Chapter 5

The Puna Plateau

5.1. The Central Andes at 25.5°S and the Puna Plateau

5.1.1. Puna and Altiplano plateaus

As already discussed in chapter 4, the Altiplano and Puna high plateaus in the central segment of the Andean orogen represent local variations in the geological history of uplift, amount of shortening, magmatism with related volcanism and changes in topography (e.g. Reutter et al., 1988; Allmendinger et al., 1997). These changes in characteristics enable a differentiation between Altiplano and Puna and can be recognized approximately north and south of the parallel 22°S, near the Bolivian-Argentine border, coinciding with the trend of an impressive fault system (Lipez lineament NW-SE – Figure 5.1).

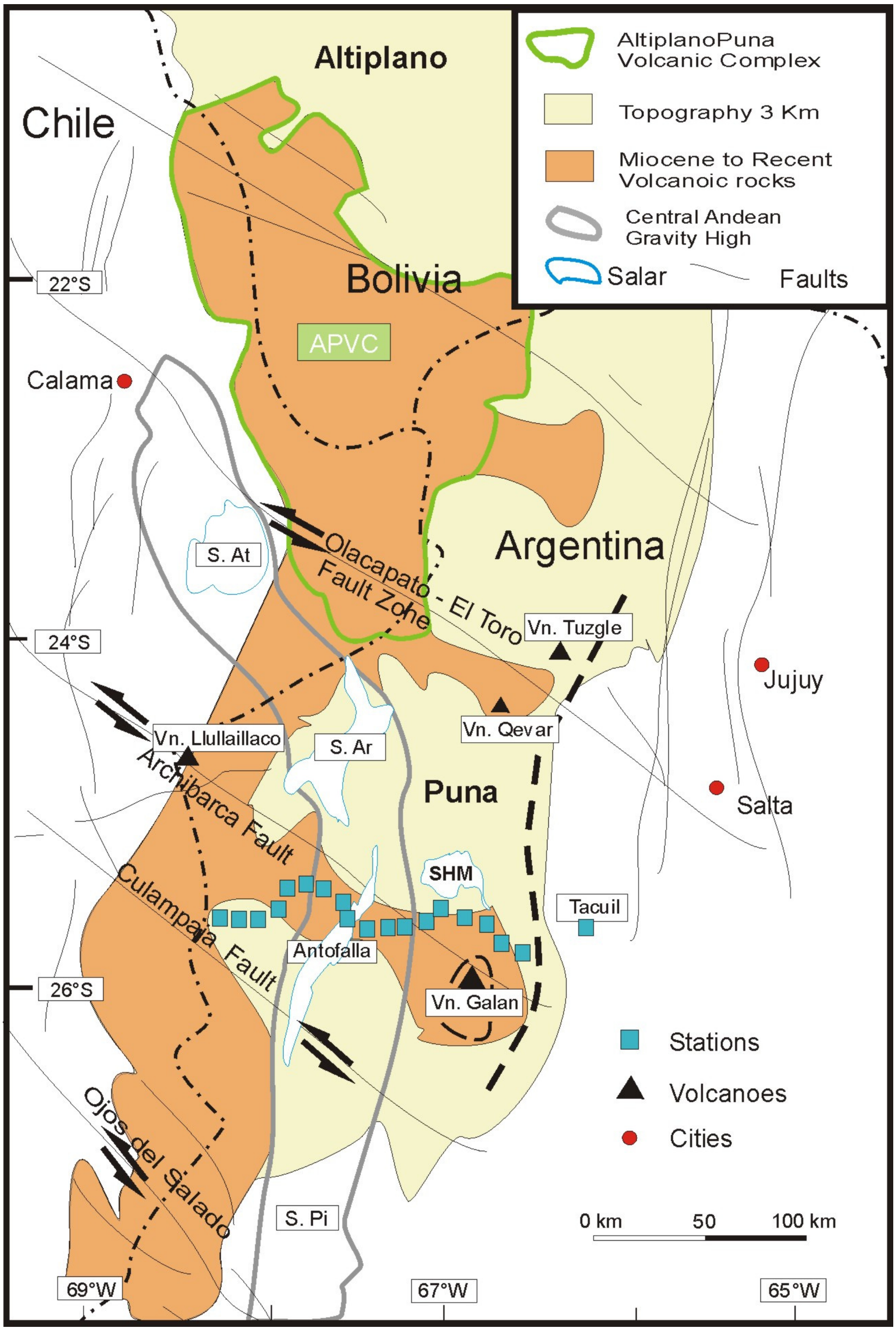

Figure 5.1: Map showing the principal lineaments of the Puna (Lipez, Olacapato-Toro, Archibarca, Culampaja and Ojos del Salado). The position of our profile in the southern end of the Puna at ~25.5°S is also plotted (blue squares). Green area: (APVC) Altiplano-Puna Volcanic Complex from deSilva (1989); S. At: Salar de Atacama, S.Ar: Salar de Arizaro, SHM: Salar del Hombre Muerto, S. Pi: Salar de Pipanaco. Gray line: encloses the region of the Central Andean Gravity High (Goetze and Krause, 2002). Black dashed line represents the eastern border of the Puna plateau where an apparent change in the temperature of the rocks in the crust is detected (Ocloyic fold and thrust belt?). (Modified after Riller et al., 2001)

The presence of the Altiplano-Puna-Volcanic-Complex (APVC by de Silva, 1989) between both parts of the plateau at the triple junction between Argentina, Bolivia and Chile has been suggested to be responsible for late Miocene to Recent ignimbrite deposits (Figure 5.1.). The APVC overlies a large part of the ALVZ which is suggested to behave as a migmatite zone in the middle-upper part of the crust (Trumbull, pers. com.). Below the APVC, the presence of a very low-velocity layer at depths between 10 and 30 km has been interpreted as a sill-like magma body named the Altiplano-Puna Magma Body (APMB) (Chmielowski et al., 1999; Zandt et al., 2003) Along