Chapter 6

Discussion and Conclusions

As outlined in chapter 3, the present data analysis is based on the teleseismic tomography approach in which a slice from a three–dimensional (3-D) object gives us information about its two-dimensional (2-D) properties. The discussion here presented is based on the interpretation of these "slices" along two profiles in the Central Andes. The described properties and velocity anomalies are restricted to our observation profiles. Therefore, no 3-D interpretation is performed for the present data as it is not possible to speculate about the extension these anomalies have north and south of our study areas at 21°S and 25.5°S.

6.1. The Altiplano anomalies

The results for both *P*- and *S-wave* analysis correlate well and show good agreement with the low- relative velocities at zones of incipient melting (Heit et al., submitted). We see a strong low-velocity anomaly that has the aspect of a layer extending from the west to the east beneath the Central Volcanic Zone (CVZ), the Altiplano plateau and the Eastern Cordillera. This anomaly is flanked by two high-velocity anomalies: one on the western side, corresponding to portions of the interacting Nazca and South American plates (fore-arc) and the other on the east related to the boundary between the Eastern Cordillera and the Interandean, Subandean, Chaco Plain and parts of the non-exposed Brazilian shield (Figure 6.1).

Between the CVZ and the West Fissure (WF), a low-velocity anomaly (QBBS) is observed and interpreted as the origin of fluids associated with the melting in the mid-crust responsible for volcanism in this part of the subduction zone. The ascent of fluids at the border of the Altiplano (Schurr et al., 2003) around the CVZ could be related to the presence of the WF that acts as a natural thermic border between blocks of different temperature. Schurr et al. (2003) found similar results for the Atacama Block in the south of our study area. The rising fluids associated with faults delimiting the QBBS are interpreted by ANCORP Working Group (2003) as the origin of this highly-reflective anomaly.

It is also possible to detect, above the subduction zone and below the Moho at ~68°W, a low-velocity zone interpreted here as partially melted mantle material (see Figure 6.1) possibly generated as a consequence of dehydration of the subducting slab and hydration-melting of a heterogeneous mantle wedge. The progressive heating of the base of the crust should affect the interior of the crust where MASH processes may occur. The MASH standing for Melting Assimilation Storage and Homogenization process was defined by Hildreth and Moorbath (1988) for those zones of possible remelting in the lower crust that produce tonalitic magmas that may differentiate later on their way through the continental crust. Results obtained by other authors also support the presence of this "underplating-like" structure (Lamb and Hoke, 1997). Another possibility for the existence of this anomaly is the presence of hot asthenospheric mantle material that accumulated at the base of the crust

The Uyuni-Kenayani Fault Zone (UKFZ) is a prominent fault system that penetrates the upper portions of the crust and reaches segments of the lower crust (in Mertmann et al., 2001 – see Figure 4.6) and correlates with a low-velocity anomaly observed at 67° W, in accordance with the results presented here and those of other authors (Yuan et al., 2000; ANCORP Working Group, 2003; Haberland et al., 2003). This low-velocity anomaly may be related to the upper limit of a partially melted intra-crustal body that coincides with the prominent layer called the Low-Velocity Zone or Altiplano Low Velocity Zone (LVZ or ALVZ) (Yuan et al., 2000).

A continuation of the ALVZ to the west correlates very well with the presence of the volcanic arc but nevertheless the most important fluid portions appear to be present underneath the Altiplano-EC border region. The anomalies in the west between ~68°W and ~68.5°W could be interpreted as a part of the ALVZ, probably not well defined due to the low resolution of the detected anomaly at depth. To the east, this LVZ appears to reach as far as 65.5° W at the easternmost part of the EC. The extension and presence of this layer of melted material could be explained in accordance with geological features such as the Main Andean Thrust (MAT) (e.g. Allmendinger et al., 1997; Mertmann et al., 2001; Scheuber et al., Submitted). The MAT is a prominent detachment zone located at the border between the EC and the Inter- Subandean provinces and could be affecting the entire portion of the upper crust (see Figure 4.6). We can speculate that the MAT represents a fault zone where changes in lithostatic pressure, temperature and mineralogical variations take place, encouraging, the movement of fluids and melts at greater depths. The role of the QBBS and the ALVZ as a potential decoupling surface and the thermally controlled weakening of the crust were suggested by other authors (e.g. Elger, 2003; Victor et al., 2004; Elger et al., 2005) as a possible subhorizontal shear zone transferring shortening eastward along a decoupling surface. In this case the MAT seems to be acting as a decoupling surface, probably reaching the ALVZ.

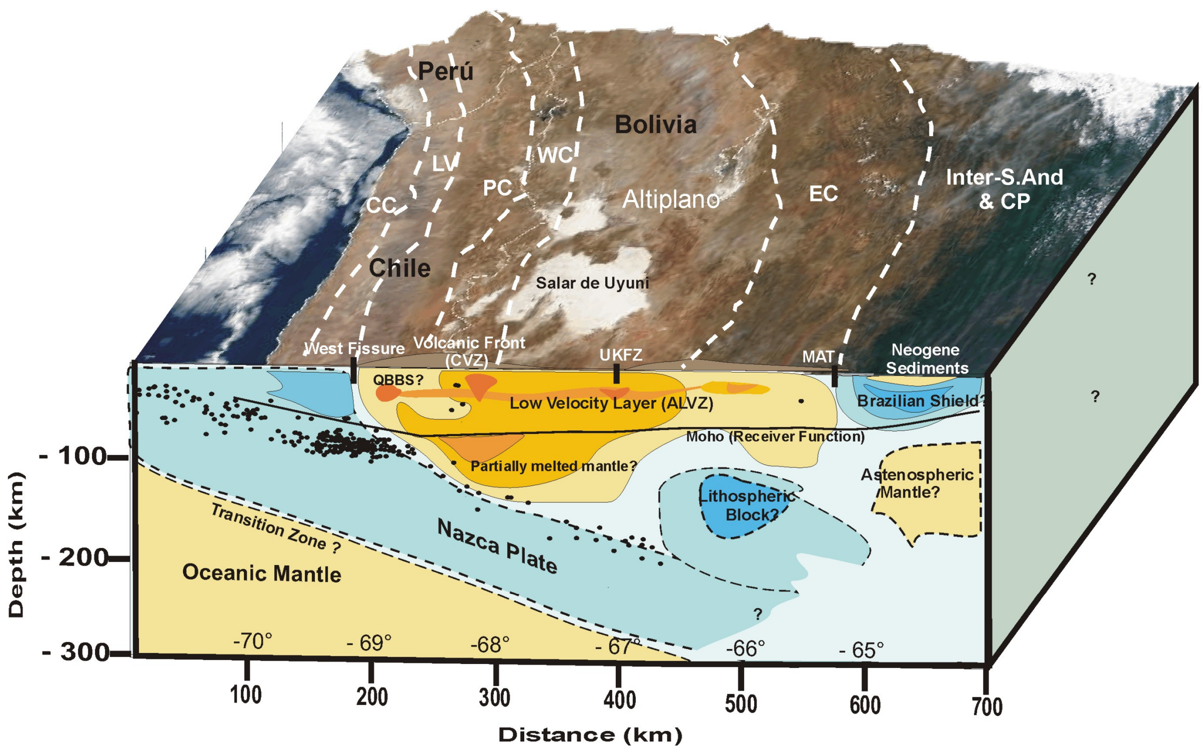

Figure 6.1: Schematic cartoon at 21°, representing the combined P- and S- wave results. Receiver function Moho (Yuan et al., 2000). CVZ: Central Volcanic Zone; QBBS: Quebrada Blanca Bright Spot; UKFZ: Uyuni-Kenayani Fault Zone; MAT: Main Andean Trust; ALVZ: Altiplano Low Velocity Zone. The positive anomaly to the east of the profile has been interpreted as the Brazilian Shield covered by Neogene and Quaternary sediments. The necessity of a Transition Zone above the oceanic mantle, beneath the subducting slab, is deduced from the variations in both P- and S-wave analysis. The Morphological units: CC: Coastal Cordillera; LV: Longitudinal Valley; PC: Precordillera; WC: Western Cordillera; EC: Eastern Cordillera; IA: Interandean; S.And: Subandean; CP: Chaco Plain.

