GEOPHYSICAL RESEARCH LETTERS, VOL. 23, NO. 2, PAGES 205-208, JANUARY 15, 1996

# Velocity structure of the Andes of central Peru from locally recorded earthquakes

Catherine Dorbath

ORSTOM and EOPG, URA CNRS 1358, Strasbourg, France

**Abstract.** Arrival-times of local events recorded during two short term seismic fied investigations in central Peru are used to determine the P and S velocity structures above the slab, in a segment of the Andes where the subduction is horizontal. The velocities in the upper-crust can be related to the sedimentary cover. The Moho depth is estimated to be about 50 km below the western part of the sub-Andean zone, and deepens gently to the west, reaching 60 km under the Western Cordillera. An uplift of high velocity material is observed under the coastal zone, associated with a cluster of earthquakes in and above the slab. The asymmetrical structure of the crust below the Eastern Cordillera reflects the convergence process in this part of the Andes: the shortening occurs along active reverse steeply west dipping faults involving the crust in its entirety on the eastern border of the cordillera, the compression then decreases toward the east.

## Introduction

In 1985 and 1986, two microearthquake surveys were conducted in central Peru, accross a complete section of the Andean chain at about 12°S, from the coast to the sub-Andean zone (Dorbath et al., 1990 and 1991, Lindo et al., 1992). The purpose of this paper is to use arrival times of P and S waves generated by earthquakes located in the subducted Nazca plate and in the crust to determine P and S velocity structures across the whole Andean orogen.

The Andean chain is the result of the convergence of the Nazca Plate and the South American Plate. Its well known segmentation along strike, characterized by changes in tectonic styles and volcanic activity, correlates with the shape of the subducted plate (Jordan et al., 1983). In central Peru, the Andes show a subhorizontal subduction angle. This segment of the Andes, where recent magmatic activity is absent, is divided, from the Pacific Ocean to the Brazilian craton, into nearly parallel morphological structures: the coastal zone, the Western Cordillera, the High Plateaus, the Eastern Cordillera and the sub-Andean zone (figure 1). The following description of the structural setting comes mainly from Megard (1978, figure 2a).

Between Lima and Pisco, the inland part of the coastal zone, with mainly Mesozoic volcano-detrital gently folded sediments outcroping, is very narrow and subsiding (Macharé et al., 1986). The Western Cordillera, with maximum elevations of 5200 m, is formed by folded upper Jurassic to Eocene marine and continental sediments intruded by the Coastal Batholith and covered by Oligocene and Miocene volcanic series. The High Plateau, about 3800 m high, is a 50 km wide intra-cordilleran zone (the Mantaro basin), characterized by marine Mesozoic formations. It is enclosed by active fault systems (Dorbath et al., 1990): Altos del Mantaro to the west and Huaytapallana to the east. The Eastern Cordillera, which culminates at 5557m, is a wide anticlinorium of essentialy Precambrian and early Paleozoic rocks. It is separated from the sub-Andean zone by active west-dipping high-angle reverse faults where crustal shortening and uplift of the crystalline basement take place. The deformation of the sub-Andean hills decreases toward the Brazilian shield. The thickness of the late Mesozoic to Tertiary terrigeneous sedimentary cover is poorly known, however, to the north, it may reach 10 km (Pardo, 1982).

The two seismic investigations give a complete section of the seismicity of this Andean segment for about 475 km from the trench (figure 1). The first one was located in the Eastern Cordillera and the sub-Andean zone. It provided a precise description of the crustal seismicity and the present tectonic activity of the region (Dorbath et al., 1990, 1991). Simultaneously, we recorded several local events related to the subduction zone under the array. The second investigation took place between the coast and the High Plateau across the Western Cordillera. Most of the seismicity recorded was related to the slab. Together with the results of previous surveys, it provided an accurate description of the shape and the stress regime of the subducted slab (Lindo et al., 1992, Lindo, 1993).

The data collected during these investigations are well suited for Local Earthquake Tomography (LET). LET offers the possibility of improving our knowledge of the velocity structure of central Peru down to the subducted slab, since the depth extent of LET models is limited by the maximum earthquake focal depth. LET can help to estimate the Moho depth and can define more precisely the shape of the subduction zone by refining the hypocenters locations. In this paper, we will invert separately the data from the two seismic surveys in order to obtain the structures below the coastal zone and the Western Cordillera (Western Survey, WS) on one hand, below the Eastern Cordillera and the sub-Andean zone (Eastern Survey, ES) on the other; the results will then be merged to present a global cross-section of the Andean chain.