the APVC area, a transition based on the chemistry of Tertiary rocks, topography and structural changes north and south of 22°S that differentiates Altiplano and Puna has also been proposed by many authors (e.g. Allmendinger et al., 1983 and 1997; Allmendinger and Gubbles, 1994; Coira et al., 1993; Whitman et al., 1996) (Figure 5.1).

The volcanic arc has been migrating to the east since the Jurassic to its actual position at the CVZ. The magmatism changes form andesitic to dacitic with some expressions of shoshonitic and basaltic rocks in the southern end of the Puna. Large volumes of silicic ignimbrites are associated with the terminal volcanic activities represented by collapse of calderas (e.g. La Pacana in Chile, Cerro Galan and Cerro Blanco in Argentina).

In general, the plateau is limited to the west by an active volcanic arc consisting of Tertiary volcanic rocks (Western Cordillera, WC or Central Volcanic Zone, CVZ) and to the east by an active westward verging thin-skinned foreland thrust belt (deformed Paleozoic rocks of the Eastern Cordillera).

By comparing the elevations of the entire plateau from 20°S to 26°S it is also possible to detect some differences. The Puna plateau in Argentina, for example, has an average altitude of ~4.2km above sea level and is about 1 km higher than the Altiplano plateau of Bolivia and Peru (~3.2km). In contrast to the Altiplano, the units that follow the Eastern Cordillera to the east are in addition, the Santa Barbara System and the Sierras Pampeanas thick-skinned foreland provinces.

The crustal structure of the Puna was studied using the seismic refraction method by Wigger et al. (1991). The crustal thickness, another feature for differentiating the Puna and Altiplano plateaus, was estimated by receiver function analysis as being ~ 55-60 km underneath the northern Puna, which is 10-15 km thinner than that of the Altiplano (Yuan et al., 2000). These results agree with those of Götze et al., (1994) who postulated that the crustal thickness of the Puna should be less than that of the Altiplano, based on regional gravity models obtained for the Central Andes.

