CHAPTER 4
PALEOTECTONICS

The Southern Andean intra-arc zone has been affected by Paleozoic, Mesozoic, and Cenozoic tectonics including increments of extension and shortening (e.g. Malumián and Ramos, 1984). The latter probably led to mountain building during the Cretaceous and Late Miocene. In this work, the term "paleotectonics" refers to tectonics which are older than Pliocene. "Neotectonics" in this work means the "newest", i.e. Pliocene to active, tectonics.

4.1   Regional Geology

The geology of the study area (Fig. 4.1) basically reflects the climatic gradient: The fore-arc as well as the back-arc (not shown on the map) are composed of pre-Andean basement whereas the intra-arc zone, actually situated on the windward eastern side of the Main Cordillera shows southward increasingly exhumed rocks of the North Patagonian Batholith (NPB). This gradient in exhumation affected rocks as young as Late Miocene (Seifert et al., in press) providing the opportunity to study relatively young geologic processes along a profile through the upper 15 - 20 km of continental crust.

In the northern part of the study area (north of ca. 39.5°S), the dominant rock types are volcano-sedimentary successions of Jurassic and Tertiary age which form the basement of the Pliocene to recent volcanic arc. The volcano-sedimentary pile consists of two main successions: a Jurassic succession of volcanic, volcanoclastic, and sedimentary rocks deposited in a marine to continental environment in an intra-arc basin (Fm. Nacientes de Bíobío, De la Cruz and Suárez, 1997) which is the easternmost part of the Neuquén basin and an Eocene to Miocene succession of volcanic, volcanoclastic, and sedimentary rocks deposited in a continental, lacustrine to fluvial intra-arc environment, the Cura-Mallín basin (Fm. Cura-Mallín, Fm. Trapa-Trapa, and Fm. Mitrauquén, Suárez and Emparan, 1997, Niemeyer and Munoz, 1983, Kemnitz et al., in press).

Structures in the northern part of the study comprise Cretaceous and Miocene folds and inverted normal faults and Pliocene to recent strike-slip faults. The volcano-sedimentary sequences are intruded by post-tectonic granitoid plutons of Cretaceous and Miocene age. Pressures during the crystallization of these plutons have been inferred to be less than 1 kbar (Seifert et al., in press, Fig. 4.1 b) constraining the crustal depth of emplacement to be less than ca. 3 km.

