) that also used CHARGE data. Previous studies using a dense local network (1987–1988 PANDA experiment) in the region around Sierra Pie de Palo (Fig. 2) covering an area of roughly 150 km north–south by 100 km east–west showed the westward increase of the Moho depth between the Sierra Pie de Palo (48–55 km) and the Precordillera (55–60 km) from P- and S-wave converted phases (Regnier et al. 1994). Interpretations of JHD PANDA station corrections also predict higher crustal velocities ($>6.4$ km s$^{-1}$) near Sierra Pie de Palo than to the west of the Eastern Precordillera (Pujol et al. 1991). These studies found differences in the crustal structure between the Cuyania and Chilenia terranes that we cannot resolve because we have very few pure Chilenia paths (Ramos 1994). Our estimates in the western terranes are in close agreement with those results, although our surface-wave-based average crustal models do not discriminate between the properties of the Pampeanas and Eastern Precordillera terranes.

5 RESULTS AND DISCUSSION

Our results for the events in this study are listed in Table 2 and shown in Fig. 9. The table includes the SMTI results for both fault-plane solutions, the orientation (azimuth and plunge) of the $P$- and $T$-axis, the seismic moment, the moment magnitude and the amplitude-misfit error. Values for $M_o$ and $M_d$ are from the global NEIC and local
Figure 7. Grid search results obtained for event 02-117 using the SMTI technique. (a) Maps of the synthetic and observed data misfit errors for each crustal \( V_p \) and crustal thickness pairs explored. Mantle parameters are as in Fig. 6. (b) Same as (a) but now varying \( V_p/V_s \) ratio and maintaining crustal \( V_p = 6.20 \text{ km s}^{-1} \). Dots show the crust models used in the inversions shown in (d). (c) Ray paths, seismic station and epicentre location used in all of the inversions shown in (a) and (b). Note that the ray paths are travelling in the western geological terranes for this grid searching. The same event, analysed with more station-components, is shown in Fig. 3. (d) Amplitude-misfit error and CLVD component (dashed line) versus source depth for different crustal \( V_p/V_s \) models (dots in diagram b).

INPRES catalogues, respectively. Even though in theory single three-component station waveforms are sufficient to estimate the moment tensor solution and focal depth, we limit the minimum number of stations to four (Fig. 8). The more stable solutions, however, are related to higher numbers of station/components used (see the number of used components (column #comp) in Table 2). We also observed that the CLVD component depends on azimuthal coverage. Hence we present solutions with a minimum misfit error around the best depth coincident with a minimum or an acceptable (<20 per cent) CLVD component.

In some cases, we observed a broad minimum instead of a more local minimum in the amplitude-misfit error curve around the best depth. However, the focal mechanism solutions did not change for the depth-range explored. In general, the depth curves were broader for events located outside of the CHARGE network with magnitudes <3.9. For example, event 00-360 in Table 2 shows a minimum at
Figure 8. Ray paths showing the different seismic velocity models (Table 1) used during the inversion of the 27 crustal earthquakes in this study. Crustal S-wave velocity variations are responsible for $V_p/V_s$ changes. Higher $V_p/V_s$ ratios produced better fits for ray paths mainly travelling in the western geological terranes (see Fig. 2) while lower $V_p/V_s$ ratios seem to better characterize the Eastern Sierras Pampeanas.

20 km depth, but errors are still acceptable for a depth-range of 20–30 km. Higher frequencies are required to constrain the depth better. However, at higher frequencies this is complicated because the solution becomes more sensitive to the complex shallow structure of the crust. Nevertheless, our determinations are still very useful to discriminate between crustal and deeper or intraslab seismic events, and to characterize their focal mechanism solution. We obtained good depth constraints for 80 per cent of the crustal earthquakes studied with an estimated error $\leq 2.5$ km.

Overall, we find the backarc has mainly reverse or thrust focal mechanism solutions with average $P$-axis and $T$-axis oriented at azimuths of 275$^\circ$ and 90$^\circ$, and plunges of 6$^\circ$ and 84$^\circ$, respectively. However, the earthquakes could be representing different styles of deformation as reflected by crustal seismicity variations as we discuss below.

