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Tomography of the Andean crust and mantle at 20°S: first results of the Lithoscope experiment

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Abstract

In 1994, a temporary network of 41 short-period seismic stations operated for 6 months along an east–west profile in Northern Chile and Bolivia. The profile crossed the entire Andean chain at a latitude of 20°S. We present the results of the teleseismic tomography of the crust and mantle along an east–west vertical cross-section beneath the profile. The most striking feature of the tomographic image is the identification of the subducted Nazca plate. It appears as a zone of high velocities which is continuous over the aseismic part of the slab down to the 660 km discontinuity. In the continental lithosphere, strong lateral P-wave velocity variations are found in the crust whereas the upper mantle shows only small velocity variations. Major lateral changes in crustal velocity coincide with the limits of the main structural units of the Andean range. The distribution of velocity anomalies in the crustal layer matches the lateral changes in crustal thickness and average velocity deduced from refraction surveys. A comparison with a previous tomographic experiment in Northern Bolivia confirms that the northern and southern segments of the Bolivian orocline have different lithospheric structures.

1. Introduction

The Andean chain is the most favourable place to study the lithospheric processes associated with the convergence of an oceanic and a continental plate. Its central part (southern Peru, northern Chile and Bolivia; see Fig. 1) is occupied by the second largest high plateau in the world after Tibet: the Altiplano-Puna. Recent field studies show that structural shortening plays a main part in crustal thickening and formation of the plateau in its eastern side (e.g. Schmitz, 1994). However, other mechanisms must be considered to account for the elevation of the western part of the Altiplano and the Western Cordillera. The change in the general trend of the Andes, from N140°E in the north to N0° south of 18°S, is associated with changes in the morphology of the chain, which implies that differences of the deep structures exist between the northern and southern segments.

A first seismic experiment was conducted in 1990–1991 across the northern branch of the Bolivian orocline from the Western Cordillera to the

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Sub-Andean Zone (Fig. 1) using portable stations of the Lithoscope network. This profile of 320 km length provided tomographic images of the lithosphere from both teleseismic events (Dorbath et al., 1993) and local events (Dorbath and Granet, 1996). In 1994, we carried out a second Lithoscope experiment across the southern branch of the orocline, approximately perpendicular to the main Andean structures at 20°S (Fig. 1). Forty-one vertical short-period seismic stations were installed for 6 months along a profile of 700 km length crossing the whole Andean chain from the Coastal Range in Chile to the Bolivian Sub-Andean Zone. Most of the profile used the same line as the PASSCAL BANJO experiment (Beck et al., 1996) but with a closer spacing between stations (20 km). This layout is suitable for investigation of the deep structures by inversion of relative teleseismic residuals and local earthquake arrival time data. This short paper is devoted to the presentation of the first results on the mantle structure deduced from the teleseismic tomography.

2. Data

During the field experiment, we recorded 250 earthquakes at teleseismic distance that were reported by international bulletins. Events with a clear P-wave onset and which were recorded by at least ten Lithoscope stations were selected. This selection reduced the dataset to 120 events. Distances and travel time residuals were computed using the Herrin (1968) tables and the Preliminary Determinations of Epicentres provided by the US Geological Survey.