To the east of our study area, we detected the presence of a high relative-velocity anomaly at ~65°W correlating well with the eastern end of the EC. This cold, high velocity unit could be indicating the presence of the Brazilian shield underlying the Interandean, Subandean and Chaco Basin provinces. The presence at this longitude of the Brazilian shield (BS) has also been proposed by other authors (Wigger et al., 1994; Watts et al., 1995; Lamb, et al., 1997; Lamb and Hoke, 1997; Beck and Zandt, 2002) and we believe that the strong Brazilian shield is the basement that supports the eastern morphological units and acts as a backstop that imposed shortening-uplift at the eastern border of the Andean chain.

The evolution of the Altiplano (e.g. Isacks, 1988; Gubbels et al., 1993; Allmendinger and Gubbels, 1996; Allmendinger et al., 1997) started at about 25Ma and went through a series of diastrophic phases that ended with the last uplift and deformation of the plateau at about 9-5Ma (e.g. Allmendinger et al., 1997). The shortening ceased in the Altiplano and shifted to the east, where the Brazilian shield is located, having a counter-effect that could be responsible for the final closing stage of the Altiplano basin and the present-day aspect of the plateau. At the same time, shortening in the Puna plateau continued up to 1-2 Ma (Allmendinger and Gubbels, 1996; Allmendinger et al., 1997) and could be interpreted as a delayed tectonic reaction proving, in this case, that the distance that separated the Puna plateau from the shield may have been bigger than in the case of the Altiplano. In other words the Puna had more time and space to continue its evolution.

However, we are unable to see the subduction of the Brazilian lithosphere as suggested by other authors (Wigger et al., 1994; Watts et al., 1995; Allmendinger et al., 1997; Lamb et al., 1997). The ideas of a subducting lithosphere can not be confirmed with the present data. The strong changes in velocities observed in the upper crust indicate that the transition between the Precambrian Brazilian shield and the morpho-units to the east is abrupt and probably related to major structures in the upper crust.

Many authors interpret Andean uplift as being the result of various interacting factors (Isacks, 1988; Cahill and Isacks, 1992; Wigger et al., 1994; Allmendinger et al., 1997; Lamb et al., 1997; Lamb and Hoke, 1997) but the remaining issues concerning lithospheric thinning, magmatic addition, stability of the plateau and the amount of shortening are still not well constrained. The movement or displacement of the basal parts of the lithosphere is usually proposed to explain the high elevation of plateaus. We believe that the high-velocity anomaly for the Lithospheric Block (100 km -150 km depth) between ~67°W and ~66°W in fact indicates the removal of the base of the lithosphere.

Observations by Hoke et al. (1994) postulated the existence of a thin lithosphere at 20°S for parts of the Eastern Cordillera, to explain the elevated ^{3}He signature in the geothermal systems. The distribution of velocities at the base of the crust and upper mantle suggest that there could be an influx of materials from the asthenosphere that affects the base of the continental crust. Myers et al. (1998) detected the presence of a zone of low Qs (~100), in the same area, that they interpreted as being evidence of near solidus conditions. If this is the case, we will expect to see a shift of the isotherms at this point, probably related to the removal of the high density base of the lithosphere(e.g. Beck and Zandt, 2002). Delamination has already been proposed by other authors (Bird, 1979; Housemann et al., 1981; Kay and Kay, 1993; 1994; Withman et al., 1996; Kay et al., 1999; Schurr, 2000) to explain the absence of a mantle lid and the rapid uplift of the Tibetan and Puna regions that produced stable plateaus over a long period of time. With this type of model, thick and cold lithosphere delaminates from the base of the crust and sinks into the mantle. Our model, presented in Figure 6.1, supports this idea for the Altiplano area. We will discuss about this particular phenomena of *delamination* in section 6.4.

6.2. The Puna anomalies

The Argentine Puna is a key region for understanding the evolution of the Central Andes and particular characteristics that are observed in our profile bring to light some questions concerning the distribution of anomalies in correlation with morphological units.

A similar distribution of positive and negative relative velocity anomalies are observed as for the Altiplano previously discussed area: low-velocity anomalies at both flanks of the Puna and a high-velocity structure on the eastern border of the plateau. However, a difference from the Altiplano is observed in the central part of the Puna plateau where a high-velocity zone is present beneath the area of Salar de Antofalla (SA) (Figure 5.4). This high-velocity anomaly has been interpreted by Goetze and Krause (2002) as a part of the Central Andean Gravity High (CAGH) that is believed to include some of the ultrabasic rocks that form the "Faja Eruptiva Occidental" (an Early Paleozoic magmatic arc) (in Coira et al., 1982).

The CAGH is a dense body about 400 km long and 100 km wide that lies at depths between ~10 and ~40 km (see Figure 5.1). A good correlation is observed for the tomographic section where low-velocities coincide with negative residual anomaly values (under volcanoes) and high-velocities correlate with positive residual anomalies (between the Antofalla and Galan volcanoes). The negative anomalies reflect reduced densities related to fluids and/or melting below the volcanoes. The positive anomalies are related to places where there is no melting/fluids, i.e. between the volcanoes where the Salar de Antofalla is located (Hackney, *pers. comm.*) (Figure 5.4).

The western border of the Puna profile shows what we interpret as the effects of ascending melts that correlate well with the position of the volcanic arc (CVZ). The negative anomalies on the eastern border of the Puna plateau may be related to a trend of volcanic edifices from north to south where some of the back-arc volcanoes are located (e.g. Tuzgle, Qevar, Galan). Observations by Schurr et al. (1999 and 2000) found high attenuation (low Qp) beneath the Tuzgle volcano to the north of our study area that may be included in the before mentioned trend.

The variations that are observed on the eastern side of the profile (border region between the Puna plateau and the Eastern Cordillera) could be related to a ductile lower crust that becomes thicker under a compressional regime. This has been also proposed by Kley and Monaldi (1998) to explain shortening in the north of Argentina and the south of Bolivia. The high velocity in the east is then related to the presence of intrusive and metamorphic Paleozoic rocks that "separate" the plateau from the Eastern Cordillera. The limit imposed by such an old structure could be responsible for the sharp difference in relative velocities west and east of the Luracatao-Tacuil ranges. A difference in temperature for intra-crustal rocks would play a role in the distribution of weak-zones for the ascent of fluids and magmatism that is observed in the back-arc volcanoes (e.g. Tuzgle, Qevar, Galan), where the signature and distribution of volcanic activity on the eastern side of the plateau suggests this situation.

In Figure 5.1 the edge of the Puna plateau is plotted and correlates well with the change in velocities and temperatures observed for both P- and S- waves. The most important back-arc volcanoes are located where these NW-SE trending lineaments intersect the structure that represents the border of the plateau (~NE-SW). Hence, we suggest that deep reaching fault systems associated with shortening, on the east of the plateau, are the conduits for back-arc volcanism concentrated on the eastern margin of the plateau. The fault systems originate in the brittle zone of the crust and propagate to the ductile zone or vice versa.

To explain the shape of the branched anomalies emanating from both clusters of events at ~100 and ~200 km (Figure 6.2), it is necessary to consider the geodynamic location of the southern Puna that represents the final portions of the plateau and is located in a transitional area involving flat-slab subduction.

