4. Geological setting – the Andean plateau

The Tibetan Plateau in the Himalayas and the Altiplano-Puna Plateau in the Central Andes are the only two examples of active plateau orogens worldwide. They are both characterized by a very large area that is hardly internally deformed and has an elevation of more than 4000 m for the Andean plateau and more than 5000 m for the Tibetan example. These flat areas are bounded to both sides by mountain ranges which have an extent of several hundred kilometers.

The Andean Cordillera stretches all along the western margin of South America, but only from ~17°S-27° the mountain range is developed as a plateau orogen. The Altiplano lies north of 24°, the Puna south of it; together they form the Altiplano-Puna plateau, in the following referred to as the orogen (excluding parts of the Andes that have no plateau). In this thesis, I focus on the Central Andean Altiplano (Fig. 4.1).

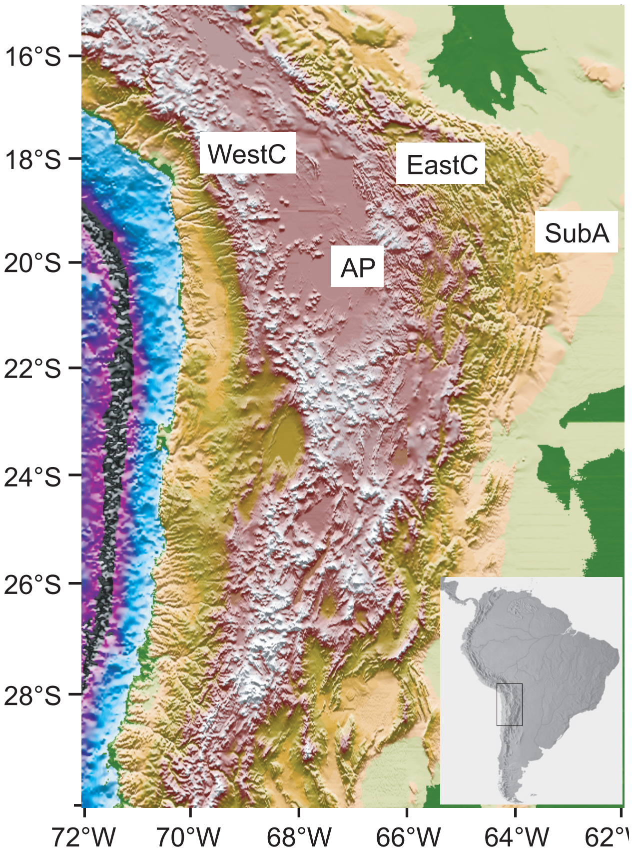

Fig. 4.1: Topographic map of the Central Andean plateau. The following abbreviations are used: WestC – Western Cordillera, AP – Altiplano, EastC – Eastern Cordillera, SubA – Subandean fold-and-thrust belt.

The orogen exhibits a symmetric oroclinal bend (Gephart, 1994) with its symmetry axis coinciding with the current direction of plate convergence. Both paleomagnetic data within the forearc (e.g., Macedo-Sánchez et al., 1992; Butler et al., 1995; Somoza et al., 1999; Roperch et al., 1999; 2000; Rousse et al., 2005) as well as differential shortening indicate counterclockwise rotation (up to 45°) north of the symmetry axis and clockwise rotation south of it explaining the oroclinal bending of the Andes (Isacks, 1988). This is supported by balanced cross sections (Kley, 1999). The convergence rate for relative motion between the Nazca plate and South America can be averaged to 7 cm/a for the last 30-40 Ma (Silver et al., 1998). At the largest distance (from forearc to foreland) of the orogen between 18-20°S, 280 km of shortening have been accumulated within the last 40 Ma (e.g., Kley, 1999).

The studied region within the Central Andes comprises the five major units from west to east: a) the fore-arc, b) the Western Cordillera (magmatic arc), c) the Altiplano plateau, d) the Eastern Cordillera, and e) the Subandean fold-and-thrust belt.