The presence of the Central Andean Gravity High (CAGH) (see Figure 5.1) has been also detected along the Cambro-Ordovician magmatic arc (Faja Eruptiva de la Puna Occidental in Coira et al., 1982). The extension of this anomaly, interpreted from isostatic residual values, represents an elongated ~N-S area that covers the city of Calama in Chile, the Salar de Arizaro and Salar de Antofalla salt pans and ends in the Salar de Pipanaco depression in the southern Puna (Goetze and Krause, 2002).

Young mafic magmatism in the southern part of the Puna has been one of the arguments proposed for distinguishing between both plateaus, supported by processes related to the delamination of a part of the lower crust and upper mantle, to explain a higher topography and a thinner lithosphere (Kay et al., 1994; Whitman et al., 1996).

The Puna also differs from the Altiplano in the fact that it does not present a well developed thin-skinned thrust belt to the east and has a more irregular surface (Allmendinger et al, 1997). The fact that the Puna surface is so irregular (Figure 6.5 -Chapter 6) has to do with a system of small closed basins that started to develop during Neogene times. A model presented by Riller and Oncken (2003) suggest that crustal segmentation of the Puna plateau into rhomb-shaped areas where the closed basins developed is responsible for localized shortening and surface uplift at both margins of the plateau. Such segmentation has produced a series of strike-slip faults limiting the structural blocks of the plateau. Some of these faults still show some seismic activity (Schurr et al., 1999; Schurr, 2000), mainly concentrated along the eastern border of the plateau and the OTL (Olacapato-Toro Lineament, displayed in Figure 5.1 as Olacapato-El Toro Fault Zone), a prominent shear zone that extends from the coast of Chile to the southeast in the proximity of Salta city in Argentina, that has registered historical seismicity on several occasions.

The addition of accretion wedges and marginal basins has not been an influence on the orogenic evolution. The uplift related to compression and crustal shortening of the Puna plateau was attributed to the increasing convergence rate and flattening subduction of the oceanic Nazca Plate, which started around ~20Ma and continued until the late Pliocene.

The plateau of NW Argentina has an early Paleozoic metamorphic basement composed by different sedimentary and magmatic units. Following compression and uplift of the plateau, a series of Palaeogene sediments deposited in the Puna foreland orogenic basins.

The filling of evaporite basins in the Puna Plateau occurred between the Neogene and the Holocene. In the Neogene magmatic arc two main types of volcanic activity developed. The andesitic and dacitic volcanism was continuous since about 20 Ma until the Quaternary and produced many of the well-known strato-volcanic complexes of the southern Central Andes. In the Central Andes between ~25 Ma and ~18 Ma the volcanic centres are well distributed across the arc and back-arc regions north of 22°S; south of 25°S the centres are confined to the arc front. Based on this distribution and the absence of volcanism during this time some authors have suggested a gap in magmatism between ~22°S and ~25°S (Coira *et al.*, 1993; Allmendinger *et al.*, 1997). The volcanic centres dating from Middle Miocene to Pliocene, are distributed over the whole Argentine Puna. The onset of large-volume dacitic to rhyolitic ignimbrite eruptions began in the late Miocene (ca. 12 Ma) (de Silva, 1989; Coira et al., 1993).

5.1.2. The Southern Puna

In the case of the Puna plateau, there are also some internal differences that enable a separation in northern and southern Puna. The northern Puna region, between ~22°S–24·5°S, is structurally separated from the southern Puna, ~25°S–28°S, by the NW–SE-trending Olacapato-Toro (OTL) lineament. North and south of the OTL, a system of parallel shear zones, considered to be zones of lithospheric weakness (Allmendinger et al., 1983, 1997), are responsible for a series of strato-volcanic edifices that account for the largest volcanoes of the Puna plateau (e.g. Llullailaco, Socompa, Ojos del Salado in the volcanic arc and Antofalla, Tuzgle in the back-arc) and include the highest volcanoes on Earth. In between these lineaments, back-arc volcanic activity is still present but not with prominent edifices (Figure 5.1).

