THREE DIMENSIONAL *P*-WAVE TOMOGRAPHY AROUND ANTOFAGASTA, NORTHERN CHILE: STRESS DISTRIBUTION ALONG THE DEEPER PART OF THE SUBDUCTING NAZCA PLATE

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INTRODUCTION

Northern Chile is located in one of the most active seismic zones of the Circum-Pacific belt, where the subduction is characterised by a young and fast oceanic Nazca plate, which is underthrusting the South American plate. This process, therefore, is associated with a strongly coupled seismogenic interplate contact, which is capable to nucleate large earthquakes. The last great earthquake in northern Chile occurred on May 10, 1877 (Mw~8.7); using the available historical reports, this event seems to be the largest one that took place there. The 1877 rupture area suggests that it ranges from the south of Arica (19°S) to the north of Antofagasta (23°S) reaching approximately 450 km of rupture along the coastline. Using the 1877 earthquake as reference and considering that during this century, has not occurred any thrust event with comparable size, this area has been defined as the Northern Chile seismic gap.

There have been many investigations oriented to better understand the tectonic framework of this gap, some of them using teleseismic data and others, locally recorded microearthquakes. The main behaviours observed with teleseismic data are in agreement with that obtained using local data, however, local data have permitted to obtain a more detailed behaviour of the subducting slab, with the advantage that locally recorded microearthquakes were obtained with only few weeks of observations in comparison with the available teleseismic data obtained along ~30 years. One of the most interesting behaviour observed in the seismicity of intermediate depths using local networks in northern Chile, is the inverted double seismic zone located at about 100 km in depth, that exhibits both normal and reverse faulting microearthquakes in a close spatial relationship, where downdip tensional events are, in general, shallower than the compressional ones [1]. Local data had been also used to obtain a 2D *P*-wave velocity model evidencing a thin layer of oceanic crust attached to the top of the subducting Nazca plate that reaches ~60 km in depth with a *P* wave velocity of 7.3 km/s [2]. Considering that the 2D body wave velocity models consisted of big structures and the focal mechanisms of the double seismic zone have nodal planes not well resolved because the networks used were located mainly near the coast, we added in this work new data recorded by a temporary network, with a complete coverage from the coast to the Andes Cordillera (Figure 1), in order to simultaneously localize the hypocenters and determine 3D lateral heterogeneous seismic velocity structure. We also performed a formal inversion of the best fitting stress tensor based on the first motion polarities of the *P* waves.

DATA ACQUISITION AND METHOD

Three different sets of *P* and *S* arrival times locally recorded are used in this study. One set corresponds to 336 microearthquakes recorded during the overlapping period of the PISCO and CALAMA networks (5 weeks). The CALAMA network consisted of 11 portable short period vertical analog and digital stations (Figure 1) of the University of Chile through a joint project with ORSTOM. The CALAMA network improved the western coverage of the PISCO network of the Collaborative Reseacrh Center 267 "Deformation processes in the Andes" formed by the Free University of Berlin, the GeoForschungsZentrum at Potsdam, and the Technical University of Berlin, and the eastern coverage of the permanent telemetric seismic network of Antofagasta, installed with the collaboration of the University of Chile, ORSTOM and the IPG, Strasbourg, France. The PISCO network consisted of 32 digital seismic stations, where the majority of them were installed to the east of the Salar de Atacama, near the Chilean and Argentina political boundary. The permanent telemetric network of Antofagasta, installed in June 1990 near the coastline, corresponds to 8 short period vertical stations and one three-component central station. The second set of data used in this study corresponds to 575 microearthquakes recorded by the permanent network of Antofagasta during 1993. The third dataset consisted in 186 events recorded during 8 weeks (1988) by 29 portable stations installed within a 150 km radius of the city of Antofagasta; this data were previously used in the determination of the 2D *P* wave velocity model for the region of Antofagasta.

The arrival times of *P* and *S* waves recorded by the local seismic networks were used to simultaneously determine the hypocenters and seismic velocity structure. The main technique is similar to that used by [4]. The area of study was parameterized as a set of constant velocity blocks of arbitrary dimension. Body wave velocities were specified and determined as independent parameters within each block. The inversion was iterated until changes in velocities became small (<0.05 km/s) and the variance reduction became insignificant (<2-3 %); generally two or three iterations were sufficient. The area of inversion was choosed considering the distribution of seismic stations of the local networks and the microearthquakes recorded. Considering the distribution of the seismicity with depth, we decided to keep the depth range of 0 to 170 km. The first 11 layers had a thickness of 10 km and the three deeper layers were 20 km thick. The strategy of the inversion was to start with megablocks with large dimension (about $120 \times 240 \times 10$ km^3) and diminish systematically the size of the blocks depending of the ability of the data to resolve them. The smaller block obtained corresponds to a volume of $30 \times 60 \times 10$ km^3.