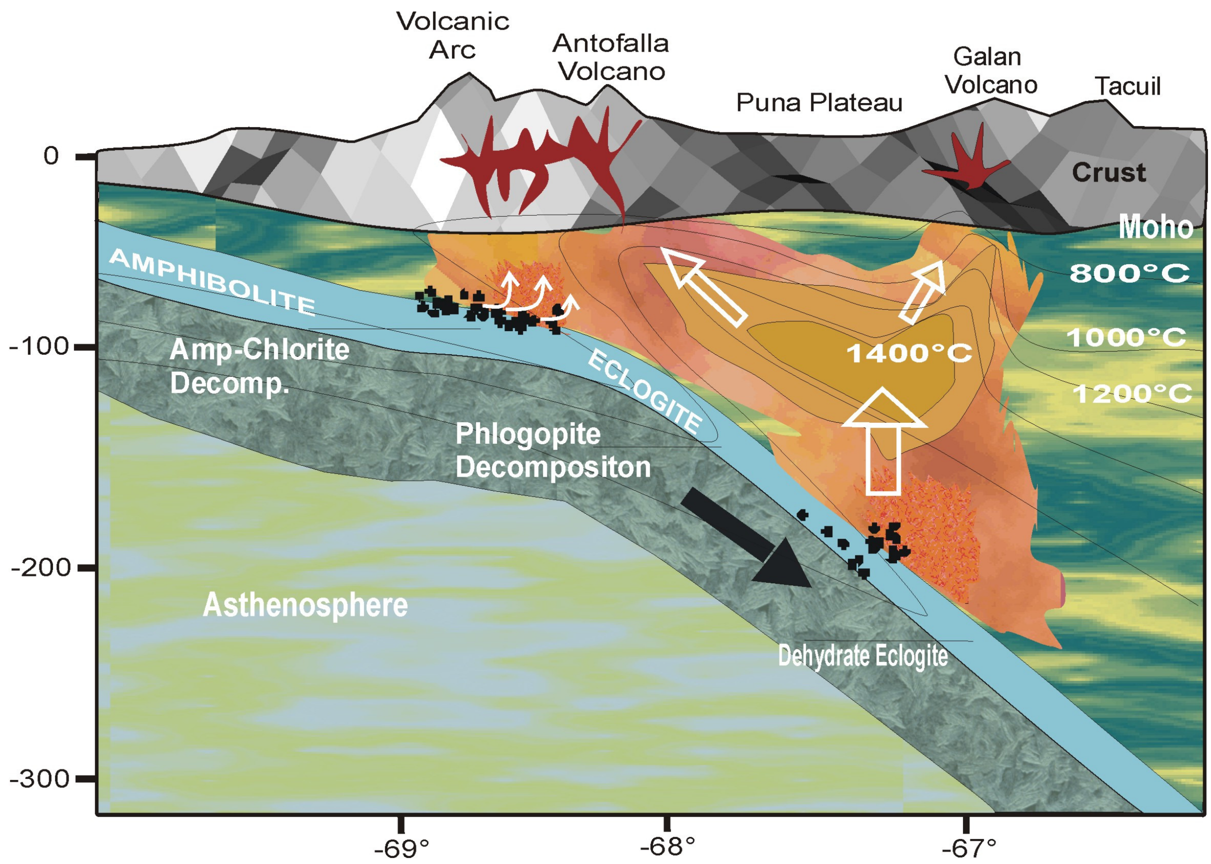

Figure 6.2: Cartoon showing the interpretation of the combined results from P- and S- wave tomography with a branched ascent path for fluids/melts rising from the Benioff zone (clusters of earthquakes) at depths of ~100 km and ~200 km. The anomalies detected in our study suggest that the presence of a highly hydrated oceanic crust is related to the thermal variation caused by an incipient flat area that enables the release of fluids at greater depths than usual. The aspect of the ascent paths can be the associated with a return flow type model for setting the fluids in movement. The position and distribution of isotherms play a very important role, as the high-T° isotherm of 1400°C should induce a flux of melted material towards the volcanic arc or nearby volcanoes as Antofalla.

The reasons for the presence and evolution of a flat-slab are still not well understood. A possible cause usually suggested for explaining this situation is the fact that the Nazca oceanic plate and its lithosphere are relatively young, thin and possibly buoyant. Other results associate the flat-slab areas on both flanks of the Central Andes with the buoyancy of anomalously thick oceanic crust where the aseismic ridges are present (e.g. Carnegie, Nazca and Juan Fernandez (e.g. Cahill and Isacks, 1992; Kay and Abbuzzi, 1996; Gutscher et al., 2000). In contrast, Lamb and Davis (2003) suggest that the flat-slab zone in the Andes is associated with climatic changes along-strike that leave zones of low shear stress where trench-sediment fill lubricates the subduction system, leading to a low angle subduction zone. Meanwhile, zones of high shear stress where the trench sediments are absent lead to normal subduction. In any case, the presence of the flat-slab area to the south of the Puna plateau indicates that the thermal state and heat-flow models for the lithosphere and asthenosphere must be affected by the presence of an unusually shallow oceanic lithosphere. It is also expected that these thermal variations associated with a down-going slab that remains at shallow depths will retain more water that can be later released at progressively greater depths (Figure 6.2).

6.2.1. Fluids and temperature in the mantle under the Puna.

The role of water is important for lowering the melting point of the overlying rocks particularly in those regions where a hydrated slab release water due to phase transformations such as amphibolite to eclogite at 100 km depth. The dehydration of amphiboles is however, a pressure-dependent effect more than a temperature dependent one (Schmidt and Poli, 1998). In all cases, the oceanic lithosphere experience progressive heating and pressure changes due to the subduction and releases the remaining portions of water from the eclogite at a depth of ~200-250 km, where the initial melting should also take place at these depths (Schmidt and Poli, 1998). By successively adding H_2O to the overlying mantle, the partially molten zone must grow upwards. Assuming the position of a high temperature isotherm (e.g. 1400°C), one would expect to see a movement of mantle material towards the volcanic arc. Some of the models of thermal activity in subduction zones show that high temperature isotherms, about 1300ºC, must be located beneath the volcanic arcs and that the depths of the subducting slab surface beneath the volcanic front are in accordance with such isotherms (Schmidt and Poli, 1998).

In the case of the Puna, all andesites and some ingnimbrites always have some mantle material for arc and back-arc volcanoes, such as the Tuzgle and Cerro Galan areas (Lucassen*, pers. comm.*). For explaining the presence of mantle material in volcanic rocks, a thermal anomaly in the asthenosphere is suggested as in Figure 6.2 (e.g. 1400°C isotherm). This thermal anomalous zone would also be responsible for the aspect and distribution of the ascending paths for melts that separate into two branches where the lithosphere-asthenosphere boundary is expected (100-130 km). The presence of a return-flow for fluids leaving the slab at the Benioff zone can be described on the basis of our interpretations and hence, fluids that leave the Benioff zone will ascend parallel to the surface of the oceanic slab, due to physical forces associated with subduction and temperature variations and/or make their way to the surface (or base of the crust) in a more or less vertical manner (Figure 6.2).

Is it possible that water remains in the eclogite that reaches the final step of dehydration at depths of ~200 km? This depends mainly on the thermal stability of the hydrous minerals. At a given temperature the mantle does not need too much fluid to trigger large amounts of melting. The suggested position of the isotherms showed in Figure 6.2 should reflect the behaviour of the peridotite at the lithosphere-asthenosphere boundary (LAB). Small amounts of water in olivine will change the physical properties of this major component in mantle rocks, in the way that changes in viscosity and ductile flow may occur at much lower temperature than indicated in most standard subduction zone models (Schmidt and Poli, 1998). Having near-solidus or melted material in the mantle does not necessarily imply having a defined large scale flow pattern, since thermal or density inhomogeneities can also define local small scale flow patterns that transmit the movement to other portions of the mantle where the final transformations deliver fluids that ascent towards the base of the crust. The plateau will therefore be affected by a thermal influx from below, induced by water movement that lowers the melting point of the neighbouring rocks. These thermal anomalies in the asthenosphere keeps the plateau elevated over a long period of time without the necessity of delamination. On their way to the surface, the melts and fluids can form melting zones in the upper crust below ~20 km. Information about the thermal expansion of the Central Andes might still be needed to confirm this statement.

It has been suggested that only a small portion of the water initially contained in the subducting lithosphere will escape from the slab in the depth range suitable for magma formation and that a large portion of the water will be dehydrated at relatively shallow levels (Schmidt and Poli, 1998). This effect should occur in the oceanic crust at depths of ~80 km, depending on the temperature of the subduction zone, where certain phase transformations allow minerals to retain some percentage of water (e.g. minerals as lawsonite-eclogites in cold subduction areas and zoisite-eclogite in warm subduction areas). If any of these minerals, according to their stability, is able to reach deeper portions of the asthenosphere, then a part of the water will also be able to descend to greater depths and will be lost for arc volcanism (Schmidt and Poli, 1998). In our case, for the Puna plateau we speculate that the water delivered to the system from the subducted slab at about 200 km is related to the decomposition of minerals that takes place between the phlogopite decomposition depths (~150 km) and the final dehydration of the eclogite (lawsonite and phengite phases) at depths of ~300 km.