The most significant north-south structural change in the back arc, along the eastern border of the plateau is the termination of the thin-skinned Subandean belt near 23°S. This change correlates with the end of Paleozoic basins and the superposition of the Upper Cretaceous rift basins in the foreland south of 24°S (Allmendinger et al., 1983; Allmendinger et al., 1997), where the thick-skinned Santa Barbara System replaces the thin-skinned Subandean belt and south of 26°S where northern Sierras Pampeanas replace the Eastern Cordillera. Most of the Puna basement consists of Paleozoic rocks that exhibit several phases of deformation usually known as reworked pre-Cenozoic continental crust.

In the area of our profile at 25.5°S the Salar de Antofalla (SA), located to the east of the Antofalla volcanic complex, forms an elongated basin of salt filing deposits with important topographical differences with the rest of the plateau. Before the crustal thickening and uplift of the Central Andes during Neogene times, the SA area was part of the Andean foreland. The Southern Puna may have been part of a large region with low relief in which only minor sedimentation occurred until the Late Eocene. Later, the SA was incorporated to the Andean orogen, to the east of the magmatic arc of the early Tertiary subduction system. Pilger (1984) suggested the subduction of lower density, young and hot oceanic crust, to explain the flat subduction angle during this time, although today this idea is not completely accepted.

An increase in the convergence rate from ~28.3-25.8 Ma (Somoza, 1998) normal to the trench due to reorganization of the Pacific plate suggest a subduction angle less than 30° at this time (Kay et al., 1988 and 1991; Mpodozis et al., 1995). Between ~20-12 Ma regional compression and the onset of volcanic activity dominated in the Antofalla area. According to geochemical data, the melts erupted through a crust with a thickness of at least ~40 km. This indicates that regional uplift was in progress. For the period between 12 and 4 Ma, a series of intra-arc basins developed due to tectonic shortening while the strato-volcanic centres were still active. During Late Miocene to Pliocene times (< 9 Ma), alluvial fan sedimentation was interrupted by a period of relative tectonic quiescence. Between 6 and 4 Ma, the relative tectonic quiescence was accompanied by the eruption of intraplate-like basaltic andesitic magmas at monogenetic centres.

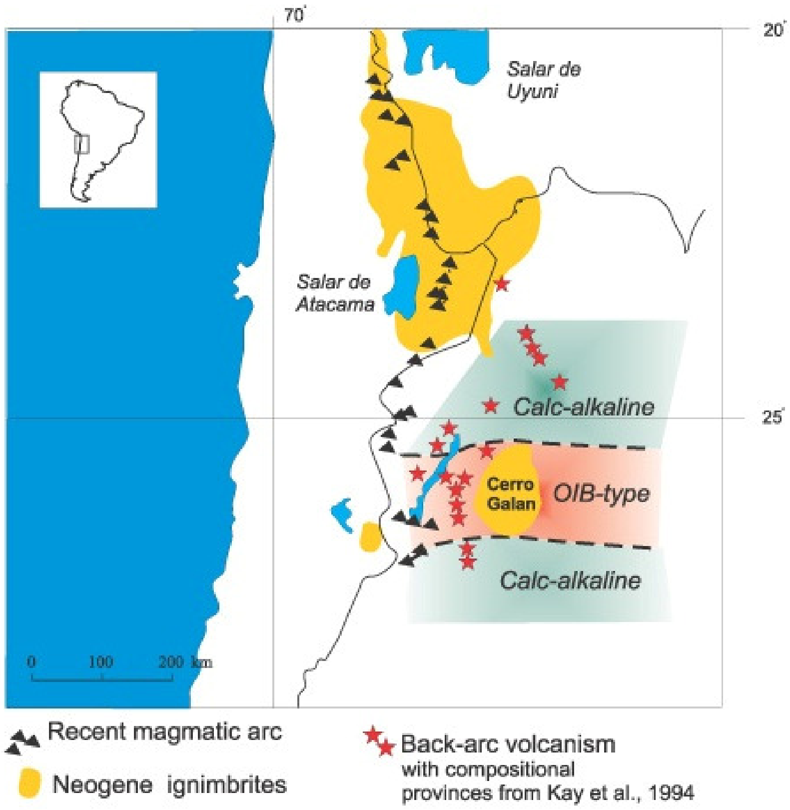

Figure 5.2: Map showing the distribution of volcanism in the Puna plateau where the lithospheric delamination is supposed to be strongest (beneath the OIB region) according to Kay et al. (1994).

