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Seismicity and Average Velocities beneath the Argentine Puna Plateau

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Abstract. A network of 60 seismographs was deployed across the Andes at ~23.5°S. The array was centered in the backarc, atop the Puna high plateau in NW Argentina. *P* and *S* arrival times of 426 intermediate depth earthquakes were inverted for 1-D velocity structure and hypocentral coordinates. Average velocities and v_p/v_s in the crust are low Average mantle velocities are high but difficult to interpret because of the presence of a fast velocity slab at depth. Although the hypocenters sharply define a 35° dipping Benioff zone, seismicity in the slab is not continuous. The spatial clustering of earthquakes is thought to reflect inherited heterogeneties of the subducted oceanic lithosphere. Additionally, 57 crustal earthquakes were located. Seismicity concentrates in the fold and thrust belt of the foreland and Eastern Cordillera, and along and south of the El Toro-Olacapato-Calama Lineament (TOCL). Focal mechanisms of two earthquakes at this structure exhibit left lateral strike-slip mechanisms similar to the suggested kinematics of the TOCL. We believe that the Puna north of the TOCL behaves like a rigid block with little internal deformation, whereas the area south of the TOCL is weaker and currently deforming.

Introduction

The Puna plateau in the NW Argentine Andes forms together with the Bolivian Altiplano one of the great plateaus on earth. Its formation above a subduction zone has sparked much debate about the mechanism accounting for uplift and crustal thickening [*Allmendinger et al.*, 1997, and references therein].

The Central Andes have been the site of several large seismic experiments in recent years [e.g. *Cahill et al.*, 1992; *Zandt et al.*, 1996; *Graeber and Asch*, 1999]. Activities were concentrated in the forearc and Altiplano, producing detailed information about seismicity, stress distribution, subduction geometry, and crustal and lithospheric structure there. In contrast the Argentine backarc region has remained relatively unexplored. In 1997 we conducted a seismological experiment centered atop the Puna, as part of an ongoing, multidisciplinary research effort of several German geoscientific institutions (Sonderforschungsbereich (SFB) 267) to study deformation processes in the Andes.

We derive a 1-D velocity model for the region and determine precise hypocenter locations for both the intermediate depth and shallow focus crustal events. For many intermediate depth and two crustal events, fault plane solutions were also determined.

Field Deployment, Data, and Analysis

The PUNA seismograph network was in operation for approximately 100 days between late August and late November 1997. Most stations were equipped with short-period 1-Hz 3-component seismometers and *PDAS* data loggers recording continuously 100 sps. Coverage in the western part of the network was complemented by seven short-period vertical seismographs operated by Universidad de Chile and the French IRD, and five broadband instruments from a long-term deployment of the GeoForschungsZentrum Potsdam (Figure 1). Average station spacing was about 40 km.

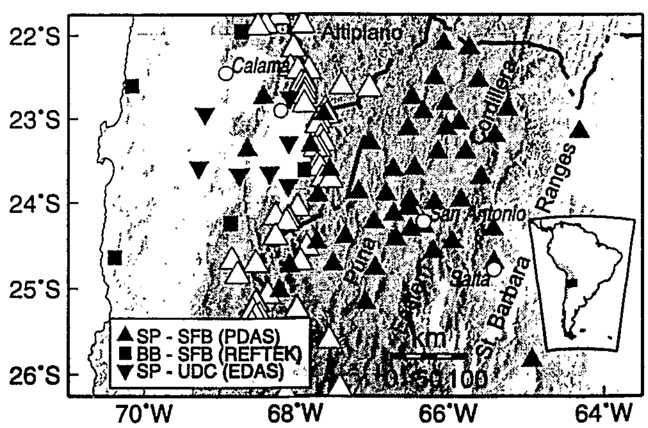

Figure 1. Map of the study area. The area elevated 3000 m above sea level is shaded, approximately indicating the extent of the Puna-Altiplano Plateau. Black symbols mark seismograph locations: UDC-Universidad de Chile, SFB-Sonderforschungsbereich. White triangles are young volcanoes.