Both Cordilleras formed at positions of inherited structures: a Permoliassic rift in the Eastern Cordillera (Sempere et al., 2002), and the magmatic arc in the Western Cordillera (Oncken et al., 2006), both operating as weakness zones where strain localized. A low-velocity zone at a depth of 20 km is imaged in seismic sections. It can be interpreted as a zone of partial melt possibly decoupling deformation of the upper crust from that of the lower crust (Yuan et al., 2000), which has probably developed since about 30 Ma ago (Babeyko et al., 2002). This horizon possibly extents to the east connecting with the detachment of the Subandean fold-and-thrust which is located in Ordovician shales (Kley et al., 1996); the Subandean might be underthrust by the cold Brazilian Shield (Babeyko and Sobolev, 2005). To the west of the Western Cordillera, the little internally deformed forearc might have acted as a "pseudo-indenter" pushing into the comparably weaker orogen during subduction (Victor et al., 2004; Tassara, 2005). Thus, the orogen would have been squeezed between two relatively stronger units: the forearc and the Brazilian Shield. I used the mentioned characteristics (cf. Fig. 4.2) as analogies to develop the experimental set-ups, which are described in detail in Chapter 6 (granular series) and Chapter 7 (viscous-brittle series).

Shortening estimates based on balanced cross sections (for all relevant latitudes) have been

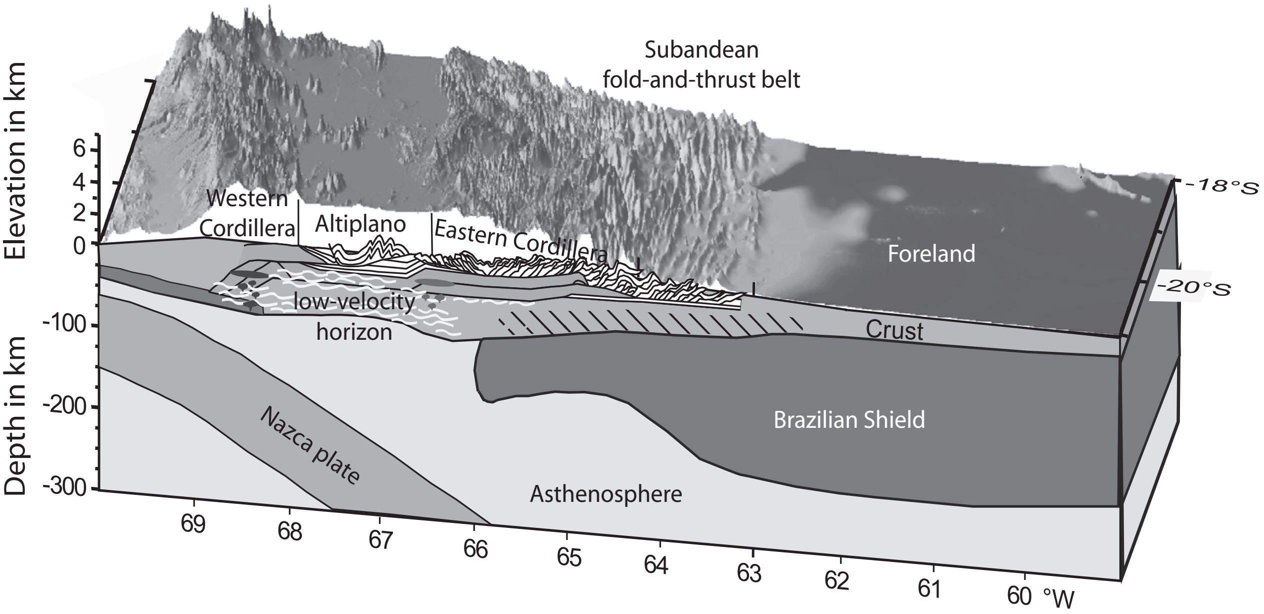

Fig. 4.2: Schematic cross section through the Central Andean plateau with the characteristics used as analogies for the experimental set-up. Further description in-text. Modified from Beck and Zandt (2002).

summarized in Oncken et al. (2006).