A larger number of events modelled in this study have locations in the western terranes Cuyania and Famatina (Figs 1, 2, 9, 10 and 11). The Pampia terrane does not show seismic activity for earthquakes of magnitudes larger than 4.0 during the CHARGE time period. However, palaeoseismic studies along Quaternary faults point out significant earthquakes that have occurred in the southern part of this terrane (Costa & Vita-Finzi 1996; Costa et al. 2000). Interestingly, we also observed crustal earthquakes further east at about 800 km from the trench near Córdoba in central Argentina related to the Río de la Plata terrane (events 02-074, 01-138b and 01-339). Another distinction already mentioned was observed in the seismic velocity structure for the region. We could fit the observed data better when using an $S$-wave seismic velocity that was 6 per cent lower in the crust (40–57 km thickness) in the western Sierras Pampeanas than in the 25- to 45-km-thick crust of the eastern Sierras Pampeanas. This result could be related to a different composition or a more fractured crust in the west or a combination. Regional surface-wave tomographic studies also show lower $S$-wave velocities beneath the thicker Andean crust than in the eastern shields for this region (Feng et al. 2004).

Earthquakes with epicentre locations in the western Sierras Pampeanas show focal mechanisms with pure thrust or reverse fault-plane solutions. These earthquakes are at depths of between 15 and 25 km, which represent middle-to-lower crustal levels. The only exception is event 01-348, which we will discuss later. In the eastern Sierras Pampeanas, we find strike-slip fault-plane solutions (events 02-019 and 01-138b) at shallower crustal depths. We were still able to model event 02-074 ($M_D = 4.0$) located $\sim 50$ km north of Córdoba using CHARGE data at epicentral distances up to 727 km with poor azimuthal coverage. Its solution shows a reverse focal mechanism and a source depth of $\sim 18$ km. This example shows that it is possible to estimate the moment tensor for $M \sim 4.0$ events with locations outside the network using the SMTI technique. Event 01-285 located in the main cordillera had the largest magnitude ($M_s = 5.6$) of the studied crustal earthquakes. Its focal mechanism solution indicates strike-slip motion at a very shallow depth (5 km). The region around Mendoza showed deeper crustal earthquakes in the range of 10–26 km and reverse focal mechanisms with some strike-slip components. The southernmost located earthquake we modelled is event 01-074, which occurred on 2001 March 15 in the eastern flank of the Andean cordillera with a moderate magnitude ($M_D = 4.9$). Our results indicate a pure-thrust fault-plane focal mechanism and a very shallow depth of 3 km for this event. Notably, the events located south of 33$^\circ$S in the high cordillera showed very shallow depths ($< 5$ km) during the CHARGE period.

5.1 Western Sierras Pampeanas–Sierra Pie de Palo and Valle Fértil Megashear

The region around the Sierra Pie de Palo (Fig. 2) is of particular interest because of the last large damaging ($M_s = 7.4$) earthquake in
Figure 9. Results of the SMTI showing lower hemisphere projection focal mechanisms and source depths in parentheses (Table 2). Dark quadrants are compressional motions. Observe the distribution of the crustal seismicity with respect to the presence of geological terranes and their boundaries. Location of the seismic data showed in the cross-section in Fig. 11 is indicated by the rectangle. Brackets at \(31^\circ\) denote the location of the geological profile shown in Fig. 11.

1977 linked to different interpretations of its causal fault (Fig. 10) (Kadinsky-Cade et al. 1985; Langer & Bollinger 1988; Langer & Hartzell 1996). This Precambrian block is one of the westernmost ranges of the Sierras Pampeanas and the most seismically active in this geological province (Figs 1 and 2). Previous seismic studies show along-strike variation of structures in this area (Langer & Bollinger 1988; Regnier et al. 1992) making it difficult to associate microseismicity with single planar faults (Smallay & Isacks 1990). The 3-D structural geometry of this range is very complex, and its relationship with the adjacent areas is interpreted as different blocks bounded by nearly north-trending thrust faults with important E–W transverse megafractures (Bastias & Weidmann 1983; Perez & Martinez 1990; Regnier et al. 1992; Zapata & Allmendinger 1996; Zapata 1998) (Fig. 10). Ramos et al. (2002) suggested an alternative interpretation for the southernmost block of the Sierra Pie de Palo, a passive roof duplex related to a basement mid-crustal wedge. We found earthquake sources deeper than 15 km for seven events with epicentral locations to the east of the main axis of the major anticlinorium of the Sierra Pie de Palo. Almost all of the fault-plane solutions have trends oriented parallel to this axis and indicate pure thrust or reverse faulting (Fig. 10).