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Velocity information for the Puna is very sparse. The only refraction experiment penetrated merely the upper crust [*Wigger et al.*, 1994] and a standard velocity model for the mantle proved inadequate, producing large RMS residuals when locating the intermediate depth events. For that reason the arrival times were used for a joint inversion of 1-D velocity structure, station delays, and hypocentral coordinates [*Kissling et al.*, 1994]. In this inversion scheme the structure is parametrized by a stack of flat layers of constant velocity over a half space. The inverted velocity model will represent ideally the average (weighted by the total raylength) of the true 3-D velocities. Such inversions provide very good hypocentral estimates that differ only marginally from 3-D tomographic inversions [e.g. *Graeber and Asch*, 1999].

The crustal events were excluded from the inversion, because their subhorizontal rays sample only the uppermost, very heterogeneous part of the model. A suite of starting models with 20 km layer thickness has been used to carefully sample the model space (dashed lines in Figure 2). The range of resulting velocity functions that fit the data equally well is shown as shaded band in Figure 2.

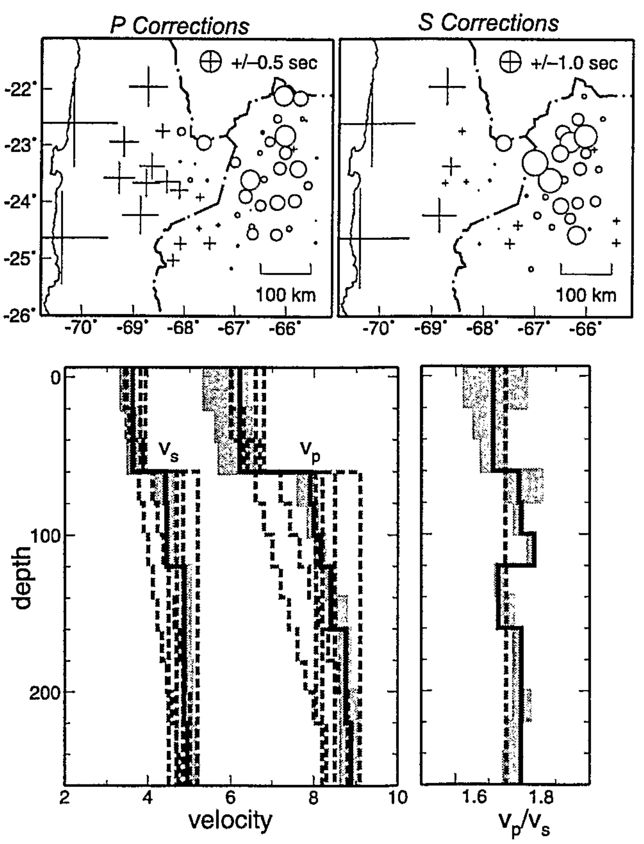

Figure 2. Velocity function from a joint inversion for 1-D velocities, station delays, and hypocentral parameters (black line). The shaded area indicates the range of models that fit the data equally well. This range comes from runs with different starting models shown as broken lines. Above, *P* and *S* station delays are shown. Crosses indicate negative, circles positive delays.

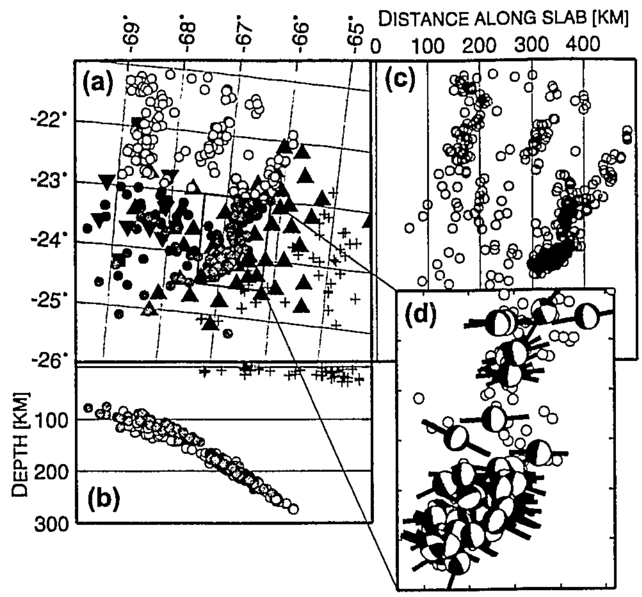

Figure 3. (**a**) Map displaying epicenters. Benioff earthquakes are displayed as circles, with filling changing at 23°S, crustal events are plotted as crosses. Black symbols indicate station locations. (**b**) Hypocenters projected on a vertical cross section. Filling of circles is as in (a). (**c**) Intermediate depth seismicity projected on top of a 35° E dipping plane. (**d**) Detail of (a) showing upper-hemispheric projections of fault-plane solutions from events of the central cluster. Black bars are tension axes projected on the surface.