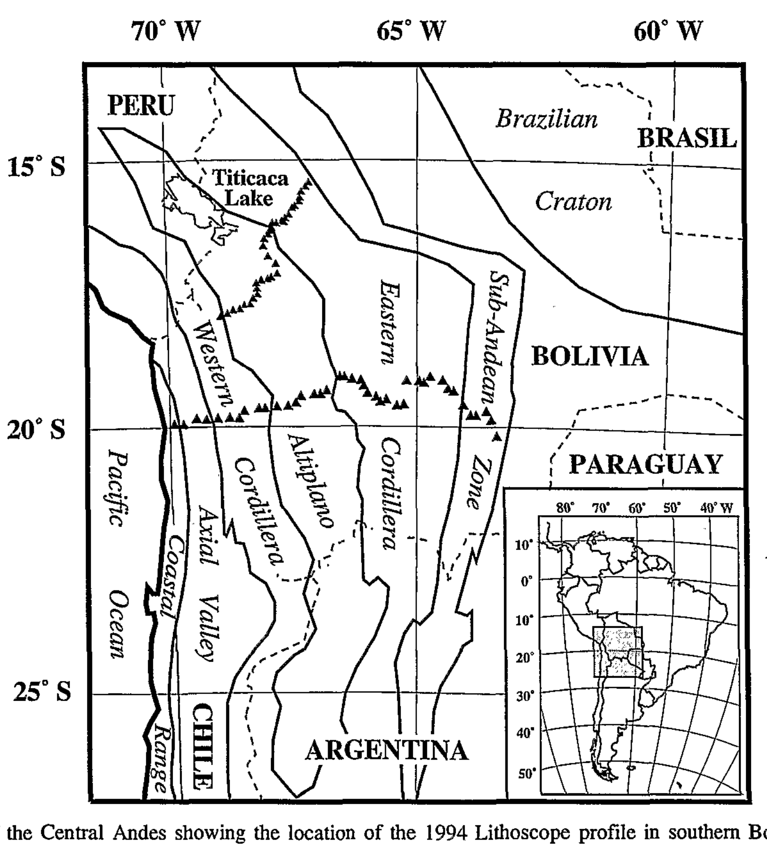

Fig. 1. Structural map of the Central Andes showing the location of the 1994 Lithoscope profile in southern Bolivia and northern Chile (large triangles). The small triangles show the location of the northern 1990–1991 profile. The 1994 transect, which is the subject of this paper, is perpendicular to the direction of the main Andean structures at 20°S and crosses the entire chain from the Coastal Range to the Sub-Andean Zone.

The epicentres of the selected events are plotted as dots in Fig. 2 on a world map which is centred on the middle of the profile. The good coverage in both distance and azimuth ensures that the dataset can be used for tomographic inversion.

The mean PKIKP residuals computed from 30 events located at epicentral distances out of the triplication zone (145–155°) are plotted in the bottom part of Fig. 2. As for all the residuals computed for this study, the PKIKP residuals are relative residuals with respect to Station B111 located on the Altiplano in the middle of the profile. Owing to the quasi-vertical incidence of PKIKP phases, no azimuthal variation of the residuals is observed. The

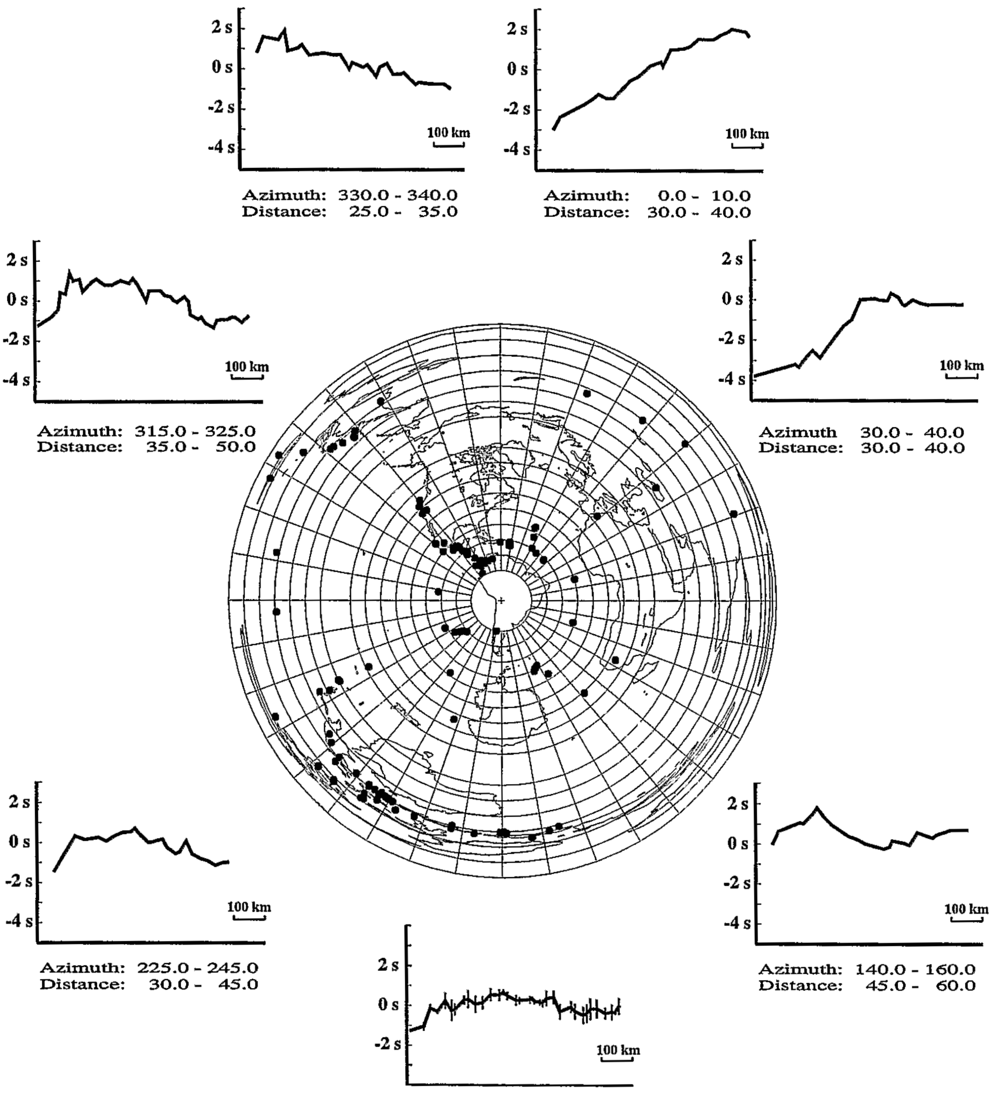

Fig. 2. Distance and azimuth distribution of the 120 events (●) selected for the tomographic study. The world map is centred on the station of the profile which is taken as reference in the residual study. Average relative P-residual curves are plotted around the map at the approximate azimuth of the source region with respect to the centre of the profile. The mean PKIKP residuals computed for the 30 highest-quality events are plotted in the bottom part of the figure with the corresponding standard errors.