Given uncertainties in the earthquake locations, it is very speculative to relate any one event to one structure. However, because of the increasing interest in relating regional seismicity with structural system or crustal volumes associated with structures in active zones, we think that it is important to briefly put our results in the context of known structures in and around the Sierra Pie de Palo (Figs 10 and 11).

Event 02-117 (Figs 3, 9 and 10) is very intriguing because its epicentre is in the vicinity of the Valle Fértil major reverse fault that borders the western side of this north-northwest trending range and is responsible of its uplift (Jordan & Allmendinger 1986; Snyder et al. 1990) (Figs 2 and 10). This 600-km-long fault zone has rocks characteristic of an ophiolitic belt (Ramos et al. 2000). It seems to have generated several earthquakes recorded by the global seismological network (Stauder 1973; Chinn & Isacks 1983; Dziewonski et al. 1988), which also detected the moderate-size \(M_w = 5.1\) event 02-117. However, the SMTI fault-plane trends of our solution and the depth suggest it may be associated with the main north-northeast oriented axis of the Sierra Pie de Palo and other non-exposed structures located to the north of this range (Zapata & Allmendinger 1996; Zapata 1998). In fact, seismic reflection interpretations at the intersection of the Sierra Pie de Palo and the next range to the east, Sierra Valle Fértil, predict a buried anticline as the northern termination of the Sierra Pie de Palo named the North Pie de Palo anticline (Fig. 10). We speculate that event 02-117 may be related to the fault that bounds the North Pie de Palo anticline to the west. This evidence has two implications: the Sierra Pie de Palo western fault branches from the Sierra de Valle Fértil fault (Fig. 10, inset) as predicted by Zapata (1998). Second, this structure correlated with
Figure 10. Major thrust faults around the Sierra Pie de Palo according to the geological references cited in the text and earthquakes in this study with maximum epicentre-location errors of ±6 km (Table 2). Approximate locations of the historical large crustal earthquake in 1944, and the foreshock (F) and mainshock (M) of the Ms = 7.4 1977 earthquake are also shown. Fractures and faults divide the Sierra Pie de Palo in several blocks. Event 02-117 (Mw = 5.1) at 21 km depth (Fig. 3) is likely related with the northern continuation of this oldest Pampean range in the buried North Pie de Palo anticline (Zapata 1998) (see the vertical projection of this event in the inset). Projections and labels for the seismic events are the same as in Fig. 9.

the northernmost part of a suture between the Precordillera and the western Sierras Pampeanas (Fig. 10) of Grenville age (Zapata 1998) is seismically active.

Another moderate-sized earthquake (event 01-348; Ms = 5.3) occurred in the southeast of the Sierra Pie de Palo (Fig. 10) on 2001 December 14. Its focal mechanism solution indicates a very shallow west-dipping plane or a vertical east-dipping plane and 6 km focal depth. Previous large events in this area like the second shock of the 1977 (Ms = 7.3) earthquake (Kadinsky-Cade 1985) and the 1941 (Ms = 6.7) earthquake (INPRES 2005) have large location uncertainties and do not show a clear correlation with exposed faults (INPRES 1977; Volponi 1979; Langer & Bollinger 1988; Langer & Hartzell 1996). Although a smaller earthquake, event 01-348, is very shallow so we might expect it to correlate with an exposed fault. This earthquake could be related to some continuation to the southwest of active faults seen on the east side of the southern block of the Sierra Pie de Palo known as the Nikizanga fault system (Ramos & Vujovich 1995; Ramos et al. 2002) (Fig. 10). The east-dipping Nikizanga fault showed a 10 km surface rupture during the large deep-crustal earthquake of 1977, and for this reason it is interpreted as a secondary fault (Costa et al. 2000). Local seismicity studies by Regnier et al. (1992) also define a cluster of seismicity around 5 km depth for this area. However, a potential occurrence of event 01-348 along the west-dipping fault-plane solution cannot be ruled out, suggesting its relationship with a shallower décollement under Sierra Pie de Palo like that modelled by Ramos et al. (2002).