Most part of the velocity model is well constrained. Only the upper 60 km (crust) show large variations. There, all rays screen the model at similar steep angles, causing a strong coupling of velocities and station delays. For our final model, layers that were not well resolved (upper and lower most parts) or had similar velocities were bundled together in the inversion.

Position and slope of the Benioff zone from our final hypocenters match favorably with the one defined by *Engdahl et al.*'s [1998] relocated teleseismic data. For 93 intermediate depth events in the center of the array fault plane solutions were determined from first motion polarities. The shallow crustal seismicity was located separately from the Benioff zone events. We also applied a joint inversion for hypocentral parameters and station delays, but did not invert for velocities. Local magnitudes range from 0.5 to 4.5.

Results and Discussion

Velocities and Station Delays

Starting with a homogeneous gradient the inversion places the crust mantle boundary at 60 km depth. To test the sensitivity of our data to crustal thickness we also tried to fit other Moho depths. Whereas a 70 km thick crust increases data variance by seven percent, thinner crust is compensated by slower average crustal velocity without loss of data fit (6.0 km/sec for 50 km, 5.7 km/sec for 40 km crustal thickness). To locate the shallow crustal earthquakes velocities of 5.8 km/sec or higher are needed in the upper 20 km, placing a lower bound on crustal thickness to 50 km. 50-60 km crustal thickness is in accordance with results from *Whitman* [1994] for the eastern margin of the plateau and also with Moho depths derived from a recent receiver function study [*Yuan et al.*, 1998].

Average velocity and v_p/v_s ratio in the crust are low, indicative of felsic material [*Rudnick and Fountain*, 1995; *Zandt et al.*, 1996]. Similar values have been found in the Altiplano, and were there interpreted as supportive of crustal shortening as cause of plateau uplift [*Zandt et al.*, 1996]. Mantle velocities increase gradually from 8 km/s to 8.7 km/s at 200 km depth. These velocities are too fast for normal mantle material. We believe that the high velocities at depth are the effect of the subducted slab. This has been corroborated with synthetic traveltimes from a 2-D slab model, showing that inverted 1-D velocities become increasingly influenced by the fast slab at depth. Although there is an increased v_p/v_s ratio in the uppermost mantle, the data do not require a low *P*-velocity zone at that depth range.

Station corrections show a systematic pattern from negative delays in the west to positive delays on the plateau (Figure 2). The negative station corrections in the W agree with a relative fast and thinning crust found there [*Wigger et al.*, 1994; *Graeber and Asch*, 1999], but may also be caused by the slab. We interpret the positive delays at the eastern stations as indicative of lower velocities in the mantle, because for these stations the effect of the slab becomes less influential.

Benioff Seismicity

In Figure 3 the Benioff zone sharply defines a 35° dipping plane, tapering off to only 10-15 km thickness with increasing depth. Different shading of events in the N and S in Figures 3a and 3b shows that the layering in the 100-150 km depth range is an effect of projection. Events in the south are systematically displaced upwards relative to events in the north.

In Figure 3c the intermediate depth seismicity is projected onto a 35° dipping plane. Three lineaments of increased seismic rate striking N to NE are perceptible. This is not an artefact caused by limited recording time, as seismicity plotted from several decades shows a similar pattern [e.g. *Cahill and Isacks*, 1992; *Kirby et al.*, 1996; *Engdahl et al.*, 1998]. Focal mechanisms and tension axes for earthquakes from the most prominent cluster are shown in Figure 3d. Most events have extensional mechanisms with their strikes approximately following the lineament (Figure 3d). The rotation of tension axes away from pure E-W extension, perpendicular to the trend of the lineament, indicates that earthquakes probably occur on pre-existing zones of weakness with roughly aligned faults and cracks. This could for example be caused by subduction of a volcanic chain or ridge as suggested by *Kirby et al.* [1996].