## Data and inversion

The data sets include the arrival times of P and S-waves from the local events recorded by the temporary stations. Complete descriptions of the networks, velocity models and data analysis may be found in previous papers (Dorbath et al., 1991, Lindo, 1993). Here we will only summarize the description of these surveys and the data processing.

The WS array included 19 short-period vertical analog and 3-component digital seismic stations. The ES array included 20 stations (11 analog and 9 digital). The uncertainties on P and S waves arrival times read on analog recorders are estimated to be less than 0.1 s and 0.5 s respectively. The uncertainties on arrival times read on digital records have to be reduced by a factor of two.

The earthquakes were located using the HYPOINVERSE program (Klein, 1978) The thicknesses of the horizontal layers were adjusted to simulate the change of the Moho depth



Paper number 95GL03778
0094-8534/96/95GL-03778$03.00

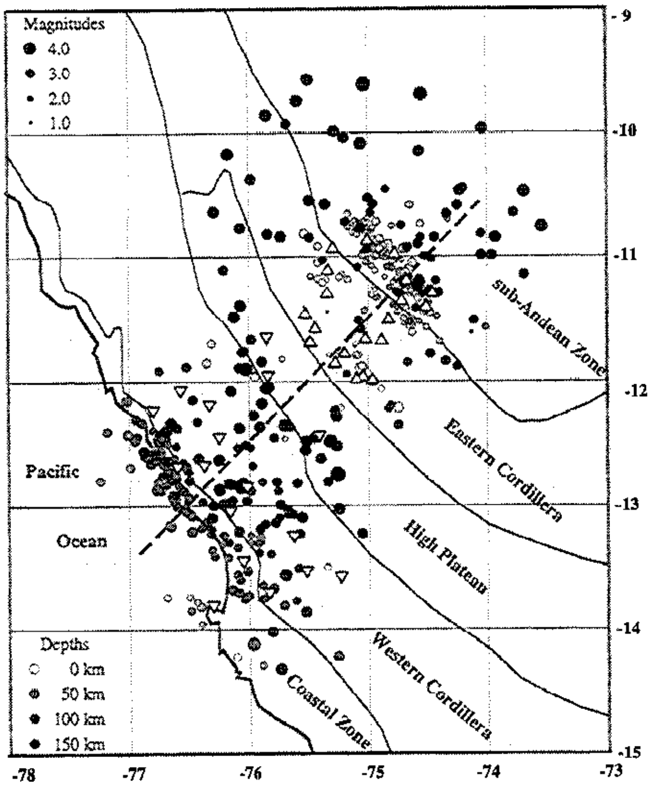


**Figure1:** Distribution of stations and events used for the Local Earthquake Tomography together with the morphostructural zoning of the central Peru. The shaded circles show the initial epicentral locations of the local events, the darker the deeper. The black triangles show the position of the temporary seismic stations (inverted: Eastern Survey - normal: Western Survey. The dashed line shows the location of the cross-sections presented on figures 2a and 2b.

across the Andean chain. We only kept hypocenters which met restrictive criteria ensuring a high quality location. For the WS, 220 events passed these tests, providing 2628 arrival times consisting of 1412 P and 1216 S arrivals; for the ES, the complete data set includes 3742 arrival times (2317 P and 1425 S arrivals) from 297 events, 197 of which are crustal events. The mean mathematical error on the location depends on hypocentral depth; the vertical error is 2.8 km for the western part of the Wadati-Benioff zone, 3.4 km in the center, 4.0 km at the eastern end, the horizontal error is one km less. We may estimate the location precision to be twice these values.

The inversion program we used is the latest version of the program adapted by Eberhart-Phillips (1986) from the method developed by Thurber (1983). The iterative process inverts P and S-wave arrival-time data for a 3D velocity structure and hypocentral parameters. The parametrization of the regions under study is achieved by assigning velocity values at fixed points on a 3D grid. An approximate ray tracing algorithm (ART, Thurber, 1983) is used to calculate travel-time between the source and the station. This method combines parameter separation and a damped least squares inversion.

The initial P-wave velocity models were smoothed versions of the models used for hypocentral locations. Vp/Vs ratio was fixed at 1.73. The velocity grids were aligned north-south and the horizontal grid spacing was one fourth (WS) or one fifth (ES) of a degree underneath the seismic stations. The starting models consisted of 15 km (ES) to 20 km (WS) spaced layers of grid points from the surface down to the maximum depth of hypocenters.