There are other events (00-356, 01-107, 01-333, 01-352 and 01-328) located near the terrane boundaries that separate Cuyania from the Famatina and Pampia terranes. They also separate the region in terms of seismicity. Very few crustal earthquakes of magnitudes >4.0 had locations in the Pampia terrane during the CHARGE period (Figs 1, 2, 9, 10 and 11).

Deformation rates based on moment tensor summation (Kostrov 1974) using the Harvard-CMT solutions for events (Mw > 5.0) that occurred in the last 27 yr around the Pie de Palo range (Fig. 2) predict 31.8 mm yr⁻¹ rate of crustal shortening (Siame et al. 2002, 2005). This estimate results from multiplying the horizontal component of the maximum shortening strain rate (oriented in the direction N93°E) times the width of 93 km for the seismically deformed volume of 123 × 93 × 30 km³, which covers the Sierra Pie de Palo. This calculation is dominated by the 1977 earthquake sequence.
Figure 11. SMTI results in a vertical projection along a cross-section around 31°S (see Fig. 9 for location) including: (a) GPS velocities (yellow circles) around this latitude from Brooks et al. (2003), and (b) Geological interpretation modified from Regnier et al. (1992), Ramos et al. (2002) and Siame et al. (2002). The western Sierras Pampeanas are more seismically active than the eastern basement blocks. Focal depths with errors of ±2.5 km are in good agreement with the proposed décollements under Sierra de Piede Palo based on geological studies. Note the different horizontal and vertical scales used in the geological cross section but not for the fault-plane focal-mechanism projections.

We calculated the earthquake moment tensor sum with the same technique using the eight closest crustal events to the Sierra Piede Palo in the 1.5 yr CHARGE period. These events are located in an area of approximately 90 × 38 km² with a seismogenic depth range of 19 km (Figs 9 and 10). In this particular volume, the main contraction axis is oriented along an azimuth 297° and has a plunge of 1°, whereas the minimum contraction axis has an azimuth of 46° and a plunge of 86°. The rate of crustal deformation along the maximum shortening strain orientation over the width of 38 km and the 1.5 yr period is about 0.45 mm yr⁻¹.

As expected, when comparing Siame et al.’s (2005) estimation with our estimate of earthquake crustal shortening, we find that their rate of deformation is nearly two orders of magnitude greater. This is not surprising given the Ms = 7.4 1977 earthquake (Fig. 2). We believe that our estimate is less likely related to the 1977 earthquake sequence since event 06-117 on 2002 April 27 (Table 2 and Figs 3, 9 and 10) in our calculation is the largest magnitude >5.0 earthquake that occurred in this area over a period of 14 yr. However, neither estimate is likely to represent the long-term rates.

Interestingly, the GPS velocity vectors observed in the vicinity of the Sierra Piede Palo and the Sierra Valle Fértil over a time span of ~3–5 yr show a decay towards the east along an azimuth of ~82° (Brooks et al. 2003) (Fig. 11), which agrees with the plate convergence direction (N79.5°E-Kendrick et al. 2003). Our estimate for the maximum shortening axis orientation and that determined by Siame et al. (2005), both based on the moment-tensor sum of earthquakes in the area during different periods of time (1.5 and 27 yr), are ~35° and ~11° clockwise rotated with respect to GPS orientation observations. This might indicate that other tectonic processes are occurring besides crustal earthquakes that affect geodetic measurements. However, it could also indicate that even though the main shortening orientation is that observed by GPS data, the pre-existing fractures and faults associated to the Sierra Piede Palo (Fig. 10) are controlling the moderate-to-large earthquake deformation.

5.2 Eastern Sierras Pampeanas

Focal mechanisms for this region are either strike-slip or reverse fault-plane solutions (Figs 9 and 11). The strike-slip events have very shallow hypocentres. Since our determinations provide the first seismic source characterizations in this region, we consider it important to discuss their tectonic implications in order to exploit their seismic information.