Crustal Seismicity in the South American Plate

Only very little is known about present day deformation in the Puna plateau. Presently, E-W compression is expressed in the seismically active fold and thrust belt of the foreland and to some extent the Eastern Cordillera which forms the eastern flank of the plateau. The easternmost, relative deep events (Figure 3a+b) belong to the Santa Barbara Ranges, a thick skinned N-S thrust belt with deep reaching, compressional seismicity. Event depths up to 30 km have been obtained from a local seismic experiment [*Cahill et al.*, 1992] and are confirmed by our data (E-W cross section in Figure 3). Both regions exhibit Quaternary E-W thrust faulting and known historic seismicity.

Almost none shallow earthquakes in the Puna plateau are reported from global networks. The events located with our local network concentrate around and south of the El Toro-Olacapato-Calama-Lineament (TOCL). The TOCL has been recognized as a major shear zone probably crossing the entire Andes from the foreland in the SE to the Chilean forearc in the NW. Up to 20 km left-lateral displacement has been postulated along the lineament [*Allmendinger et al.*, 1983]. Several of the recorded microearthquakes align along this struc-

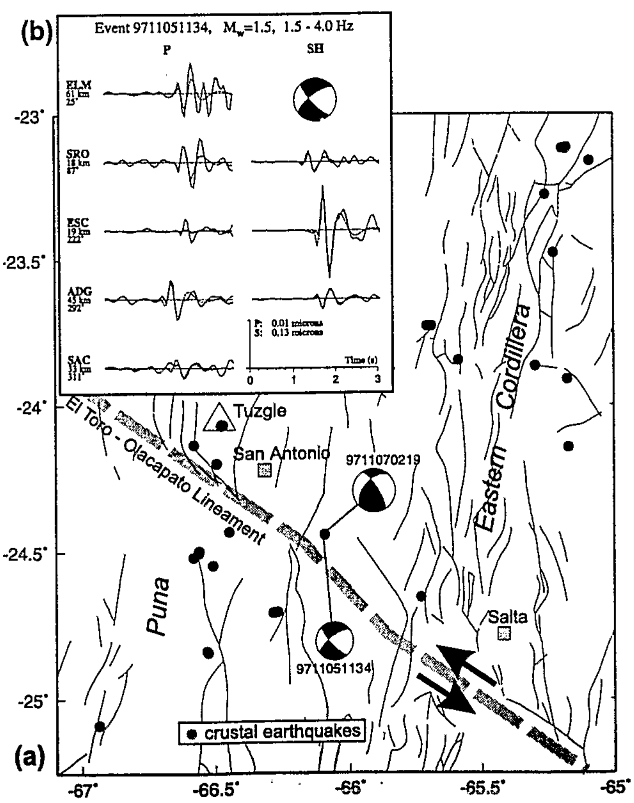

Figure 4. (a) Detailed map of crustal earthquakes. Black lines are mapped faults. Thick grey line indicates the approximate course of the TOCL. Beachballs are lower hemispheric projections of the double-couple estimate from waveform inversion. (b) Example of waveform fit for a small crustal event. Solid lines are observed seismograms, broken lines best-fit synthetics.

ture (Figure 4). For two events we were able to determine the focal mechanism from waveform inversion [*Schurr and Nàbělek*, 1999] (Figure 4). Their strikes and sense of slip agree well with presumed direction and kinematic of the TOCL, suggesting ongoing deformation along this shear zone.

All other microseismicity in the Puna locates south of the TOCL, suggesting that the TOCL marks a structural boundary in the plateau. This is supported by the fact that young mafic backarc volcanism is restricted to the area south of the TOCL [*Allmendinger et al.*, 1997]. It is thought that these dense lavas can rise through the thick crust only along fault zones. The northern Puna appears to be devoid of such zones of weakness, and behaves like a stable block with little internal deformation, whereas the southern Puna is weaker and internally deforming.

Conclusions

Velocities from a 1-D traveltime inversion are low in the crust and high in the underlying mantle. Low crustal velocities indicate a felsic composition. Velocities in the mantle are difficult to interpret because of the presence of a high velocity slab at depth. From the hypocenter locations and focal mechanisms we believe that the earthquakes in the prominent cluster ~200 km beneath the Puna occur along a preexisting zone of weakness, as could be caused by the subduction of a volcanic ridge or seamount. Crustal earthquakes occur along the fold and thrust belts of the foreland and Eastern Cordillera, and along and south of the TOCL in the Puna. The activity along the TOCL suggests that this major shear zone is still active. Because the northern Puna is devoid of both historic seismicity and microseismicity we believe that it behaves as a rigid block with little internal deformation.

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