first striking result of this study is the weakness of the variations of the PKIKP residuals along the profile. The lowest residuals (i.e. the highest velocities) are observed at the western end of the transect and the highest residuals (i.e. the lowest velocities) are observed in the Altiplano region. The maximum difference in arrival time is 2 s, but it decreases to 1 s if the two very low residuals of the westernmost stations are excluded. From west to east, the mean PKIKP relative residuals increase by approximately 1 s from the Chilean Coastal Range to the Western Cordillera. They undergo only slight variations in the Western Cordillera and the Altiplano before decreasing suddenly by about 0.8 s at the transition to the inner part of the Eastern Cordillera. The residuals are almost constant across the eastern end of the profile (eastern part of the Eastern Cordillera and Sub-Andean Range). These variations are significantly smaller than those observed at the northern profile, where a decrease of 1.8 s was measured between the Altiplano and the central Eastern Cordillera (Dorbath et al., 1993).

Fig. 2 also shows the average relative P residuals computed for teleseismic events from six source regions. Each of the residual curves is plotted around the map at the approximate azimuth of the source region with respect to the centre of the profile. These P relative residuals are affected by strong azimuthal variations. Between 140° and 340° azimuth, variations along the profile of P relative residuals are qualitatively similar to those of the PKIKP phase, with the highest values in the volcanic arc of the Western Cordillera. The peak-to-peak maximum amplitude in P residuals for the azimuth range 140–340° is 2.5 s. For events with north-eastern azimuths (0–40°), the residuals have a very different variation along the profile. The strongest positive residuals are found at the eastern end (Sub-Andean Range) and the maximum difference in travel time along the profile is 4 s.

As core phases light up the lithosphere and asthenosphere at very steep incidences, PKIKP residuals give a crude picture of the lateral velocity variations under the profile. Considering the weak amplitudes of the PKIKP residuals, we can already conclude that strong lateral velocity contrasts are absent under this transect. The very strong azimuthal dependence of the P relative residuals is related to the absence of axial symmetry. Such three-dimensional effects were expected in this part of the Central Andes because the range changes its general orientation here.

3. Inversion

A total of 1670 arrival times were picked including similar numbers of P and PKIKP phases. This dataset was inverted to obtain a model of P-wave velocity anomalies using the ACH technique (Aki et al., 1977). The initial one-dimensional velocity model (given in Table 1) is the same as in the first tomo-

Table 1
Initial P-velocity model

Layer	Velocity ($km\,s^{-1}$)	Thickness (km)	No. of blocks north–south	Dimension (km)	No. of blocks east–west	Dimension (km)
1	6.0	10	1	cones	42	cones
2	6.3	20	3	150	16	50
3	6.8	30	3	150	16	50
4	8.0	40	3	150	10	50
5	8.1	40	5	150	16	50
6	8.2	50	5	150	16	50
7	8.35	50	7	150	18	50
8	8.5	60	7	150	20	50
9	8.8	80	9	150	24	50
10	9.2	80	11	150	28	50
11	9.75	100	13	150	28	50
12	10.3	100	15	150	32	50

graphic study of the Central Andes conducted by Dorbath et al. (1993). It includes a crust of constant 60 km thickness.

The first inversion aimed at finding the velocity perturbations in a two-dimensional model. Owing to its axial symmetry, the resulting two-dimensional velocity model did not succeed in predicting the observed azimuthal dependence of the P residuals and reduced the initial data variance by only 50%. We therefore carried out a three-dimensional inversion. Each layer, except the shallowest one, was divided into blocks with sizes given in Table 1. In the upper layer, a separate block was assigned to each station, following the procedure of Evans and Achauer (1993). As rays that reach two neighbouring stations rarely overlap in the shallowest layer, of 10 km thickness, the inversion process computes a delay, called the cone delay, for each station. This delay accounts for all the perturbations owing to lateral heterogeneities of the upper crust including the sediments.

Our final three-dimensional velocity model is centred at 19.5°S, 66.5°W. The large aperture of our profile (700 km) makes it possible to obtain a velocity model down to a depth of 660 km. The initial variance is reduced by 82% in the final three-dimensional velocity model, about 10% of this variance reduction being accounted for by the station delays.