Our results for the largest size (Mw = 4.8) earthquake in this region, event 01-138b (Table 2) that occurred close to Soto, Córdoba, indicate a shallow depth (7 km) and a strike-slip solution (Figs 6 and 9). No movement along strike-slip Quaternary faults has been recognized in the region (Costa et al. 2001). However, interpretations of the orogenic evolution mainly during the Pampean cycle of the Sierras de Córdoba describe eastward dipping, NW–SE trending, sinistral wrench faults for the studied region (Baldo et al. 1996). This observation is coincident with the movement along the N16°W
nodal plane of the SMTI solution for event 01-138b. Tectonic studies by Kraemer et al. (1995) and Martino et al. (1995) also find evidence of north-south trending shear zones related to the suture between Rio de la Plata and Pampia terranes (Fig. 2). Their studies also suggest the reactivation of old faults since the Pliocene. Thus it is likely that the historic 1908 ($M = 6.5$) earthquake that occurred very close to the event 01-138b epicentre and affected the same population centres (INPRES 2005) was generated by similar structures in the region. These findings are important in helping to identify structures that generate seismicity at very shallow depths that might affect cities like Córdoba, which has one of the larger populations in Argentina.

We also constrained two reverse fault-plane focal mechanisms for events 01-339 and 02-074 with magnitudes of $\sim 4.0$ and depths of 16–18 km, respectively (Fig. 9). They represent the easternmost Andean compression acting over a region where the flat slab finally resumes its inclination to the east (Cahill & Isacks 1992) (Fig. 1). Magnetotelluric studies revealed a strong contrast in resistivity for crustal structures that could be responsible for the uplift of the Sierras de Córdoba along east-dipping faults (Booker et al. 2004). The same studies also predict their extension into deeper levels (40 km).

Event 01-349 ($M_D = 4.1$) in the northern region is close to the Sierra Velasco (Figs 2 and 9) and has a dip-slip source mechanism. Both fault-plane trends are parallel to the main NE-trend of the eastern front of this range. This deformation contributes to the uplift of the basement of the eastern Sierras Pampeanas, sometimes generating larger reverse focal mechanisms such as the recent Catamarca earthquake (2004 September 7, Harvard-CMT solution: depth $= 80$ km; $M_D = 6.1$; fault plane: $\phi = 37^\circ$, rake $= 80^\circ$; fault plane $P$: $\phi = 48^\circ$, rake $= 97^\circ$) that occurred approximately 90 km northeast of event 01-349.

Our results in the eastern Sierras Pampeanas are consistent with GPS observations. According to Brooks et al. (2003) the crustal seismicity deformation observed far away from the trench contributes to recover the elastic loading component, which is the entire GPS velocity field observed in the eastern Sierras Pampeanas (green curve in Fig. 1a). Their study assumes that the locking of the converging Nazca and South America plates as far as 600 km to the west generates this elastic component.

5.3 Mendoza

Mendoza has been the site of several destructive earthquakes (Fig. 2). The 1985 ($M_D = 5.9$) earthquake took place $\sim 15$ km south of the city of Mendoza in the Barrancas anticline (Fig. 2) above blind thrust faults (INPRES 1985; Triep 1987; Chiaramonte et al. 2000). The region is characterized by oil fields hosted by the Triassic Cuyo basin hydrocarbon reserves (Moratello 1993; López-Gamundi & Astini 2004). Tectonic analysis based on seismic imaging suggests that the faults involve the basement and the productive traps could be the result of the reactivation of older structures (Dellapé & Hagedus 1993; Ramos et al. 1996).

At this latitude, the Precordillera loses its surface expression, and the Cuyo rift basin as well as a series of low-elevation anticlines continue to the south (Fig. 2) with less east–west contraction (Ramos et al. 1996; Brooks et al. 2000). An estimate of a shallow décollement at about 15 km has been identified to the west of the Cuyo basin (Fig. 2) using balanced cross-sections and seismic reflection data (Kozlowski et al. 1993; Ramos et al. 1996). The west-dipping thrust-fault-plane solutions for events 01-127 and 01-189 ($M_D = 4.2$) at similar depths might be related to this décollement system (Table 2 and Fig. 9).