In cold subduction zones, where the stability fields of some hydrous phases and serpentine overlap, water will be transported to greater depths and will not be available for arc magmatism (phlogopite decomposition in Figure 6.3), playing maybe a role in the evolution of back-arc volcanism as observed in the Central Andes. In relatively warm subduction zones fluids will be delivered from the dehydration of serpentine (Schmidt and Poli, 1998).

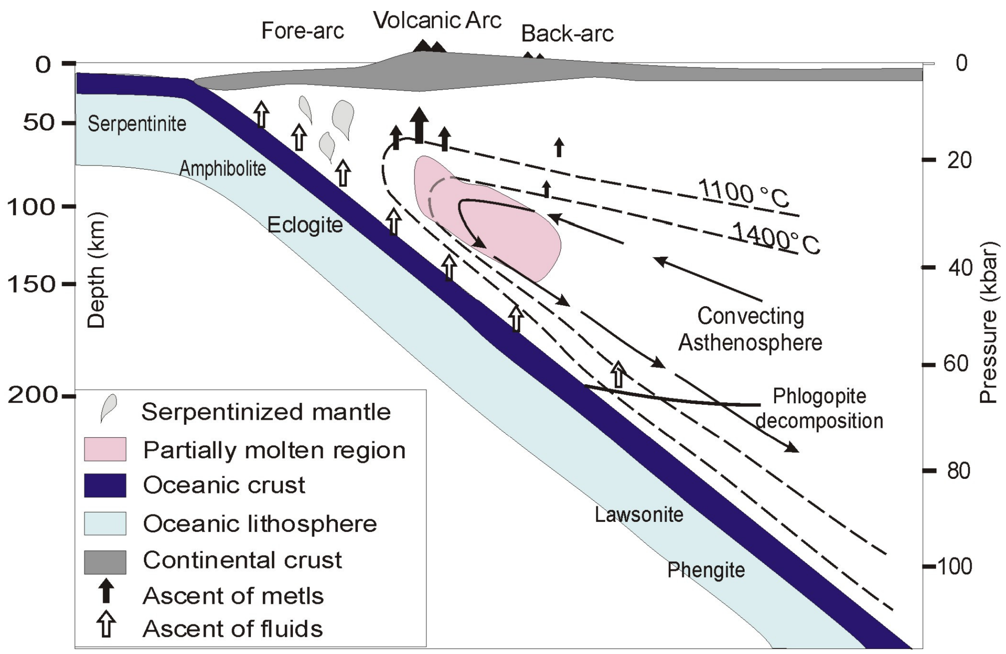

Figure 6.3: this diagram after Schmidt and Poli (1998), shows a possible model for the formation of a volcanic front. Open arrows indicate the rise of fluid, short solid arrows indicate rise of melts. Long arrows indicate flow in the mantle wedge. The diapirs represent a partly serpentinized mantle. Phase transformations at 200 km depth are directly related to the thermal state of the subducting plate.

As shown in Figure 6.3, the dehydration from peridotite and oceanic crust occurs at almost any depth to ca. 150–200 km, with water generally available above the subducting lithosphere. The region in the mantle, where partial melt is indicated, will therefore play a significant role in the movements of the melts towards the crust, so that the volcanic arc will develop above those areas where the amount of melt is sufficient to give rise to arc magmatism.

Continuous water release to depths greater than 200 km due to the breakdown of lawsonite and phengite will ensure the formation of hydrous peridotite in the mantle wedge via the direct release of fluids towards the overlying mantle (Poli and Schmidt, 1995) as suggested in our cartoon (Figure 6.2). A return-flow type model for the Andes, based on our observations seems to be the most possible case for explaining the movement of fluids towards the crust, while other authors suggest, that a corner-flow model as the driving mechanism for the type of tectonics can be discarded as it induces an excessive stress regime on the overriding plate (Husson and Ricard, 2004); without a significant corner-flow below the volcanic fronts mantle wedge temperatures would rapidly cool down and arc magmatism would cease (Schmidt and Poli, 1998). Both examples reflect how little these mechanisms are understood.

Romanyuk and Rebetsky (2000) analyzed the density and rheology of the down going oceanic crust of the Nazca plate and the importance of fluids in the weakening of mantle minerals. They suggested a model where the subducting plate is composed of two layers, one mainly of basalt and the other of gabbro being the later denser than the first (2.8-g/cm ^{3} basalt and 2.95-g/cm^3 gabbro); with density increasing with plate depth. The Nazca plate mantle is believed to consist of harzburgite, which is gradually replaced by spinel-lherzolite at greater depths. Where the oceanic crust experiences dehydration and metamorphism, the basalt–eclogite transition provides significant amounts of fluid that are released by the oceanic plate. This fluid leads to the serpentinization of the mantle and to the "wet" melting of the mantle wedge peridotites that form the volcanic arc (Davies and Stevenson, 1992). These ideas also support our interpretation cartoon from Figure 6.2.

The analysis of noble gases (e.g. Helium) released as bubbles in geothermal springs or contained in olivine crystals of mafic rocks, provide a clue as to the volatile components origin. This simple method illustrates the degassing signature of ^{3}He and shows that there are two main factors that govern the helium degassing rate and isotopic evolution in a mantle undergoing whole mantle convection: (1) the ocean crust formation rate, and (2) the concentration of helium in the mantle (Class and Goldstein, 2005). Therefore, a distinctive signature for volatiles that have a mantle or crustal origin can be obtained.

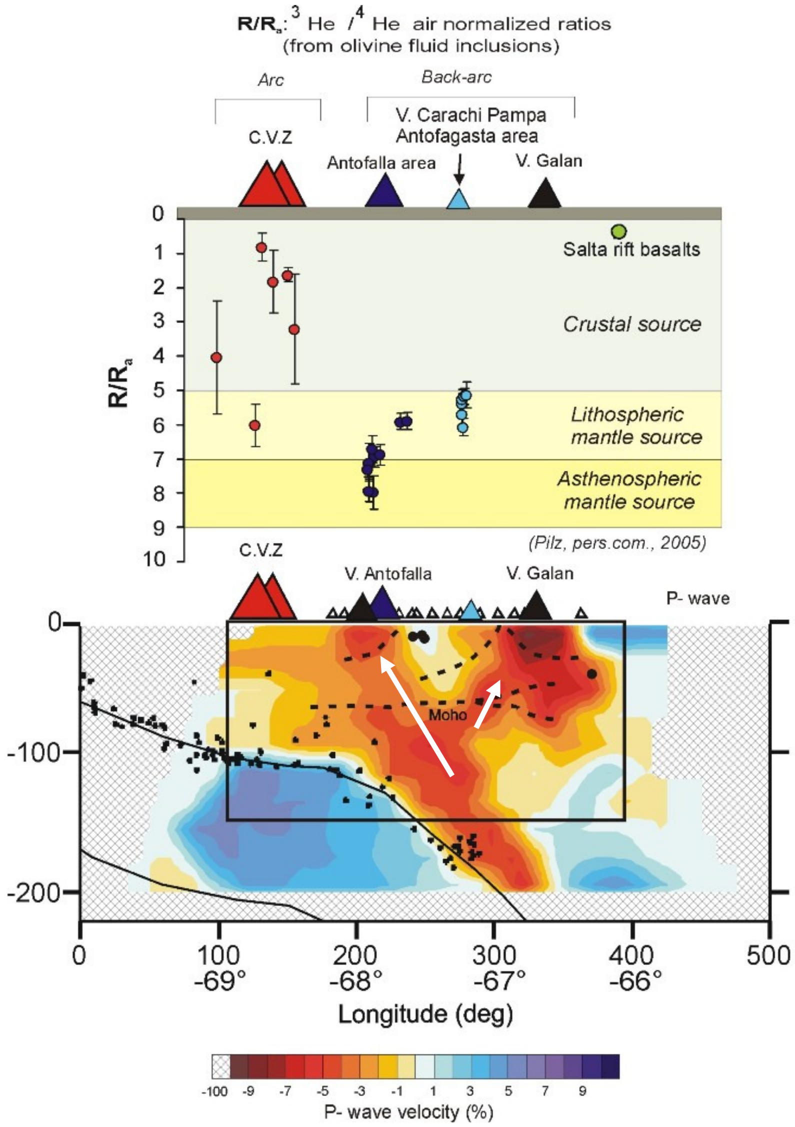

Figure 6.4: ^{3} He / ^{4}He analysis from olivine inclusion (Pilz, pers. comm., 2005) and one of our teleseismic tomography sections (P- waves). Long white arrow: fast ascent of hot melts with little contamination; Short white arrow: slow ascent of colder melts with higher crustal contamination. Rectangle shows the area of comparison. Moho topography and crustal structures from receiver function data.