Deformation activity has been grouped into five main time windows by Oncken et al. (2006). The beginning of deformation in both Cordilleras (West: 46 Ma, East: 40 Ma) falls within the first window (46-37 Ma), the intramontane basin of the plateau area between these Cordilleras (Elger et al., 2005) has started to deform in the second time window (36-30 Ma) with the Cordilleras remaining active. Subsequent windows (29-20 Ma and 19-8 Ma) are marked by the north-south extent of deformation activity and the last (7-0 Ma) is characterized by transfer of the active deformation away from the plateau into the Subandean fold-and-thrust belt (e.g., Allmendinger and Gubbels, 1996). Responsible for this spatiotemporal strain distribution was a special combination of parameters including differential trench-upper plate velocity evolution, high plate interface coupling from low trench infill, and the lateral distribution of weak zones in the upper plate (Oncken et al., 2006).

Several separate mechanisms have been previously suggested to be responsible for the plateau formation and its uplift, but they cannot explain the strain evolution in space and time. The following mechanisms have been proposed: a) changes in plate convergence (e.g., Pardo-Casas and Molnar, 1987; Coney and Evenchick, 1994; Scheuber et al., 1994; Silver et al., 1998; Somoza, 1998); b) changes in the properties of the downgoing slab (e.g., Gephart, 1994; Giese et al., 1999; Gutscher er al., 2000a,b; Yanez et al., 2001); c) mantle processes (e.g., Isacks, 1998; Wdowinski and Bock, 1994a; Allmendinger et al., 1997; Kay et al., 1999; ANCORP-Working Group, 2003; Garzione et al., 2006); d) intra-plate strength variations (e.g., Allmendinger and Gubbels, 1996; Hindle et al., 2002; Lamb and Davis, 2003); and e) climate-related variations (e.g., Masek et al., 1994; Horton, 1999; Montgomery et al., 2001; Lamb and Davis, 2003; Sobel et al., 2003).

The two plateau areas, Altiplano and Puna, differ in many aspects. The Altiplano accumulates more shortening, deforms thin-skinned, is underthrust by the cold Brazilian Shield (Babeyko and Sobolev, 2005), and thus has a thicker crust. In contrast to this, the Puna accumulates less shortening, and has a less wide plateau area, which attains higher elevations than the Altiplano. It deforms thick-skinned and has a thinner lithosphere; this is probably due to mantle delamination (Kay and Kay, 1993; Sobolev and Babeyko, 2005), as the Brazilian Shield does not underthrust the Santa Barbara fold-and-thrust belt (which is the southern equivalent to the Subandean fold-and-thrust, that belongs to the Altiplano region).

Allmendinger and Gubbels (1996) have classified two different deformation modes for the two areas: the Altiplano deforms in a simple shear mode, meaning that localization of deformation in the upper crust does not occur in the same vertical column as the lower crust. This “pure shear” mode is true for the Puna plateau. In addition, the Puna plateau is still actively deforming and uplifting, while the Altiplano has already attained its final elevation.

Most of these characteristics can readily be explained by along-strike variations in the initial system: the location and type of sedimentary deposits, which determine the style of thrusting in the fold-and-thrust belt, the boundaries of a continental shield, or the fact that delamination of the lithospheric mantle affects the thermal state of the system and therefore its strength (e.g., Allmendinger and Gubbels, 1996). Thus, some characteristics in our models are rather true only for the Altiplano, like the fact that two stronger blocks compress a weaker orogen, analogue to the indenting forearc to the west (Victor et al., 2004; Tassara, 2005) and the strong Brazilian Shield to the east. The natural counterpart of the decoupling horizon is the layer of partial melt beneath the Altiplano (Yuan et al., 2000).

The Andean plateau is the natural example for my study, as the resolution of strain data is very high. This is due to unique outcrop exposures as a result of the arid climate, and also to the preservation of datable syntectonic deposits in intramontane basins. Therefore, the area is extensively studied and deformation is dated very well and highly resolved. All available data compiled from literature yield a comprehensive deformation database (Appendix A, previously published in Oncken et al., 2006). These data form the base for further analysis (Chapter 5).