One intriguing ($M_D = 4.8$) earthquake that we modelled in this region is event 02-005. The focal mechanism has a very low angle southeast-dipping fault plane or a high-angle west-dipping fault plane. Also, it has the deepest depth estimate (26 km) in our study (Table 2 and Fig. 5). Its epicentral location lies below the gentle domes of the Cuyo basin (Figs 2 and 9). Based on seismic reflection data, a tectonic suture in the Cuyania terrane between the Precordillera and the western Sierras Pampeanas basements has been proposed further north (~100 km) at similar middle-to-lower crustal depths (Cominetti & Ramos 1991). Perhaps the very low angle fault-plane solution of event 02-005 is related to the same proposed suture providing evidence of its present activity. This coincides with a major change in the geometry of the subducted Nazca plate near this latitude where the plate changes from a near horizontal dip to a steeper dip of $30^\circ$ (Fig. 1).

The seismic moment tensor solutions for this region give average $P$- and $T$-axis orientations of azimuth $82^\circ$ and $156^\circ$, and plunge of $11^\circ$ and $55^\circ$, respectively.

5.4 High Cordillera

Shallow seismicity that occurs in the high Cordillera is absent north of 33° S for magnitude 4.0 and larger events (Fig. 1). An interesting concentration of crustal seismicity began in 2001 October around Tupungatito volcano (Fig. 2), generating 80 aftershocks in approximately one month (Barrientos et al. 2004). This location marks the beginning of the crustal seismicity to the south. Coincidentally, the downgoing plate changes to a steeper subduction angle, an active volcanic arc starts to the south, and the Precordillera and the basement-cored faulting of the Sierras Pampeanas are absent in the foreland (Figs 1 and 2) (Ramos et al. 2002). We modelled the main shock ($M_D = 5.6$) of this sequence (event 01-285) that occurred on 2001 October 14 southwest of Cordón del Plata in the Frontal Cordillera (Fig. 2). This was the largest continental crust event recorded during the CHARGE experiment. Its focal mechanism corresponds to a near-vertical strike-slip fault-plane solution with a very shallow focal depth of 5 km (Fig. 9). The CMT focal mechanism solution, constrained for a fixed depth of 15 km, is very similar (Table 2).

Another larger ($M_D = 6.5$) earthquake (event 04-241 in Fig. 9) with an epicentre located 190 km southwest of event 01-285 occurred on 2004 August 28, generating 85 aftershocks in 2 days. The Harvard-CMT focal mechanism solution also indicates a strike-slip solution ($M_D = 7.18 \times 10^{18}$ N m; $M_D = 6.5$; depth $= 16$ km; fault plane: $\phi = 21^\circ$, dip $= 61^\circ$, rake $= -178^\circ$; fault plane $P$: $\phi = 390^\circ$, dip $= 88^\circ$, rake $= -29^\circ$). If this earthquake and event 01-285 are being generated by the same major strike-slip Andean structure, we infer that the most likely ruptured plane is very approximate to the NE-striking nodal plane of both solutions. A detailed study including the aftershocks detected for this earthquake following a methodology similar to that used by Fromm et al. (2005) may lead to a determination of the activated fault plane.

Although not in an organized pattern, recent GPS data show a slight clockwise rotation of the backarc vectors with respect to the forearc vectors relative to the stable cratonic South America (Brooks et al. 2003). Studies in other segments (18° S–28° S and 39° S–43° S) of the Andean cordillera have documented strike-slip faults more than 1000 km long (Armijo & Thiele 1990; Hervé 1994). The strike-slip faults in the High Cordillera are also related to the location of an active volcanic arc and a more inclined position of the subducted
Nazca plate segments. Interestingly, those faults do not show high seismic activity (Siame et al. 1997) as it is observed in the 33° S–36° S Andean segment.