The occurrence of intraplate-like volcanism has been usually associated with crustal heating, the generally higher topography and a thinner lithosphere of the Southern Puna compared with the Altiplano. From 4 Ma until the present, crustal thickening and uplift resulted from renewed contractional movements. Da Silva (1989), Francis et al. (1989) and later Kraemer et al. (1999) interpreted the large-volume ignimbrites as reflecting large-scale crustal melting due to crustal thickening and heating by basaltic intrusions The presence of the Cerro Galan collapse Caldera and its associated large ignimbrite deposits have been suggested by Kay and Kay (1993) and Kay et al. (1994) as a region for understanding delamination processes (Figure 5.2). Among the characteristics the authors consider, are the combination of factors such as uplift, stress regime and the different magmatic signatures compared with the rest of the plateau in the Galan area.

A difference in the composition of the extrusives parallel to the CVZ as used in figure 5.2 to explain basic type-volcanic deposits in the back-arc area may reflect the chemical characteristics of the fluids involved in upper-crustal melting. For the ascent of melts, Schurr et al. (1999 and

2003) proposed a model for the Puna at ~24°S where the ascending path for the fluids causes melting of the overlying mantle and suggested a strong horizontal component for melt ascent (Figure 5.3). This model was obtained from local tomography data from the PUNA 97 (Puna Untersuchung Nord Argentinien) seismological project. Results that correlate the Puna magmatic evolution with associated shortening on the Eastern Cordillera, and their influence on crustal thickness variations from north to south are still required to understand the present state of the plateau within the context of deformational processes.

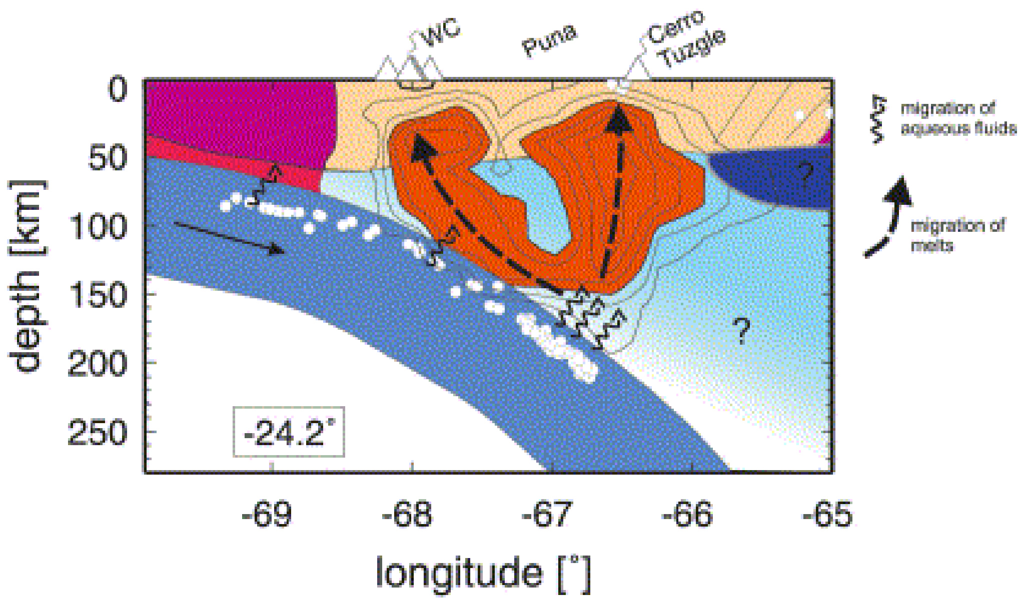

Figure 5.3: Fluid movement based on interpretations from of the PUNA project data. The ascent paths for melts that migrate through the mantle are shown in orange. Note the bifurcation of fluid patterns from the cluster of earthquakes between 150 and 200 km depth (From Schurr, 2000 and Schurr et al., 2003).