RESULTS OF THE P-WAVE VELOCITY INVERSION

P-wave velocity is parameterized initially by a 1-D horizontally layered model, where the velocity gradually increase with depth. This initial velocity model is an average of the 2D model obtained by Comte et al. [1994]. As a first step, we divided the area of study in three NS megablocks with an EW extension of 120 km, those blocks are named as Western, Central, and Eastern, and for each one it was determined a 1D *P*-wave velocity model. It can be observed that, in general, all of the three models exhibit significant variations from the initial model, mainly along the first 60-70 km in depth. In the Western and Central models, the *P* wave velocity in the layer located at a depth of 30-40 km is smaller than in the embedded layers, suggesting a low-velocity with an average velocity of about 6.7 km/s. This tendency vanishes in the Eastern model. In the second step, the 1D velocity model obtained by the inversion is the initial starting model of the 3D laterally heterogeneous *P*-wave velocity model. A serie of inversions with different size of blocks were run to find the smaller block able to be resolved by the inversion. The obtained model has blocks with variable sizes, which depend on the distribution of the ray paths between hypocenters and seismic stations. The resulting 3D velocity model was composed by 680 blocks for the *P* and *S* velocities, using ~27.000 *P* and ~21.000 *S* arrival times. The minimum number of rays crossing each block was constrained to 20; with this condition, only 571 blocks were conserved for the inversion. The distribution of the *P* and *S* residual errors is less scattered in the 3D heterogeneous model than in the 1D model, as it was expected.

The lowest values of *P*-wave velocity are located in the first shallower layer of 10 km depth, mainly beneath the Quaternary volcanoes, at distances of about 300 to 400 km from the trench, with an average velocity of 5.7 km/s. It can be observed that, in general, the velocity increases with depth, except into the range depth of 30 to 40 km, that presents a lower value of ~6.7 km/s, and extends from the coast up to a distance of about 400 km from the trench, that is, beneath the active volcanic arc. The strongest changes in velocity with depth, can be found at a depth of 40 km near the coast, where the velocity increases of ~6.7 km/s to 7.5 km/s. In the obtained 3D velocity model, it is difficult to identify the geometry of an inclined structure associated with the upper part of the subducting slab, however, it can be observed that, between the coastline and distances of about 200 km from the trench, the average P wave velocities reaches ~7.5 km/s, up to depths of ~70 km . Moreover, at depths greater than ~80-90 km, velocities higher than ~8.0-8.1 km/s start to be observed, suggesting the presence of the lower part of the descending slab. The highest velocities of the obtained model (8.2-8.5 km/s) observed at depths between 90 to 170 km, and at distances of about 230 to 400 km from the trench, suggest the presence of an upper-mantle wedge. The blocks located above this upper-mantle wedge exhibit lower velocities, ranging from 7.0 to 7.9 km/s, and may be interpreted as the root of the Andes Cordillera in this region.

The reliable hypocenters determined with the obtained 3D model are shown in Figure 2, where it can be observed that the seismicity recorded during July-August, 1988; January-December, 1993; and March-April, 1994 exhibits the same distribution pattern. There is no reliable shallow seismic activity that could be associated with the Atacama Fault System. The shallowest events observed in the profile are blasts of the copper mines existing in the region. There is a southward flattening of the slab observed mainly from distances of about 200-300 km from the trench. This flattening is in agreement with the void in seismicity mentioned before, and could be interpreted as an inflection point of the changes in the dip angle of the downdip the slab.

FOCAL MECHANISMS OF THE 1994 CALAMA AND PISCO EXPERIMENT

The focal mechanism solutions of 77 microearthquakes recorded by the CALAMA and PISCO field experiment were selected because they exhibit an adequate azimuthal distribution of the polarities. The majority of these events (70) are located beneath the CALAMA AND PISCO stations, that is, at depths >70 km; however, the cluster of seismicity of depths of ~200 km have a lower definition of the auxiliary nodal plane, due to the network is located to the west of them. The cluster of microseismicity with depths between 90 and 140 km, exhibited normal reverse faulting events, therefore, the earthquakes of this cluster were divided in two groups (tensional and compressional), and for each group we did a joint focal mechanism and stress tensor inversion from the first motion polarities using the Rivera and Cisternas [3] algorithm. It was found that there are 42 normal faulting and 11 reverse faulting events, the average depth of the tensional events is 102±2 km, which is smaller than the average depth of the compressional events (110±3 km). This observation is in agreement with that obtained previously by Comte and Suárez [1]. The main component of the obtained stress tensor (σ_3 for the tensional and σ_1 for the compressional events) is approximately horizontal and the lower component of the stress tensor (σ_1 for the tensional and σ_3 for the compressional) is roughly vertical. The second cluster located at depths of ~200 km, exhibits also normal (7) and reverse (10) faulting events, where the average depth of the tensional events is 201±6 km, while the average depth of the compressional events is deeper, reaching 229±7 km. This is an interesting behaviour of the deeper part of the slab in northern Chile, because this is the first time that there is evidence of compressional events occurring together with tensional seismic activity.