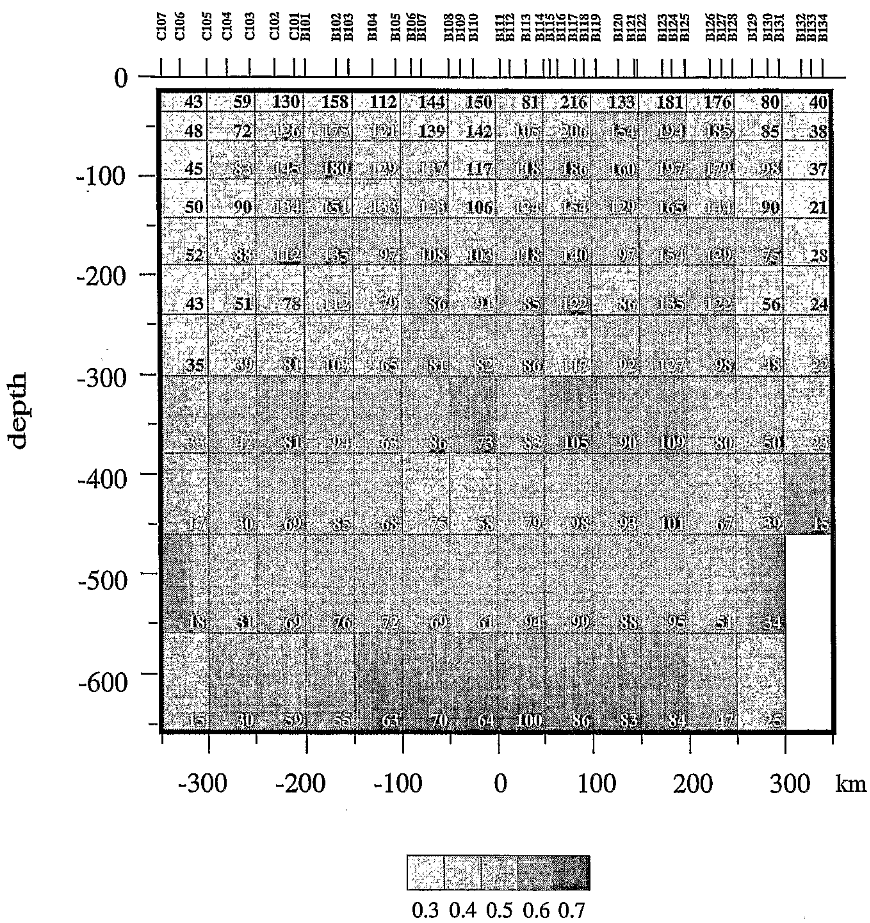

Fig. 3. Value of the diagonal element of the resolution matrix corresponding to each block of the east–west cross-section under the seismic profile (grey scale). The darkest blocks have the highest resolution in the inversion. The numbers indicate how many rays crossing each block were used in the inversion. We did not consider blocks sampled by less than 15 rays.

The central part of the model, located immediately beneath the profile, is the best resolved part, as it is crossed by the largest number of rays. We present here a vertical cross-section through this best resolved part of the three-dimensional velocity perturbation model, i.e. beneath the seismic transect. Fig. 3 shows the value of the diagonal element of the resolution matrix corresponding to each block and the number of rays crossing each block of the vertical cross-section.

4. Tomographic image

A vertical east–west cross-section at 19.5°S through the original block velocity perturbation model is presented in Fig. 4(a), together with the smoothed iso-perturbation contours. In Fig. 4(b), the historical seismicity reported by the National Earthquake Information Service for the time period 1974–1994 is superimposed on the smoothed image of the velocity perturbations. It should be noted that Fig. 4