By comparing our results with ^{3} He / ^{4}He air normalized ratios, we see a strong correlation between the plotted values for the area of the volcanic arc that suggest a crustal source of the volatiles with our interpretations of the observed anomalies related to the group of earthquakes at ~100 km. Towards the east of the CVZ, ^{3} He values for the Antofalla volcano show a deeper source located in the asthenospheric mantle, with a few values indicating a shallower origin. East of the block from Salar de Antofalla, where the Carachi Pampa volcano is located, the ratios indicate lithospheric to crustal sources. This evidence can be used to reinforce our previous explanation that the branched-shape anomaly represents melt ascent paths towards the volcanic arc that in some cases reach the surface with less crustal contamination caused by the influence of unusual high temperatures (1400°C isotherm) or by favourable fault zones guided ascent. To the east, the materials reach the base of the crust at lower temperatures (~800°C) and experiences greater crustal contamination (Figure 6.4)

We are not able to observe clear evidence for delamination or its remnants in the analyzed section. Receiver function data (RF) for our profile does not show crustal variations that could be interpreted as being affected by such a process. The Moho depth remains more or less constant between 60 and 65 km. The eastern side of the RF profile shows a double layered Moho interface, with one surface above and the other below 65 km. Since this is the border area of the Puna plateau, we interpret this as a place where the shortening of the Eastern Cordillera affects portions of the entire crust.

6.3. The Role of Delamination in Andean Evolution

The role of lithospheric delamination has been proposed in recent years to explain the rapid uplift and thin lithosphere with the associated magmatism in the Puna Plateau (Kay and Kay, 1993; Kay et al., 1994; Schurr, 2000; Schurr et al., 2003). This process remains controversial as it involves the removal from lower portions of the crust and upper-most lithosphere that "detach or delaminate" and sinks into the asthenosphere. The process of delamination must be related to density increase in the mafic lower crust (basaltic crust at the Moho discontinuity), changes in mineral composition and thermally-anomalous materials.

When modelling the South American subduction zone, Sobolev and Babeyko (2005) presented lithospheric delamination as being triggered by the gabbro-eclogite transformation in a thickened continental lower crust. Mechanical failure of the sediment cover at the margin of the plateau, where the BS is located, should be the leading process in the mechanical weakening of the overriding plate during tectonic shortening. Active delamination, in their opinion, impedes corner-flow by delaminated material that intensifies shortening as a result of increased coupling between the plates and the mechanical weakening of the overriding plate. In their model, the trench (roll-back) has also been migrating westward, contributing to the delamination and shortening that they interpret as a major factor on controlling Andean orogeny.

In our study we are able to see the effects of some sort of delamination in the Altiplano plateau (Figure 6.1) where we interpret the presence of a lithospheric block for explaining high-velocities at depths of ~150 km. The shape of the high-velocity anomaly detected below the Altiplano can be supported by the model showed in Figure 6.5. For the Puna region, at 25.5°S, we are not able to see a similar effect that could be interpreted as delamination of the lithosphere.

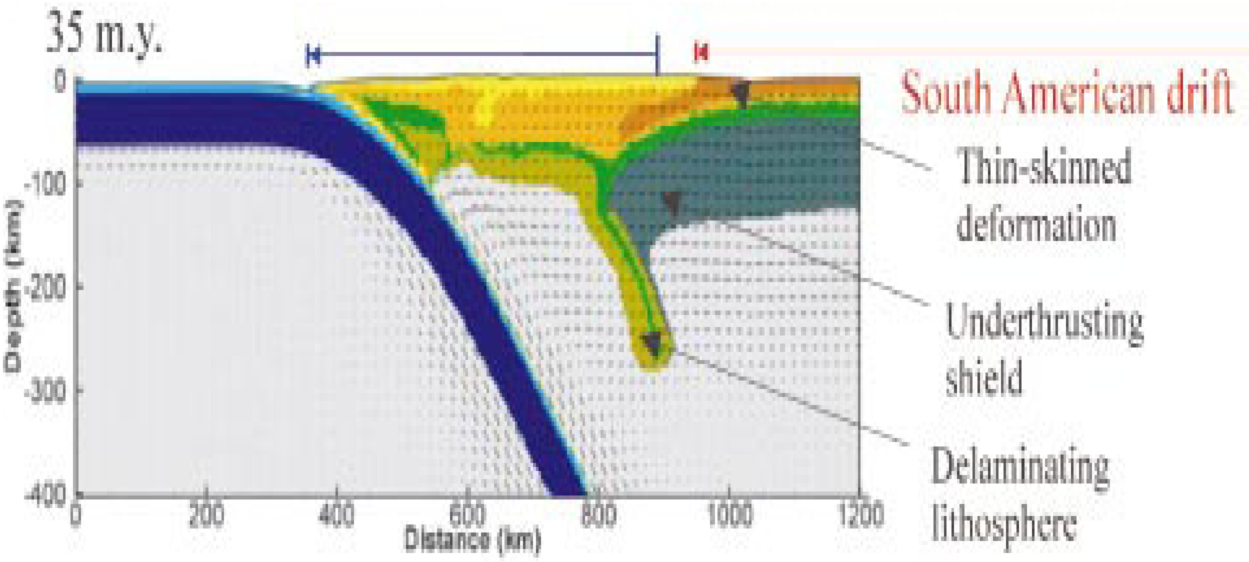

Figure 6.5: After 35 Ma the evolution of the Central Andes show an important drift of the continental plate to the west and effects of delamination in the back-arc area. An incipient underthrusting of the shield can also be seen (Sobolev and Babeyko, 2005)

For considerations about the thickness of the crust for both plateaus –a thicker Altiplano crust was usually proposed for differentiating between Altiplano and Puna- we use the receiver function images obtained from present data as explained in the next section.

6.4. Receiver Function data and the tomographic images

The receiver function (RF) method is based on the fact that teleseismic waves passing through a seismic discontinuity convert part of their energy to S- waves which arrive at the station after the onset of the P- wave within the P- wave coda. By analyzing the energy released during the conversion from P- to S- waves it is possible to detect some discontinuities beneath the station, for example the Moho. The different aspects of the receiver function method and the quantification of the data is outside of the focus of this work. For more information the reader is referred to the literature cited at the end of this work (e.g. Yuan, 1999; Yuan et al., 2000; Yuan et al., 2002).

The results from teleseismic tomography are compared with receiver function images obtained from the analysis carried out in our group at the GFZ using the same data set. For the first time, the final results of the Moho discontinuity at ~21°S (Altiplano) and ~25.5°S (Puna) are obtained from our data and show a good correlation with tomographic images and previously obtained RF images (Yuan et al., 2000; Yuan et al., 2002).

In the case of the Altiplano, it is possible to see a uniform Moho discontinuity that goes from depths of ~75 km beneath the volcanic arc, thins beneath the Altiplano to ~65 km and thickens again below the Eastern Cordillera region to 75-80 km. To the east the Moho flattens to depths of ~50 km between 65°W and 64°W. The subducted slab can be clearly seen down to a depth of ~120 km, while at greater depths it is only inferred by the position of some interfaces that are not properly defined. The presence of a low-velocity anomaly in the upper crust can be seen as a zone of horizontal extension along a large part of the profile (Figure 6.6, centre and bottom in the case of the Altiplano).

For the Puna area, the Moho discontinuity beneath the plateau remains flat at about ~60 km. We are not able to see the portions of the Moho beneath the volcanic arc due to the limited extent of the profile. To the east, below the border between Puna and the Eastern Cordillera, the Moho is not well resolved, but a thickening beneath the Galan volcano area at ~67°W is expected, as in the case for the Altiplano profile. The slab cannot be detected in this RF image. Meanwhile, the presence of internal structures in the crust remains a subject of discussion, but nevertheless, there is a good correlation between those structures and the limits of low- and high- velocities in the tomographic images (Figure 6.6).