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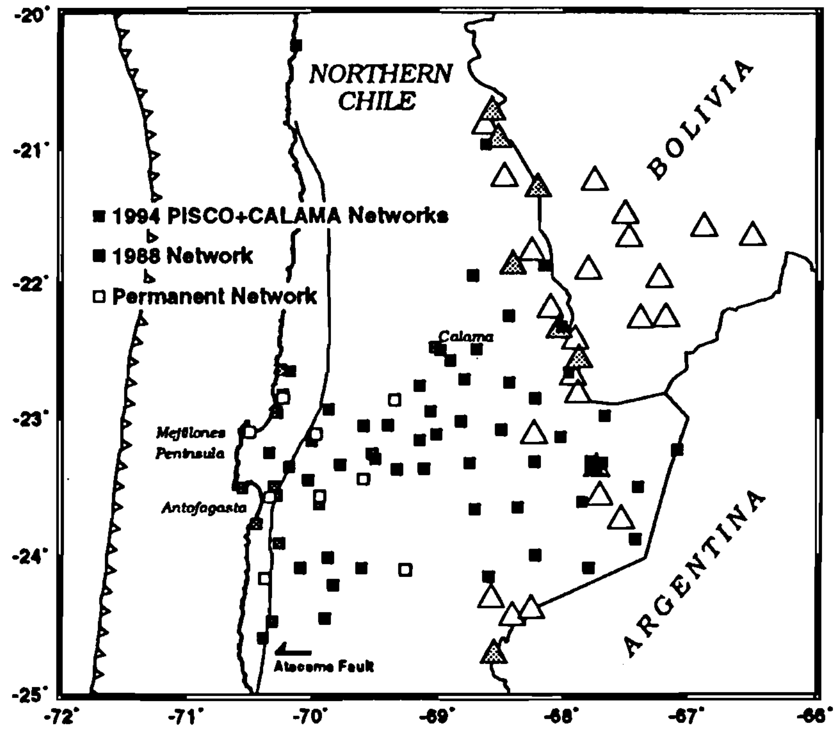

Figure 1.- Distribution of the seismological stations of the networks used in this work. Quaternary and active volcanoes are shown by open and grey triangles, respectively.

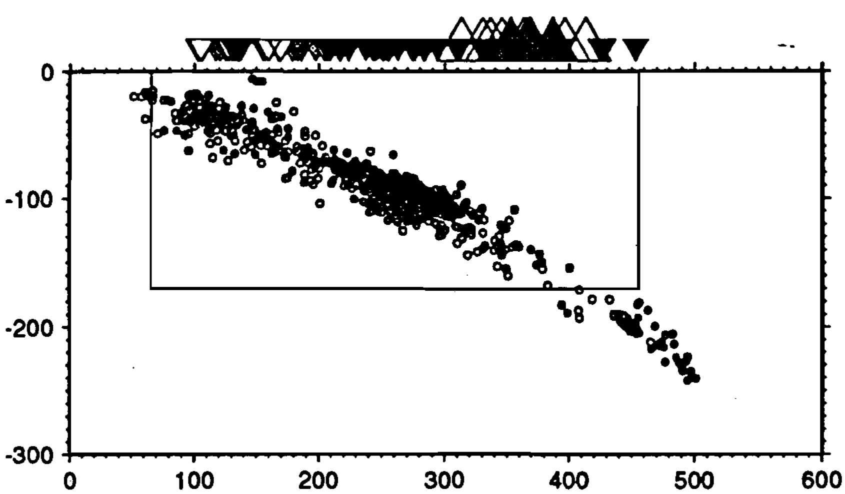

Figure 2.- Depth distribution of the seismicity recorded by the networks used. Dark, grey and open circles corresponds to the events recorded by the 1994 PISCO+CALAMA, 1988, and 1993 permanent networks, respectively. The origin of the profile is the trench, The projection of the seismic station and the quaternay and active volcanoes are also shown. The rectangle corresponds to the area of the 3D inversion.