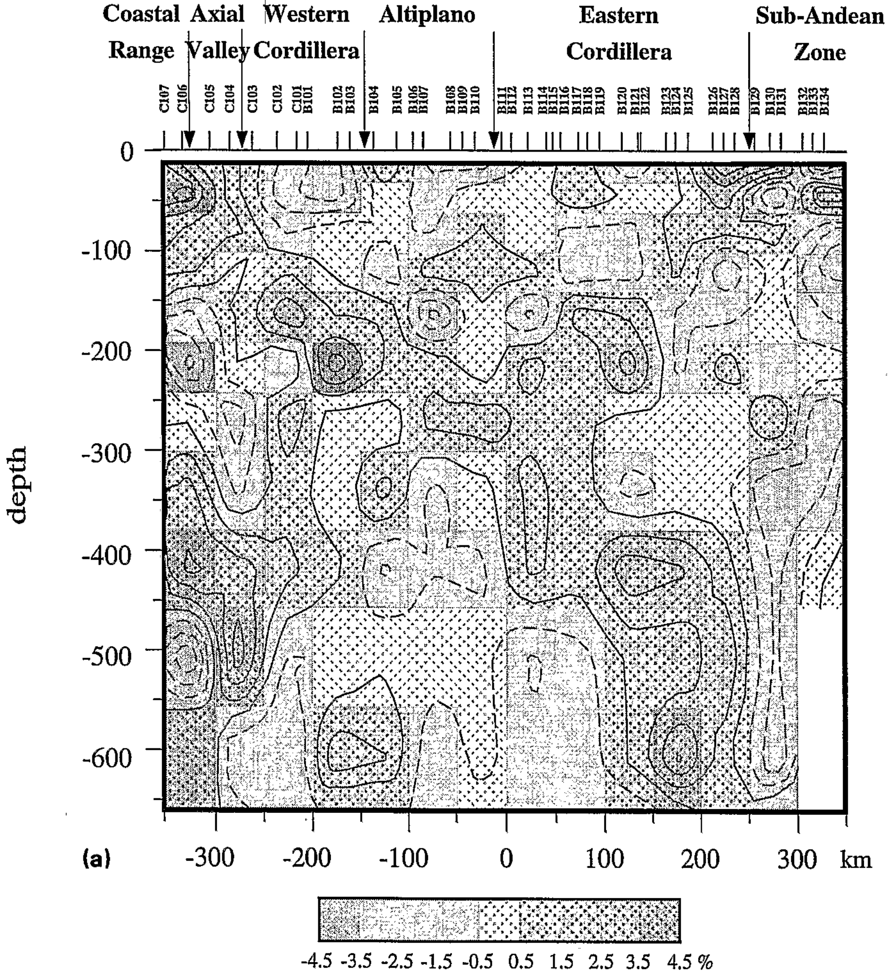

Fig. 4. Tomographic image of P-wave velocity perturbations in a vertical cross-section located underneath the Lithoscope profile. (a) Original block model resulting from the inversion of teleseismic travel time residuals. Iso-velocity contours obtained after smoothing are shown. (b) Smoothed velocity model plotted together with the slab seismicity reported by the National Earthquake Information Service for the past 20 years. Regions with higher velocities than the reference model are characterized by positive perturbations and filled with dotted grey patterns.

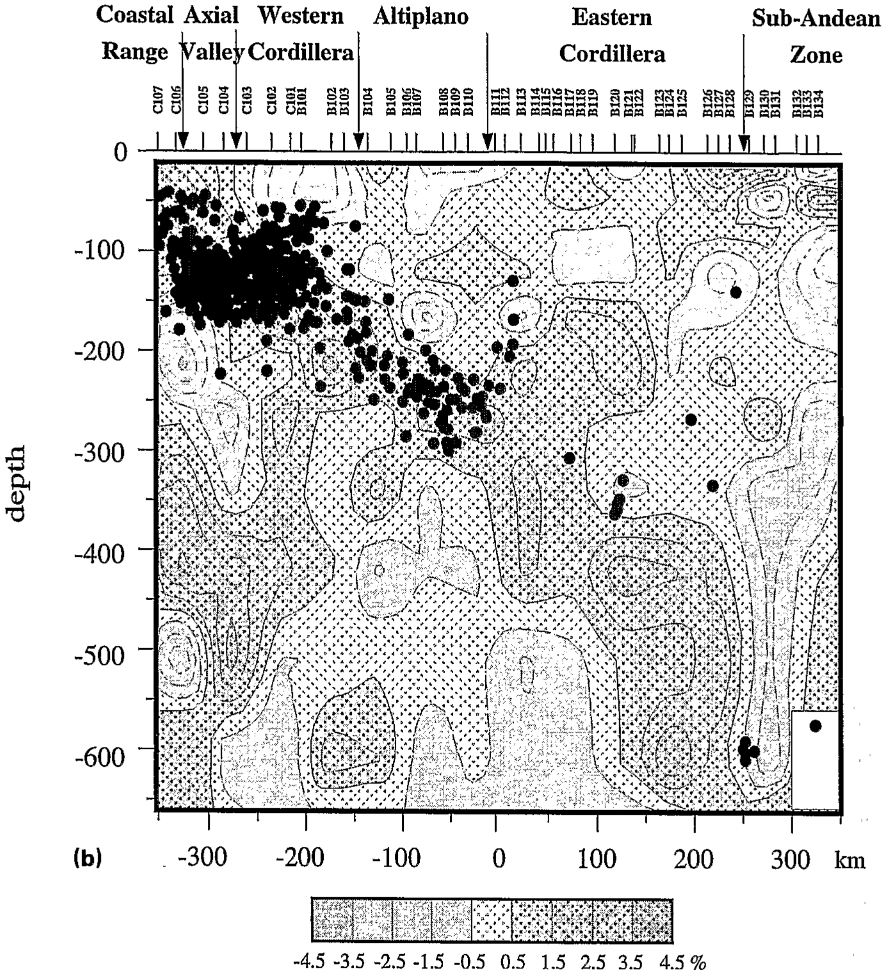

Fig. 4 (continued).

is a vertical cross-section through a three-dimensional model presenting velocity perturbations in all directions. It is only representative of the deep structure in an east–west vertical slice which is 150 km thick in the north–south direction.

The Wadati–Benioff zone associated with the subducted Nazca plate dips at a moderate angle (30°) under the South American plate in the Central Andes. As pointed out by Barazangi and Isacks (1979) and Cahill and Isacks (1992), the shape of the slab is well defined by a high level of seismicity down to a depth of 320 km and by very deep large magnitude events between 520 and 600 km. The two groups of events are separated by a gap which is between 200 and 250 km thick.