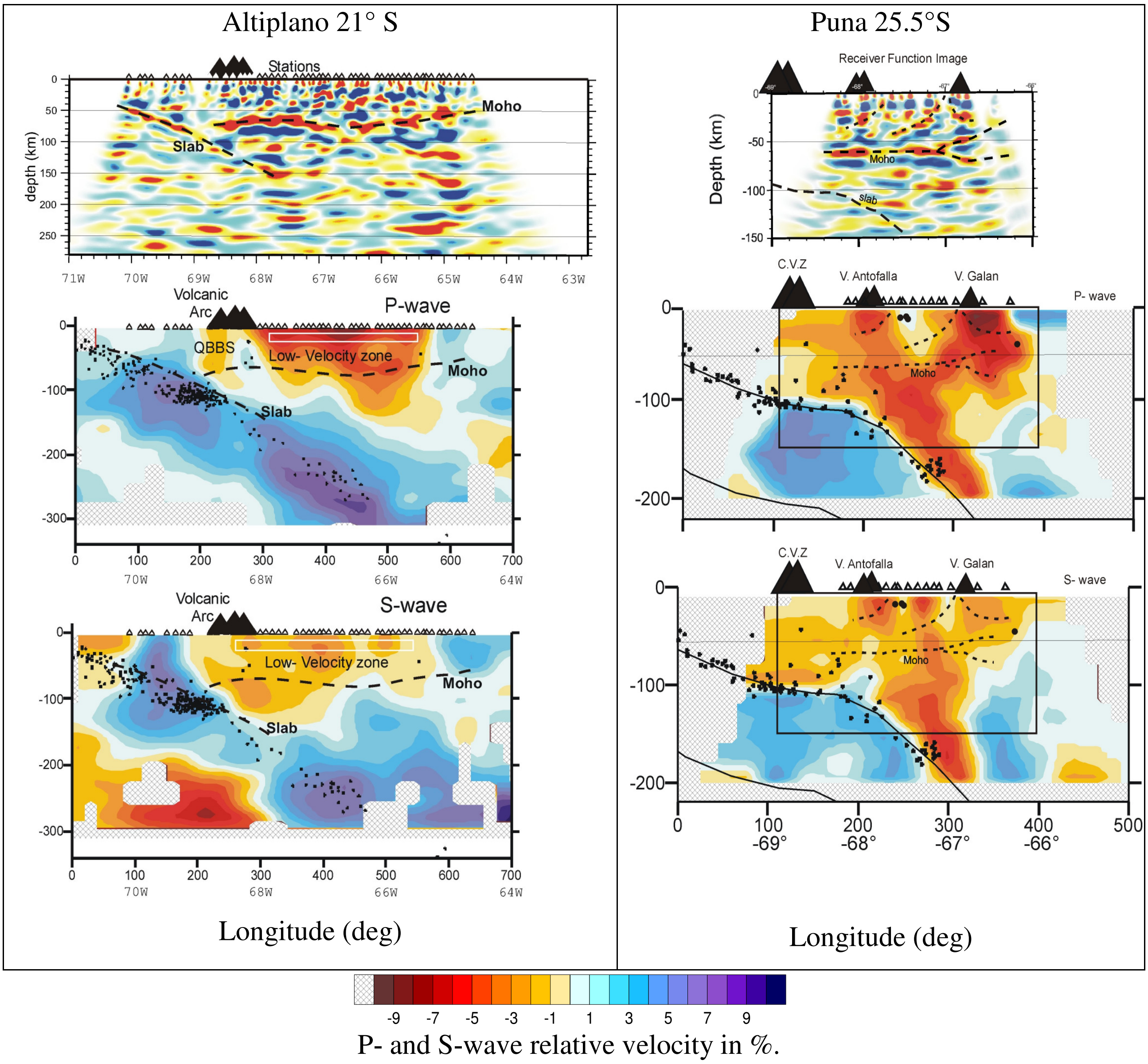

Figure 6.6: Top: Results from receiver function (RF) data, indicating the Moho and oceanic slab discontinuities for both Altiplano and Puna. Centre and bottom: P- and S- wave teleseismic tomography with information from RF. The white rectangle for Altiplano highlights the position of the low-velocity zone beneath the Altiplano.

By comparing receiver function data from both plateaus (Figure 6.7), particularly information about the Moho topography and depth, it is possible to see that there are small differences between both profiles at 21°S (Altiplano) and 25.5°S (Puna). In the case of Altiplano the Moho beneath the central part of the plateau can be seen at ~65 km depth while for the Puna the depth of the same feature is some ~60 km. Internal structures detected by receiver functions suggest a very similar response for both plateaus along major crustal faults (e.g. MAT).

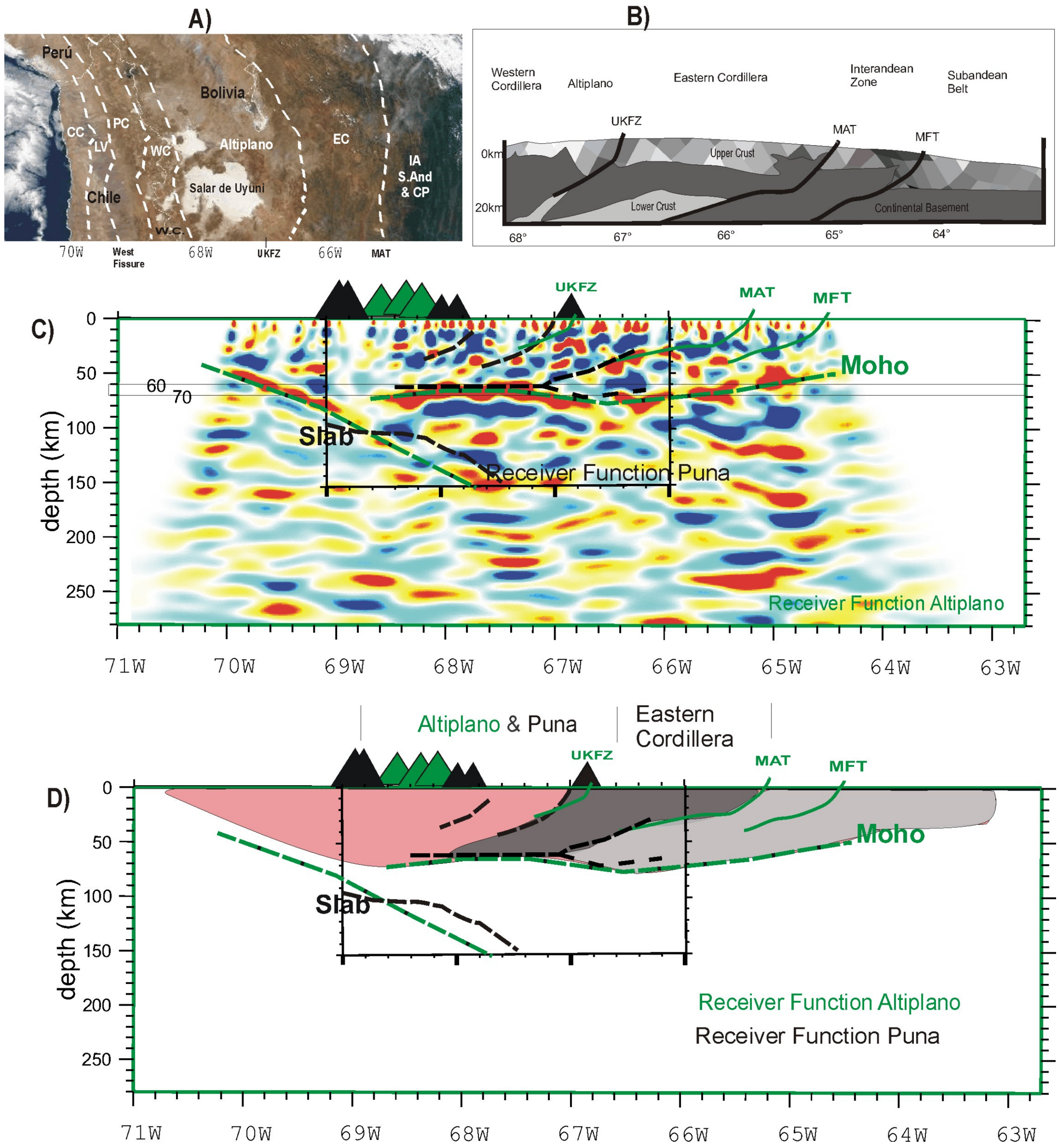

Figure 6.7: A) The Morphological units: CC: Coastal Cordillera; LV: Longitudinal Valley; PC: Precordillera; WC: Western Cordillera; EC: Eastern Cordillera; IA: Interandean; S.And: Subandean; CPlain: Chaco Plain along Altiplano profile at 21°S. UKFZ: Uyuni-Kenayani Fault Zone; MAT: Main Andean Trust. B) The MAT (Main Andean Trust), along the section at 21°S, which affects the entire structure of the upper crust and reaches portions of the lower crust Simplified after Mertmann et al., 2001 and Scheuber et al., submitted. C) Receiver Function Altiplano showing the crustal structures and the Moho (green lines) plotted together with Puna RF data (black lines). D) Diagram integrating RF crustal discontinuities with slab for Altiplano (green line) and Puna data (black line and small rectangle in figure). The different colours (pink and greys) are suggested structural blocks that can be identified and reflect the similarities along both profiles.