For selecting the optimal value of the damping factor, we constructed "trade-off" curves of resolution against solution error, and found values of 25 for WS and 10 for ES. The stability of the solutions has been tested by the usual methods: variation of the spacing of the grid points, variation of initial velocities, etc. The initial data variances were 0.091 $s^2$ (WS) and 0.089 $s^2$ (ES) for P waves and 0.266 $s^2$ (WS) and 0.217 $s^2$ (ES) for S waves. After 4 iterations, the variances reduce to 0.017 $s^2$ (WS) and 0.012 $s^2$ (ES) for P waves, 0.084 $s^2$ (WS) and 0.061 $s^2$ (ES) for S waves. Thus, the final reductions of the initial data variance are significant: 81 per cent (WS) and 86 per cent (ES) for P-waves, 73 per cent for S-waves.

The reliability of the inversion is indicated by the value of the diagonal elements of the resolution matrix, with the value of off-diagonal elements displaying the coupling between neighbouring nodes. Thus, when drawing the velocity models, we take into account only the nodes where the diagonal element of the resolution matrix is greater than 0.4. Below 75/80 km, the resolution becomes les than this because the nodes are not well sampled by rays.

## Results and discussion

In this short paper we choose to present only vertical cross-sections orthogonal to the Andean trend and passing through the best resolved nodes of both models, since this shows more clearly lateral variations in velocities than map views of the different layers.

The figure 2a shows a geological cross-section through the Andes just to the north of the studied area and in the same tectonic context (Megard, 1978). The figure 2b and 2c present cross-sections trending N45°E through P- and S-wave velocity models and through the relocated hypocenters. Although the relation between the geological section (figure 2a) and the seismic sections (figures 2b or 2c) is not straigh forward in detail, the cylindrical symmetry of the Andean chain allows some comparison.

A comparison of the original and final locations shows small variations of 0.06° on average of epicentral positions; the mean depth variation is 4 km. These changes on hypocentral locations are not significant with regard to the crustal seismicity. On the other hand, these changes appear to better deliniate the seismicity of the subduction zone. The slab dips at about 30° from the trench to a depth of about 100 km where its slope changes and the slab becomes subhorizontal. The seismicity is scarse in the central part of the subhorizontal segment of the slab, below the High Plateau.This lower level in seismicity is observed on any study of the subduction in Peru and is not due to the station spacing between the two networks. A dense swarm is seen above the slab under the coast line. This seismic activity is concentrated between 40 km and 80 km in depth (figure 1). Such a swarm may be identified on the cross-section for the same region from selected ISC and USGS data for a different time period. Therefore, this seismicity is a long period feature. The relocation of locally registered events underlines the clustering of these events.

A noticeable feature seen on figure 2b is the close correspondence of the iso-velocity lines between the two P-wave models obtained by completely independent inversions. Unfortunately, there is a small gap in the velocity structures between the two surveys, due to the poor resolution of the external nodes in the inversions. However, the interpolation of the velocities between these two models is direct. The interpolation is not so easy for S-wave velocity models, because far fewer nodes are resolved and thus the gap between the two models is larger. It is also interesting to note that the general behaviour of P- and S-wave velocities is very close. This is confirmed by the fact that the Vp/Vs ratio computed for the different structural units does not vary significantly.