We are currently carrying out a tomographic inversion of travel times of local and regional events that will provide a more detailed image of the Andean crust and mantle over the slab. We therefore focus here on the main trends in the tomographic inversion of teleseismic arrival times without going into details in tectonic interpretations.

4.1. The crust

Within the continental crust, the results of the tomography can be compared with the results of seismic refraction profiles recorded at 21°S by the Freie Universität Berlin (Schmitz, 1993, 1994; Wigger et al., 1993). Because the teleseismic inversion has a poor resolution in the vertical direction, we restrict the comparison to the global structure of the crust, i.e. its thickness and average velocity. To make the comparison of the average crustal velocities possible, we performed a second inversion with a new initial velocity model. It included a single crustal layer of 60 km thickness with a velocity of 6.3 km s^{-1} replacing the two-layer crust of the first model given in Table 1. The deeper layers of the initial model remained unchanged.

A comparison between the results of the inversion in the crustal layer and the average crustal velocities measured from refraction data is presented in Fig. 5. On the teleseismic tomography, the highest velocities are observed beneath the western end of the profile, below the Coastal Cordillera of Chile. This positive velocity anomaly is partly explained by the difference between the actual crustal thickness (40–45 km) measured from reversed refraction profiles (Wigger et al., 1993) and crust of the 60 km thickness of our initial model. The positive anomaly is also consistent with the high average velocity of 6.3–6.5 km s^{-1} found in the crust by the same workers. Further east, the positive velocity anomaly decreases rapidly when entering the Axial Valley, as shown by the rapid increase in PKIKP delay times (Fig. 2). This observation also matches the refraction data, which show an increase in crustal thickness from 40–45 km beneath the Coastal Range to 65–70 km beneath the Western Cordillera associated with a decrease in average velocity from 6.3–6.5 km s^{-1} to 5.9 km s^{-1}. The active volcanic arc (Western Cordillera) is characterized by a negative velocity anomaly which fits the low average crustal velocity (5.9 km s^{-1}) determined by refraction. This low-velocity anomaly was interpreted by Schmitz (1993) to be due to partially molten material associated with the volcanic structures of the Western Cordillera. A weak (negative) anomaly is observed under the Altiplano, where the crustal thickness (60–65 km) and velocity structure (6.0 km s^{-1}) are close to those of the reference model. The eastern part of the Eastern Cordillera is characterized by a positive velocity anomaly, which can be explained by the simultaneous thinning of the crust and of the sedimentary cover. At the eastern end of the profile, the weak negative anomaly observed in the Sub-Andean Zone corresponds to a low average velocity (5.9 km s^{-1}). In conclusion, the teleseismic tomography gives a smoothed image of the lateral velocity variations within the crust that matches the structure determined from refraction surveys.

4.2. The continental upper mantle

Fig. 4 shows that the upper mantle between 60 and 140 km has lateral variations in P-wave velocity

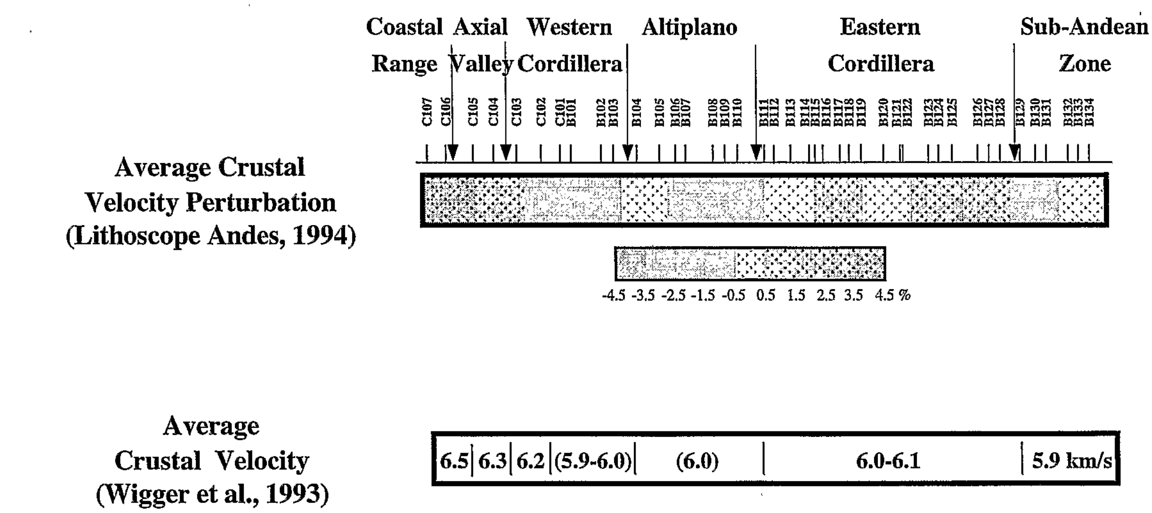

Fig. 5. Comparison between the crustal velocity perturbations given by the teleseismic tomography (perturbation with respect to a crust with constant thickness, 60 km, and velocity, 6.3 km s^{-1}) and the average crustal velocity deduced from seismic profiles at 21°S by Wigger et al. (1993).