Reverse and thrust faults are the main mechanisms that account for shortening in the Principal and Frontal Cordilleran over the steep subduction zone (Kozlowski et al. 1993; Ramos et al. 1996; Cristallini & Ramos 2000). However, the slightly oblique convergence direction between the Nazca and South America plates could also provide a potential source for right-lateral slip parallel to the Andes (Siame et al. 1997). Previous workers have recognized Palaeozoic strike-slip movements along a main fault that limits the Principal and Frontal Cordilleran (Fuentes et al. 1993). The same structure is believed to have been reactivated as a thrust fault that uplifted the Frontal Cordilleran as a block (Kozlowski et al. 1993). The possible occurrence of event 01-285 along this major structure would be an indication of a modern reactivation of this structure as a right-lateral strike-slip fault. In addition, other strike-slip deformation in the High Cordilleran has been recognized constraining focal mechanisms in the Chilean side (Alvarado 1998). The 1958 (Ms = 6.9) Las Melosas, Chile, earthquake is another example of a strike-slip seismic source in the Cordilleran (Piderit 1961; Barrientos et al. 2004) (Fig. 2).

We also modelled another moderate-sized earthquake, event 01-074 (Ms = 4.8), that occurred at the orogenic front of the Principal Cordilleran (Figs 2 and 9). Its location lies 25 km east of the Sosneado volcano in the Malargüe fold and thrust belt (Fig. 2). This thick-skinned deformation was developed by tectonic inversion of a pre-Andean rift system during late Cenozoic times (Ramos et al. 1996). Our focal mechanism solution indicates a shallower west-dipping or a more inclined east-dipping fault plane and a very shallow focal depth of 3 km. The main dip-slip movement found is consistent with the active structures recognized in the area.

Previous seismic studies cited above and our analysis show that moderate earthquakes occur at very shallow depths in this region. Therefore, seismic studies can have important implications in the evaluation of seismic hazards in Mendoza, as other authors have already indicated (Bastias et al. 1993; Zamarbide & Castano 1993; Moreiras 2004).

### 6 CONCLUSIONS

Seismic moment tensor inversions using regional broad-band CHARGE waveforms were determined for 27 moderate-sized (3.5 < Ms < 5.3) earthquakes located in the Andean backarc crust over the flat-slab segment (30° S–33° S) and in the high Cordilleran south of 33° S. The SMTI solutions show mainly reverse or thrust events with average P- and T-axis orientations of azimuth 275° and 90°, and plunge 6° and 84°, respectively. This is also in agreement with previous results around San Juan from the PANDA experiment (Regnier et al. 1992).

The basement of the Cuyania terrane had the highest seismic activity during the 2000–2002 CHARGE deployment. The Valle Fértil fault, which is also a terrane boundary, separates the region in terms of level of seismic activity. Very limited crustal seismicity of Ms > 4.0 occurs in the Pampia terrane over the flat subduction region. However, this terrane transfers deformation into the adjacent cratonic areas. We observed reverse-fault crustal earthquakes in the Rio de la Plata terrane as far as 800 km east of the oceanic trench. This is also observed in the historical seismicity of the region.

Most of the reverse or thrust earthquakes occurred at mid-crustal focal depths (>21 km) in the western Sierras Pampeanas and near Mendoza. They are probably associated with brittle behaviour on deeper décollements that are related to reactivation of old faults and terrane sutures. Although strike-slip focal mechanisms are not dominant features in the region, we found two earthquakes with very shallow strike-slip solutions, one in the eastern Sierras Pampeanas and the other in the high Cordilleran. The event in the Cordilleran as well as another crustal earthquake that occurred in 2004 in its vicinity could be accommodating strike-slip deformation produced by the slightly oblique plate convergence.

Our tests of different average seismic velocity models for the crust indicate that a higher P-wave seismic velocity (6.2–6.4 km s⁻¹), a lower S-wave seismic velocity (VP/VS = 1.80–1.85) and a higher crust (45–52 km) represent the western Sierras Pampeanas, while a VP of 6.0–6.2 km s⁻¹, a VP/VS ratio of 1.65–1.70, and a thickness of 27–37 km characterize the eastern Sierras Pampeanas. Exposed basement rocks in the eastern and western regions also exhibit different compositions. The western Sierras Pampeanas have a more mafic composition with very little granitic rocks compared to the eastern Sierras Pampeanas. The high seismic activity and different seismic properties in the western Sierras Pampeanas terrane may be due to its different basement composition, to a more fractured crust in the western part, or to a combination of these factors over the flat subduction region.

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