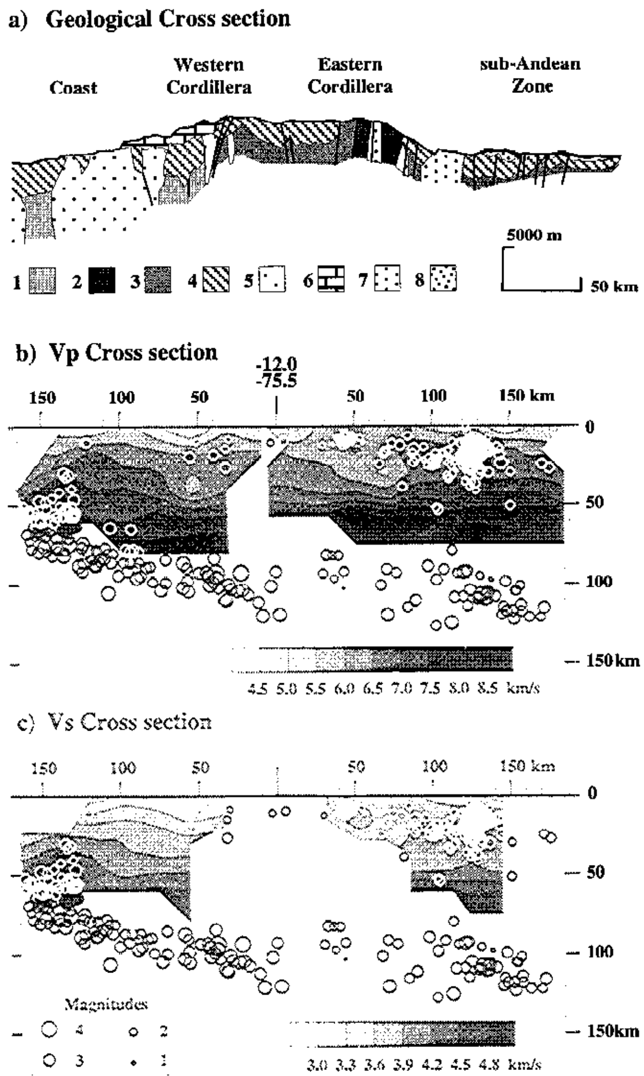


**Figure 2:** a) Simplified geological and structural cross-section of central Peru after Mégard (1983). The numbers of the legend refer to the different units: 1: Precambrian Substratum. 2: Precambrian. 3: Paleozoic. 4: Mesozoic. 5: Late cretaceous terrigeneous series. 6: Late Eocene - Miocene folded volcanical series. 7:Permian granitoids. 8: Andean granitoids. 9: Mesozoic volcanic flows. b) Vertical cross-section through the relocated hypocenters and the smoothed P-wave velocity model of the 3D inversion. To get the smooth pattern, we applied a bilinear interpolation function to the discrete velocities. This smoothing preserves the amplitudes in individual nodes of the original discretization. Higher are the velocities, darker are are the grey areas. c) Same as figure 2b for the S-wave velocity model.

P-wave velocities smaller than 5.5 km/s in the uppercrust are clearly related to the presence of sediments (figure 2b). Slow zones are observed under the Western Cordillera, the High Plateau and the sub-Andes. In contrast, the Eastern Cordillera, where Precambrian series outcrop, is characterised by velocities higher than 5.5 km/s. The same feature can be seen on the figure 2c in the areas where S-wave velocities are constrained. Thus, the correlation between the velocities observed in the upper crust and the geological units is very good.

The 6.0 km/s P-wave velocity contour is located at a depth of about 15 km, more or less paralleling the 5.5 km/s contour where it exists. It deepens smoothly from the western border of the Eastern Cordillera down to 40 km and then rises suddently by more than 20 km at the eastern border of the cordillera; this jump is also seen in the S-wave velocity model. It is outlined by the crustal seismicity associated with the active faults which mark the boundary between the range and the sub-Andean zone. Thus, the Eastern Cordillera is characterized by an asymmetrical zone having a uniform velocity ranging from 5.5 km/s to 6.0 km/s from the surface down to 35 km.

From the far east of the profile to the western border of the Western Cordillera, the P-wave velocity contours 6.5, 7.0 and 7.5 km/s are parallel and smoothly deepening to the west by no more than 10 km. The coastal zone corresponds to an uplift of high velocity material by about 20 km, well marked on both P-wave as on S-wave velocity models and directly associated with the cluster of earthquakes in and above the slab. Finally, taking the iso-7.5 km/s P-wave velocity line as an estimation of the bottom of the crust, as it is commonly used for this region, we find a Moho depth of 50 km under the western part of the sub-Andean zone increasing smoothly toward the west. It reaches 60 km under the Western Cordillera and then steeply decreases to 40 km under the coastal zone.

These results have to be discussed in the light of previous studies in Peru. The only velocity models deduced from seismic study have been obtained by Cunningham et al. (1986) for the coastal zone and the Western Cordillera in southern Peru, above the area where the dip of the slab changes. These authors observed an increase in the crustal thickness (from about 40 km beneath the coast to 70 km beneath the Western Cordillera), which occurs abruptly above the flat slab but is gradual above the 30° deeping slab. No seismic clustering is associated with the abrupt change. Cunningham et al. propose as a possible explanation that the crust above the steeply deepening Moho in the northwest represents a deformed version of the crust in the southeast. Such an abrupt change in the dip of the Moho is also suggested in central Peru by the gravity anomalies. Fukao et al. (1989) constructed models from gravity surveys along two routes enclosing our area of study. These authors observe that the Moho deepens from 25 km near the coast to 65 km beneath the Western Cordillera. They infer that the Moho in central Peru may be even more steeply deepening.