that are smaller than $\pm 1.5\%$, which is of the order of the standard deviation. This result agrees well with the low peak-to-peak amplitude in PKIKP residuals. The only correlation with the velocity structure in the crust is the extension of the high-velocity zone under the western part of the Eastern Cordillera down to a depth of 120 km. The velocity contrast associated with this anomaly is smaller in the mantle than in the crust. Under the Western Cordillera, the Altiplano and the eastern part of the Eastern Cordillera, the lack of correlation between the velocity anomalies in the mantle and the crust shows that the mantle and crust probably are decoupled. We do not find any evidence for a wedge of overheated mantle material under the thickened crust of the Altiplano and Western Cordillera which could strengthen the hypothesis of lithospheric delamination (e.g. Isacks, 1988).

As this is the second tomographic experiment conducted in the Central Andes, it is possible to study the along-strike variations in the deep structure of the chain. The previous experiment was carried out in northern Bolivia across a narrower part of the chain, which included only the Altiplano and the Eastern Cordillera (Fig. 1). A comparison of the lithospheric part of Fig. 4 with the tomographic

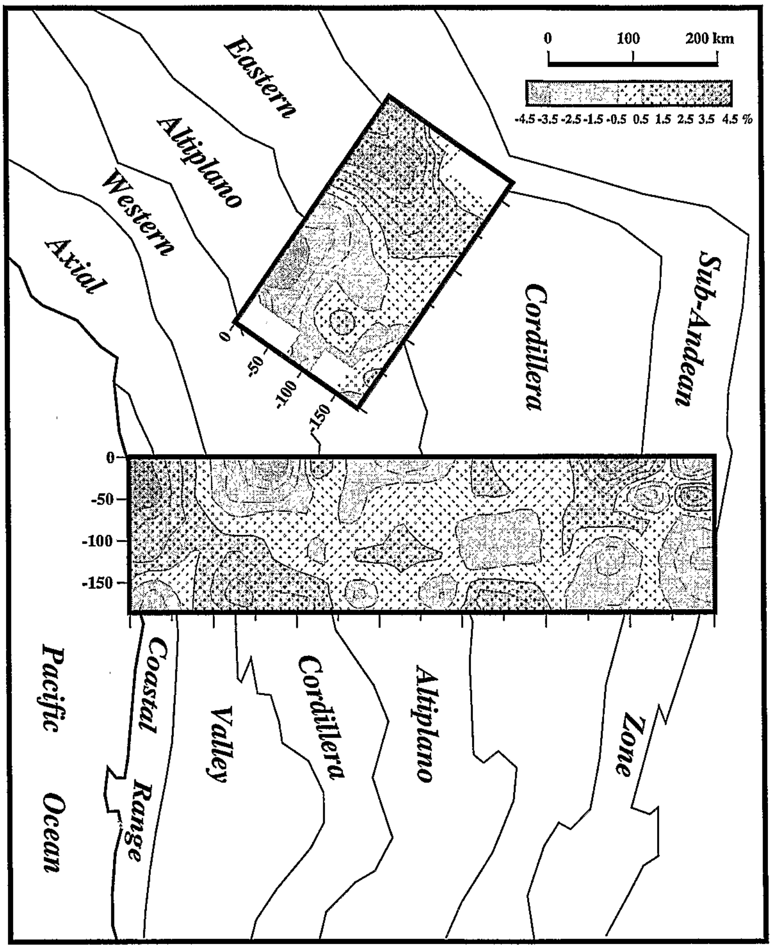

Fig. 6. Velocity perturbations in the lithosphere along two segments of the Andean chain obtained by the two Lithoscope experiments. The 1990–1991 profile (Dorbath et al., 1993) crosses the Altiplano and the Eastern Cordillera perpendicular to the northern segment of the Andes, which strikes N140°E. The 1994 profile crosses the complete southern segment, which strikes N0°. Both cross-sections are plotted on the structural map with their top limit corresponding to their location on the map.

model of the northern profile (Dorbath et al., 1993) is shown in Fig. 6.

The two velocity models are significantly different, as velocities in the shallowest layers have weaker and smoother lateral variations along the southern than along the northern transect. However, the most characteristic features are observed on both profiles. The Altiplano is characterized by low velocities down to the Moho depth. According to Martinez et al. (1995), Cenozoic sediments of the Altiplano are thicker in the north than in the south, which is why the negative velocity anomaly owing to the sedimentary filling is stronger for the northern profile than for the southern one. The difference in sediment thickness is due to along-strike variations in the geometry of the underthrusting of the eastern margin of the Andes under the Brazilian craton. In the northern narrow segment, the underthrusting occurs along steeply dipping faults (Dorbath et al., 1993) whereas in the southern wide segment it occurs along low-angle ramps (Roeder, 1988). This difference in geometry might be also reflected by the difference in the amplitude and shape of the high-velocity anomaly under the axial zone of the Eastern Cordillera. In the northern profile, this anomaly affects a thick part of the continental lithosphere, it is strong, and its western termination is sharp and vertical. The velocity contrast between the eastern and western parts of the Eastern Cordillera is stronger in the north than in the south. This difference possibly indicates that the lateral change in crustal thickness when entering the Altiplano is abrupt in the north and progressive in the south.