5.2. Results and Observations for the Puna plateau

The position of the profile and the most important salt-lakes (Salar) and volcanoes are displayed in Figure 5.4, along with the results from the inversion of P- and S- waves. Other tomographic results for P- and S- waves are presented in vertical and horizontal sections in the Appendix The morpho-structural units involved in the study area of the Puna plateau and its bordering regions are: the volcanic arc (CVZ) to the west at ~68° and the Eastern Cordillera (EC) to the east at ~66° (Tacuil).

The images obtained for the Puna plateau at 25.5° S show a strong low-velocity anomaly that virtually "emanates" from depths near ~200 km. This negative anomaly reaches the Moho discontinuity (~60km obtained with receiver function) as a single zone and then bifurcates or separates into two branches; one at the position of the CVZ and the other where the border of the plateau to the east is represented by some back-arc volcanoes (e.g. Galan). At the depth of the Moho discontinuity a broadening of the low-velocity anomaly is detected. The difference in temperature and composition below and above the Moho where accumulation of melts may occur, should cause heating of the lower crust leaving internal zones of weakness. The weakened crust affected by ductile deformation is then affected by the movement of partially melted material towards the surface.

From west to east it is possible to describe the following anomalies (Figure 5.4).

1- The region between 69°W and 68°W west of the Salar de Antofalla (SA) is characterized by low-velocities for P- waves. The anomaly region is located in the upper crust.

The low-velocities area; west of the Antofalla volcano seem to extend towards the main volcanic arc on the west and can be associated with a group of earthquakes at a depth of ~100 km. Unfortunately the upper part of this region can not be well resolved due to the limitations of the profile, but nevertheless, the shape of the anomalies suggests the presence of materials with low-relative velocities ascending towards the volcanoes of the CVZ (not displayed in Figure 5.4). An additional strong low- velocity anomaly in the upper crust at a depth of ~20 km is interpreted as part of a melting zone beneath the Antofalla strato-volcanic complex.