The uplift of the Moho beneath the coastal zone comes with an even larger uplift of high velocity material for both P- and S-waves. The crustal model compiled by Wilson (1985), passing through Pisco and Huancayo, shows a large mass of high density (3.0 g/cm$^3$, it is 6.6 km/s) material beneath the coastal zone, required to satisfy the gravity high found there. The same feature is found on two other profiles further to the north (as far as Trujillo 10°S). However, this dense rock is apparently not required where the slab dips at 30° and Cough et al. (1981) associated this high density arch-like structure along the axis of the marginal basins with basic rock. The presence of Cretaceous basic rocks in the marginal basin of central Peru is established by Atherton et al. (1983). It stops at about 15°S, where the subduction regime changes from sub-horizontal to normal. Thus, this Cretaceous complex may be responsible for the strong positive gravity anomaly (Atherton et al., 1983). The velocity results of the tomography are quite consistent with those of gravity surveys. The distribution of the hypocenters provides more information about this dense mass. Since a cluster of seismic events is associated with it, it is competent, highly fracturated, and corresponds to an active phenomena. A study of focal mechanisms by Lindo (1993) indicates shortening perhaps associated with the dense material rising. Thus, below the coastal zone, the fracturing of the continental crust allows the uplift of dense materiel above subhorizontal subduction, where the thermal effects associated with partially melted asthenosphere are absent. Moreover, the active uplift of dense material involves the bending of the free surface with extensional stresses and normal faulting activated by gravity (Lindo,1993). Thus, this uplift could explain tectonic features observed in central Peru such as the subsiding of the coastal zone and the presence of normal faulting off shore.

Fukao et al. (1989) observe a steplike change in Bouguer anomaly along the eastern slope of the Eastern Cordillera which they interpret as a sudden change in the crustal thickness at the junction of the Andes and the continent. Such a dicontinuous change in crustal thickness across the sub-Andes is not observed on our results. Only a jump of the iso-6.0 km/s P-wave velocity contour occurs at the eastern border of the Eastern Cordillera, more or less coinciding with the emergence of crustal seismicity. The spatial distribution of this seismicity is not very clear on this general cross-section, due to the projection on a similar plane of distant events. A previous study of seismicity and tectonic deformation in the same area (Dorbath et al., 1991) shows that the activity is concentrated in the eastern foot of the high chain and defines a fault striking northwest-southeast and dipping steeply to the west. The focal mechanisms indicate mainly reverse faulting with hypocenter depths reaching 35 to 40 km. The stress tensor shows a strong compressive regime in a nearly east-west direction. The seismicity on this fault is associated with the uplift of the Eastern Cordillera and the widening of the Andes toward the north-east. Although the whole crust is competent in the western part of the sub-Andean zone, the tomographic study does not show anomalous vertical velocity layering. On the other hand, the velocity structure beneath the Eastern Cordillera differs from the other Andean units. If the higher velocity in the upper crust can be explained by the outcrouping of Precambrian series, the lower crust presents on the contrary a velocity lower than the surrounding areas. Moreover the velocity structure under the Eastern Cordillera is not symmetrical. Our cross section emphasizes the difference between the two Cordilleras. Although the Moho depth is nearly the same under both, the velocity structure of the crust is different. In the uplift of the Andean range, two processes of crustal thickening act jointly: magma intrusion and horizontal shortening. As pointed out by Kono et al. (1989), these two processes have taken place in spatially separated places in the central Andes, where the chain is wide and the two cordilleras separated. The volcanic activity has been very strong in the Western Cordillera from the beginning of the subduction process (about 80 My ago), began to diminish 8 My ago, and disappeared totally during Pliocene time (Noble and McKee, 1977), when the subduction became subhorizontal. The thickening of the crust under the Western Cordillera could then be due mainly to the addition of magmatic material. On the other hand, in the eastern edge of the chain, the crust is pushed strongly to the west by the Brazilian craton, and the shortening occurs by reverse faulting and folding. The asymmetrical structure of the crust under the Eastern Cordillera might reflect the compression of the crust. It is strongly compressed in its eastern part by the craton, the contact occuring on steeply dipping trans-crustal faults, and the deformation gradually diminishes toward the west. The gradual thickening of the crust under the western part of the Eastern Cordillera and its sudden thinning under its eastern part could reflect the convergence process.