4.3. *The subducted Nazca plate*

The dominant deep structure of this transect is the oceanic Nazca plate subducting below the continental South American plate. In a recent paper, Engdahl et al. (1995) used inversion of travel time residuals of teleseismically recorded events of South America to compute tomographic images of the subducted Nazca plate. Their main result is the observation of high-velocity anomalies spatially correlated with the Wadati–Benioff zone which can be traced down to the base of the upper mantle. Although Engdahl et al. (1995) found that the highest-velocity structures are only 2.5% faster than the reference model and do not exhibit a strong continuity, they are similar to the anomalies associated with the subducted Pacific plate below island arcs of the northwest Pacific (Van der Hilst et al., 1991).

A striking feature revealed by our tomography (Fig. 4) is the eastward-dipping positive anomaly which crosses the entire profile from the coastal zone to the Sub-Andean Zone. This structure, of 100 km thickness, is characterized by a +2% average velocity contrast with respect to the surrounding mantle, with a maximum contrast of +4%. It is associated with the slab because it coincides with the Wadati–Benioff zone in its seismic part (Fig. 4(b)). The continuity of the high-velocity anomaly down to 660 km indicates that the slab is continuous across its aseismic part. The strengthening of the velocity perturbation at 400–450 km depth and the general shape of the anomaly indicate that the dip of the descending slab increases below 400 km depth.

In agreement with results presented by Engdahl et al. (1995), we find that the velocity anomaly associated with the Nazca plate is only of +2 to +3%, which is at the lower bound of the resolution power of tomographic techniques. Similar results have been obtained by teleseismic tomography for other subduction zones. For example, Hirahara (1981) measured a perturbation of +2.5% for the subduction of the West Philippine Sea plate under Kyushu, and Harris et al. (1991) found +3 to +4% for the subduction of the Juan de Fuca plate under the western margin of the North American plate. On the other hand, subduction zones investigated by local earthquake tomographies display higher velocity contrasts, reaching +6% for western Pacific subduction zones (Zhao et al., 1992). The resolution tests performed by Engdahl et al. (1995) show that the amplitude of the velocity perturbations can be underestimated by several per cent in teleseismic tomography. A slab with a shallow dip such as the Nazca plate cannot induce strong lateral contrasts in the layers of the initial model. Consequently, the inversion method used here gives a lower bound for the velocity perturbation associated with the slab.

5. Conclusion

The close spacing (20 km) between receivers of this second Lithoscope experiment in the Central

Andes and the appropriate distribution of recorded teleseismic events made it possible to perform a tomographic study of the crust and mantle at 20°S. Owing to the large aperture of the profile, we were able to obtain a velocity perturbation model across the whole width of the Andean chain and its underlying lithosphere and asthenosphere down to the base of the upper mantle.

The lack of strong lateral P-wave velocity variations is revealed by the small variations of the relative PKIKP travel time residuals along the profile. On the other hand, the strong azimuthal variations observed on relative P travel time residuals show the absence of axial symmetry of the Andean structures.

A comparison of the lithospheric part of the velocity perturbation model of the southern profile with the results for the northern profile of 1990–1991 shows that the most characteristic velocity anomalies are present beneath both profiles, although they are less contrasted in the south than in the north. Moreover, the different average crustal velocities and crustal thicknesses associated with the main structural units, as reported by refraction studies in southern Bolivia, appear clearly on the teleseismic tomography.

Low-velocity anomalies are observed under the volcanic arc in the continental crust. A negative anomaly was also expected between the slab and the continental lithosphere, as this transect of the Central Andes is characterized by abundant Quaternary volcanism, suggesting the existence of a mantle wedge of asthenospheric material. According to Isacks (1988), the wedge would have its tip located below the volcanic arc of the Western Cordillera and would play a major part in the building of the Altiplano Plateau. However, this tomography does not show any clear low-velocity anomaly above the slab under the Altiplano but only some small slow anomalies. This feature was already reported for the northern profile. The tomographic images that will be computed from delay times of regional earthquakes should provide better constraints on the existence of this mantle wedge.

The most striking feature of the tomographic image is the identification of the slab by high velocities. The positive anomaly associated with the Wadati–Benioff zone of the Nazca plate is continuous at depths down to the 660 km discontinuity, providing new evidence for a continuous slab over regions with gaps in seismicity. Further studies are required to investigate the precise geometry of the subducted slab.

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