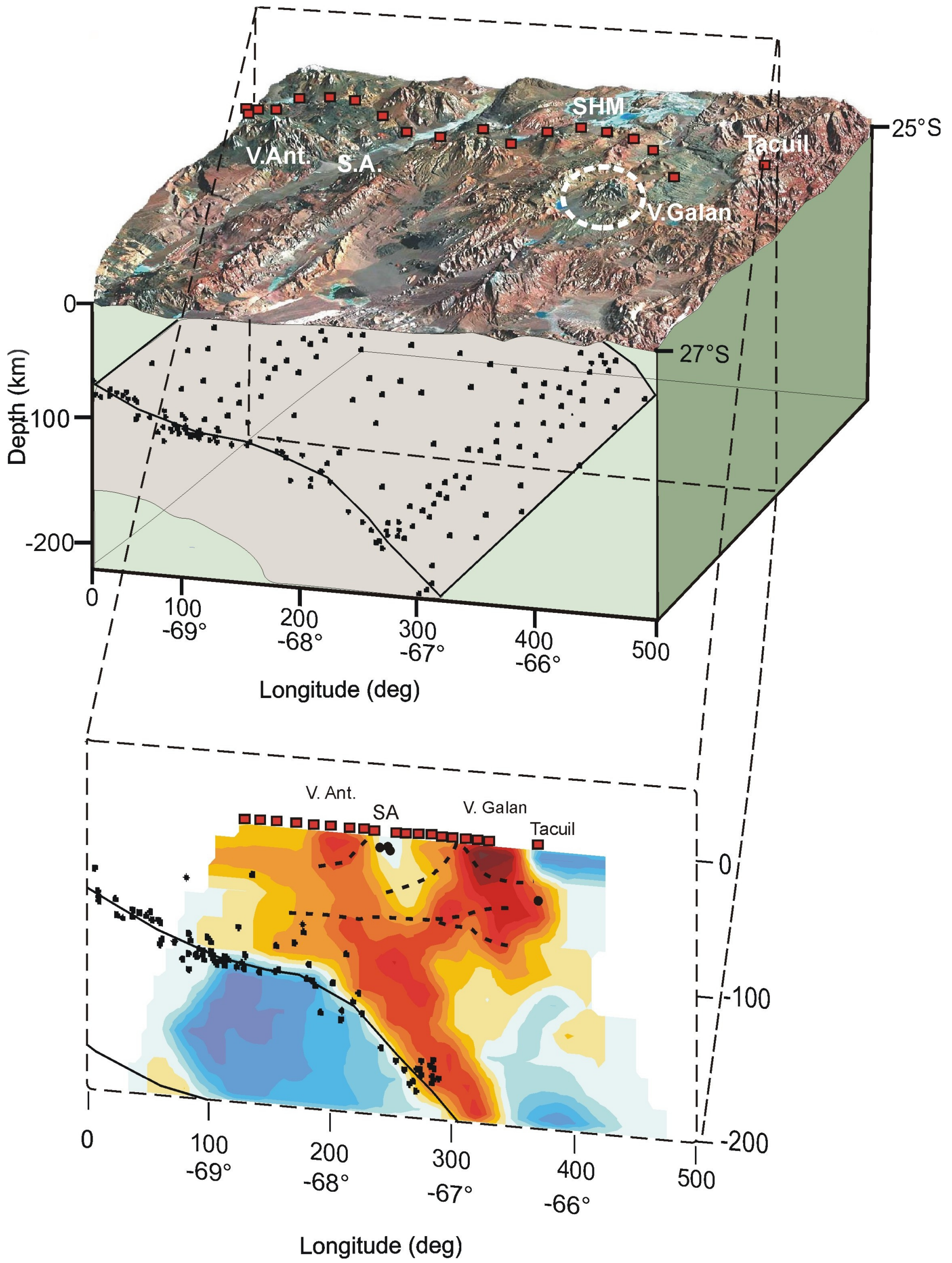

Figure 5.4: Block diagram showing the distribution of stations at 25.5°S (red squares) and the position of the slab inferred from earthquake hypocenters (Engdahl et al, 1995). The tomographic section represents P- wave results. Receiver function Moho and crustal discontinuities are plotted along this section. V. Ant.: Antofalla Volcano, SA: Salar Antofalla, V. Galan: Galan Volcano.

2- The Salar de Antofalla is related to a high-velocity anomaly in the upper crust and is located between low-velocity anomalies. The long shape of this NE-SW trending salt pan, the topographic differences between the surface of the salar (~3400m), the surrounding ranges (~4000m) -*see Figure 6.5, chapter 6* - and its narrow basin (max. 10 km) could be related to deep reaching structures delimiting not only the presence of this block but inducing the divergence in velocity anomalies at this latitude. The location of intra-crustal seismicity is also plotted and interpreted as consequence of deep reaching structures (faults) limiting the salar.

3- A low-velocity anomaly is located at ~67°W and seems to correlate well with the presence of the Cerro Galan volcano on the surface. The extension and depth of this anomaly is probably interrupted due to the presence of a high-velocity structure on the eastern border of the plateau.

4- The high-velocity anomaly beneath the Tacuil station, at the eastern end of the profile, is interpreted as the western ranges of the Eastern Cordillera, which seem to indicate a sharp change in velocities. The limit between positive and negative velocity anomalies west and east of this range is clearly represented by Palaeozoic granitic and metamorphic units in the upper crust, with higher velocities to the east of 66.5W.