The distribution and aspect of the Moho topography shows changes that occur at both latitudes that are probably related with the limits between morphological units on the surface. In Figure

6.7D the position of some crustal discontinuities reflect the similarities between both plateaus, and may be indirectly, the influence such structures had in the geodynamical history of the Central Andes. There is a possible relationship between magmatism in the upper plate and deformation that suggests that, the formation of the Altiplano Plateau was initiated by magmatically-controlled thermal weakening of the crust (Elger, 2003 and Elger et al., 2005). The presence of this thermal effect may indicate that the major deformation structures are related to the low-velocities in the upper-crust detected in our study.

It is not possible to see variations in Moho topography that suggest a variation in lithospheric thicknesses in both cases. From the tomographic images it is possible to see that Puna profile presents a more "hot" material that can be interpreted as astenospheric influx that could affect the thickness of the lithosphere but evidence supporting this idea is still missing. If this is the case, the effect of a thinner lithosphere beneath the Puna can be questioned as our data suggests. In the future, S-wave receiver function analysis of the present data should clarify the position of the LAB as this is the only method that can provide information about the LAB discontinuity.

6.5. Puna and Altiplano: a space problem?

This section discusses the results obtained by the tomographic analysis where it is possible to observe the presence of the Precambrian Brazilian shield on the east of the Altiplano profile. The correlation between the position of the BS and mayor structures limiting this unit should have played a role that influenced the plateau formation and evolution.

Why did a plateau developed in the Central Andes? Why is the history of both the Altiplano and Puna so different in many ways and similar in others? If we compare topographic sections, as noted by other authors in the past years (Whitman et al., 1996; Allmendinger et al., 1997), both Puna and Altiplano differ in the distribution and shape of the basins and in their altitude above sea level (Figure 6.8). The Altiplano presents a more or less regular surface with no major changes in topography inside the plateau. The Puna on the other hand shows a more irregular aspect and displays large topographic variations delimiting the internal closed basins.

One possibility is that plateau growth is a result of the Brazilian craton compressing the weak lower crust under the Altiplano from the east (Beck and Zandt, 2002), while from the west there is the ENE movement of the Nazca plate. As the Altiplano was compressed, the remaining portions of the plateau in the south (where the Puna is located) were left as a triangle-shaped block that experienced additional uplift, and has been active for a longer time than the northern parts. As discussed before, the uplift in the Altiplano started ~25 Ma and had stopped by ~10-9 Ma, while in the Puna this process continued until ~2Ma.

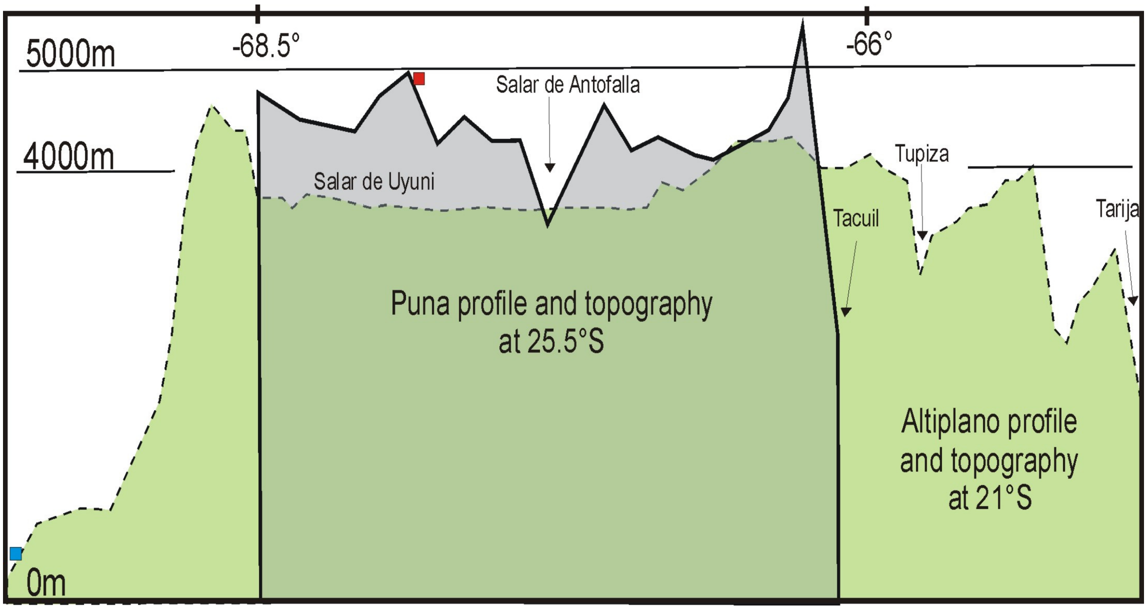

Figure 6.8: On the green background the topography along the profile at 21°S (dashed line). The topography of the Puna profile (grey background and thick bold line) shows the differences in altitude and the irregularities along ~25.5°S. The red square shows the position of the highest station of the project (PC04, ~4800 m). The blue square shows the position of the lowest station (RF01~360 m).

One question that comes out when studying the Andes is related to the different dipping angles of the Nazca oceanic plate during the geodynamical history of the subduction zone and its variation with latitude from the Colombian Andes in the north to the south of Chile. The differences in subduction angles may be associated with thickness variations of the oceanic plate (with a possible direct impact in the confluence rate of the plates) and particularly with buoyant subducting lithosphere (Gutscher, et al., 2000; van Hunen et al., 2002a,b). In this context, the position of the Altiplano-Puna plateau is emplaced in a particular geographic position with respect to other major morphological units in the South American continent, e.g. old, cold Precambrian Brazilian Shield (Figure 6.9). We may therefore, establish a parallel between the existence of flat-slab regions where those fragments of old, cold units are not present or at least have been displaced by the presence of some mayor lineaments (i.e., TBL: Trans-Brazilian Lineament in Cordani and Sato, 1999).

Along with the TBL in the south it is possible to find another major suture or lineament present in the area between the Guyana and Brazilian shields, which is located beneath the Amazonian Basin, where the presence of a proto-rift has been suggested (Burke and Lytwyn, 1993). At the same time, those parts of the Andes that developed large-scale volcanic activity, mountain uplift and a plateau, in the case of Bolivia and northern Argentina, are located in areas where subduction has found opposition, in some way represented by old stable regions such as the BS (Figure 6.9). The presence of the remaining parts of the shield and the lineaments that separate its portions could also be responsible for the differences in morphology between the Altiplano and Puna plateaus.