**Acknowledgements.** This work is part of a French-Peruvian cooperative project between the Institut Français de recherche Scientifique pour le Développement en Cooperation (ORSTOM) and the Instituto Geofisico del Peru (IGP). Financial support was provided by ORSTOM and INSU. L. Dorbath, G. King, R. Madariaga and an anonymous reviewer provided helpful comments on the manuscript.

## References

Atherton, M.P., Pitcher, W.S. and Warden, V., The Mesozoic marginal basin of Central Peru, *Nature, 305*,303-306, 1983.

Cough, R., Whitsett, R., Huehn, B. and Briceññio-Guarupe, L., Structures of the continental margin in Peru and Chile. In *Nazca Plate: Crustal Formation and Andean Convergence.* (Kulm, L.D., Dymond, D., Dasch, E.J. and Hussong, D.M. eds.) *Mem. geol. Soc. Am., 154*, 703-726, 1981.

Cunningham, P.S., Roecker, S.W. and D. Hatzfeld, Three-dimensional P and S velocity structures of southern Peru and their tectonic implications, *J. Geophys. Res., 91*, 9517 - 9532, 1986.

Dorbath, C., Dorbath, L., Cisternas, A., Deverchere, J. and M. Sébrier, Seismicity of the Huancayo Basin (central Peru) and the Huaytapallana Fault, *J. South Am. Earth Sc., 3*, 21-29, 1990.

Dorbath, L., Dorbath, C., Jimenez, E. and L. Rivera, Seismicity and tectonic deformation in the Eastern Cordillera and the sub-Andean zone of central Peru, *J. South Am. Earth Sc., 4*, 13-24, 1991.

Eberhart-Phillips, D., Three-dimensional velocity structure in Northern California Coast Ranges from inversion of local earthquake arrival times, *Bull. Seism. Soc. Am., 76*, 1025 -1052, 1986.

Fukao, Y., Yamamoto, A. and M. Kono, Gravity anomaly across the Peruvian Andes, *J. Geophys. Res., 94*, 3867-3890, 1989.

Jordan, T.E., Isacks, B.L., Allmendinger, R.W., Brewer, J.A., Ramos, V.A. and C.J. Ando, Andean tectonics related to geometry of subducted Nazca plate, *Geol. Soc. Am. Bull., 94*, 341-361, 1983.

Klein, F.W., Hypocenter location program HYPOINVERSE, U.S. Geol. Survey, Open File Report, 78-694, 113pp, 1978.

Kono, M., Fukao, Y. and A. Yamamoto, Mountain building in the Central Andes, J. *Geophys. Res., 94*, 3891-3905, 1989.

Lindo,R., Dorbath, C., Cisternas, A., Dorbath, L., Ocola, L. and M. Morales, Subduction geometry in Central Peru from a microseismicity survey: first results, *Tectonophysics, 205*, 23-29, 1992.

Lindo, R., Sismotectonique des Andes du Perou central: apport des données sismologiques de haute précision, Thèse de Doctorat, U. Strasbourg, France, 75pp, 1993.

Macharé, J., Sébrier, M., Huaman, D. and J.L. Mercier, Téctonica cenozoica de la margen continental peruana, *Bol. Soc. Geol. Perù, 76*, 45-78, 1986.

Megard, F. Etude Géologique des Andes du Perou central, ORSTOM, Mémoire 86, 310pp, 1978.

Noble, D.C. and McKee, E.H., Comment on Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America, *Geology, 5*, 576-578, 1977.

Pardo, A., Caracteristicas estreucturales de la faja subandina del Norte del Peru, in *Simp. Exploration petrolera en las cuencas subandinas de Venezuela, Colombia, Ecuador y Peru,* Assoc. Colomb. Geol. Geofiss., Bogota, 1982.

Thurber, C.H., Earthquake locations and three-dimensional crustal structure in the Coyote Lake area, Central California, *J. Geophys. Res., 88*, 8226-8236, 1983.

Wilson, D.V., The deeper structure of the Central Andes and some geophysical constraints. In *Magmatism at a plate edge, the Peruvian Andes.* (Pitcher,W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D. eds.), Blackie, 13-18, 1985.

C. Dorbath, EOPG, 5 rue René Descartes, 67084 Strasbourg-cedex, France. (e-mail: cath@mapu.u-strasbg.fr)

(Received June 30, 1995; revised November 6, 1995; accepted November 10, 1995)