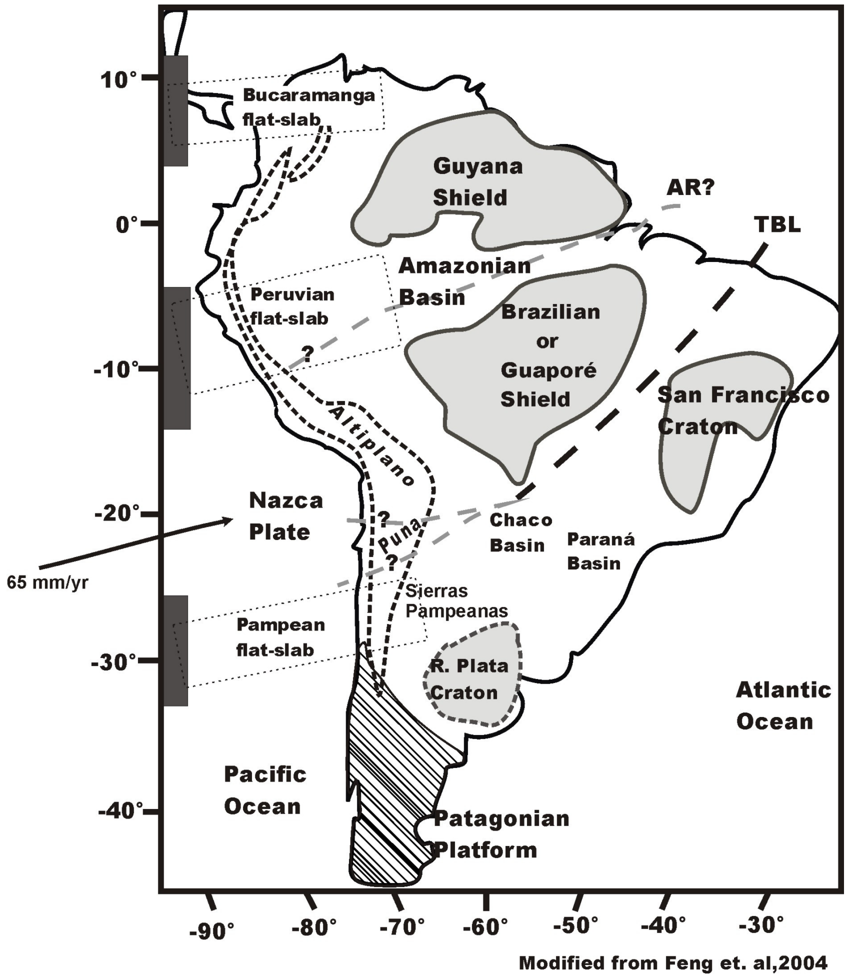

Figure 6.9: Bold Rectangles and related grey dashed rectangles show the correlation between the flat-slab regions and the absence of the exposed Precambrian cratons. Note the position of the Altiplano plateau flanked by units with opposite directions of movement i.e., Nazca Plate (Angermann et al, 1999) and Brazilian Shield (Guaporé). North and south of the BS, large sedimentary basins developed covering continental suture zones. TBL: Trans-Brazilian Lineament (in Cordani and Sato, 1999) could be responsible for structural differences in the plateau (Altiplano and Puna). The grey dashed line at the Amazonian basin is the AR? (Amazonian Rift) as suggested by Burke and Lytwyn (1993). Black dashed area from north to south represent the Andean chain with the position of the plateaus marked

As discussed in Section 6.1.3, shortening and uplift in the Central Andes are responsible for two characteristic features: the thrust belt (Eastern Cordillera) and the thin-skinned fold-thrust belt (Subandean) on the eastern side of the Altiplano plateau. According to Sempere et al. (1997), this compressional regime started around ~90 Ma and has been migrating eastward since then (Jordan et al., 1997; Sempere et al., 1997). This could be related again with to presence of the Brazilian shield.

In addition to the previous statements, we suggest that the presence of the Brazilian Shield is responsible for the change in block rotation (clock- and counter clockwise) (Beck, 1998) from the Arica bending line between 19°S and 23°S, that has been proposed as the symmetry axis of the Central Andes (Gephard, 1994). This is also the trend of the movement of opposite direction on either side of the zone of maximum shortening (Isacks, 1988, Kley and Monaldi, 1998).

An argument that opposes this theory is that the subduction angle of the oceanic plate in the central Andes has changed many times, going from normal to flat and again to normal, while the velocity of convergence and subduction has also changed. Presently, the subduction takes place at a convergence rate of 8.5cm/yr (Angermann, 1999) and the angle, on the basis of seismicity, is ~30° (Engdahl et al., 1995). The subduction angle could therefore have changed due to velocity variations of the subducting plate due to a retarded effect of mineral transformation.

Other authors (e.g. Lamb and Hoke, 1997; Lamb and Davies, 2003) suggest that climatic change during the Cenozoic is responsible for the unusual width and height of the central part of the Altiplano in the Andes and that the onset of hyper aridity along the west coast of South America is responsible for the lack of a thick sediment layer that would otherwise lubricate the interface between the Nazca and South American plates, as is the case in the southern Andes. Erosion, sediment deposition and climate controlled sediment starvation cause high shear stresses in the western part of continental South America where the resulting deformation created a high plateau without collision.

A higher Puna plateau with a more irregular surface could be indicative of less space for shortening and uplift than the Altiplano, probably induced by differences in lithospheric structures or the absence of a thick layer of sediments. This idea is demonstrated by Babeyko and Sobolev (2005) who suggested that different shortening modes for the Central Andes (pure and simple shear accompanied by thin or thick-skinned tectonics) might be controlled by strength of the foreland uppermost crust and by temperature of the foreland lithosphere. In their model the sediment cover plays a key role in tectonic evolution and support the idea of Allmendinger and Gubbels (1996) that the thick layer of Paleozoic sediments at the Bolivian foreland might be responsible for the simple shear mode of shortening north of 23°S. For the southern Puna at 24°S, the weak sedimentary layer is almost absent and the deformation pattern is described as simple shear with thick-skinned foreland tectonics, similar to the Santa Barbara system east of Puna at ~26°S. A higher temperature of the leading edge of the Brazilian shield south of 24°S than north of 23°S implying a thermal effect for explaining the thick-skinned tectonics in the Puna foreland has been suggested. In our models (for Puna and Altiplano anomalies) we are able to detect changes in relative velocities and therefore interpret that they are related to these temperatures variations. The case of the Puna having a different thermal state than the Altiplano can be then linked to the distribution of deformation that resulted from shortening variations in the crust for both plateaus.

Considering these ideas, it is possible to explain the evolution of the plateau as result of the confluence of many factors that determine the characteristic aspects of the central Andean region, but there is little that we can say about which factors played the larger role.

6.6. Conclusions

The profile of stations at 21°S tomographic analysis confirms previous observations about the presence of a low-velocity layer in the upper crustal portions of the Altiplano (Altiplano Low-Velocity Zone) and shows that this zone may extend towards the Eastern Cordillera where a similar low-velocity body has been detected. A possible relationship between this low-velocity zone and the anomalies observed to the west beneath the volcanic arc can also be interpreted from tomographic results. To the east of the profile, the transition to the Brazilian Shield seems to be abrupt and is governed by the presence of major fault systems, such as the Main Andean Thrust. The tomographic images do not show any evidence supporting underthrusting of the Brazilian lithosphere as suggested by some authors. The presence of a high-velocity block at a depth of ~150km can be associated with the detachment or delamination of lithospheric material. The crustal thickness from receiver function analyses shows a Moho that deepens beneath the volcanic arc probably related to partially melted material accumulated at the base of the continental crust. In the centre of the profile, beneath the Altiplano plateau, the Moho thins to ~65km and thickens again beneath the Eastern Cordillera. Towards the Chaco Plain in the east, the Moho thins to ~ 40km. It is not possible to observe variations in Moho topography that could suggest variations in lithospheric thicknesses

The Puna profile data indicate the presence of a two-branched pathway for the ascent of fluids and melts. The western part is clearly related to the fluids that reach the volcanic arc (Antofalla volcano) while the other coincides with the presence of back-arc volcanoes along the eastern border of the plateau (e.g. Galan volcano). The tomographic images suggest that the thermal state of the Puna plateau is different from that of the Altiplano, indicating the presence of colder asthenospheric material beneath the Altiplano and hotter beneath Puna. This thermal effect can be interpreted as responsible for the higher elevations of the Puna plateau. Beneath the Salar de Antofalla, a high-velocity block with seismic activity is interpreted as part of the old and cold Paleaozoic magmatic arc (Faja Eruptiva de la Puna occidental). The presence of this block is may be responsible for the distribution of volcanic activity localized at both sides of this anomaly.

Receiver function data provide the best (up to date) images of the Moho topography for both profiles (along 21°S and 25.5°S), indicating that there are no big differences between Altiplano and Puna crustal thicknesses (~65km and ~60km respectively). The internal structures of both plateaus, in terms of the distribution of fault systems and detachment zones in the internal portions of the crust, seem to be more similar than originally thought.