In the southern part of the study area (ca. 39 - 42°S), the dominant rocktypes are granitoids of Carboniferous to Miocene age assigned to the North Patagonian Batholith forming the basement of the Pliocene to recent volcanic arc. These granitoids emplaced at depths of ca. 5 – 15 km as constrained by estimations of their crystallization pressures (Seifert et al., in press, Fig. 4.1 b). Structures in the southern part of the study area comprise Cretaceous and Miocene ductile shear zones and Pliocene to recent strike-slip faults. Volcano-sedimentary strata of Triassic to Tertiary age unconformly overlying the granitoid basement occur along the western foothills of the Main Cordillera and, as erosional remnants, in higher parts of the Main Cordillera.
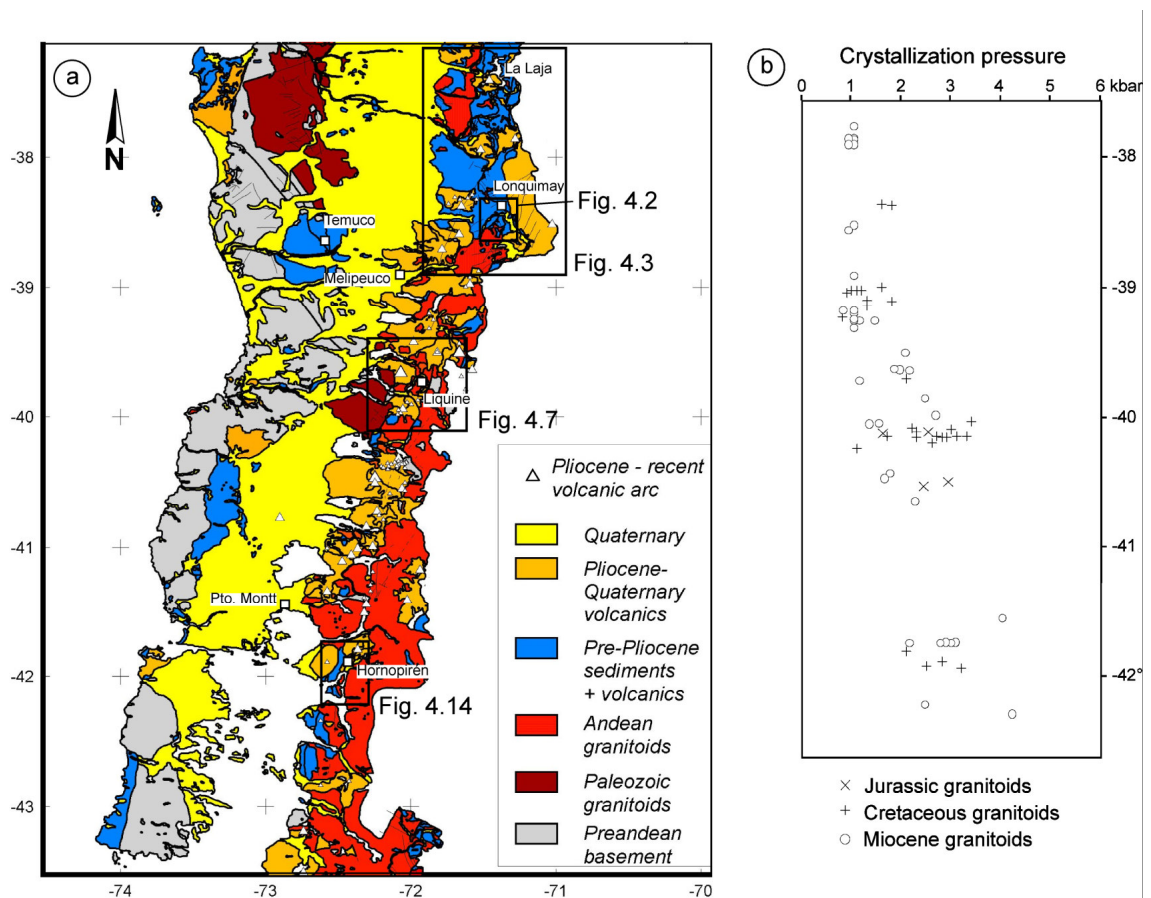

Fig. 4.1: *Geological overview of the study area: a) Simplified geologic map, boxes indicate locations of detailed maps; b) Crystallization pressure of granitoid basement along strike of the volcanic arc (Al-in-hbl data from W. Seifert, GFZ, see Appendix V, Tab. A4).*

In three subareas of the study area, near Lonquimay (38°S), Liquiñe (40°S), and Hornopirén (42°S), detailed studies revealed marked differences in paleotectonic (i.e. pre-Pliocene) tectonic style along strike of the intra-arc zone. In the Lonquimay area, Cretaceous and Late Miocene compressional phases led to basin inversion. In the Liquiñe and Hornopirén areas, in contrast, Cretaceous and Late Miocene transpressional tectonics led to the uplift and exhumation of basement blocks.

4.2   Paleotectonic structures in the Lonquimay area (38°S)

4.2.1   Inversion of the Neuquén basin

Rocks of the easternmost part of the Neuquén basin, the Fm. Nacientes del Biobío (De la Cruz and Suárez, 1997), crop out in the mountain range south of Lonquimay (Cordillera Lonquimay, Fig. 4.2). At the eastern border of the range, the Jurassic succession is in tectonic contact with the Tertiary rocks of the Fm. Cura-Mallín. To the north and to the west, the Jurassic succession represents structurally as well as morphologically the shoulder of the NNE-SSW to NE-SW striking Lonquimay graben system. To the south, outcrop of Jurassic strata is reported until the E-W striking Melipeuco valley.

Open cylindrical folds with hundreds of meters wavelength are developed within this strata. Corresponding fold axes are subhorizontal and strike NNE-SSW (Fig. 4.2). The folds are intruded by Cretaceous granitoid plutons (Grupo Plutonico Guallétue, Suárez and Emparan, 1997) whose intrusive contacts clearly crosscut the folds. These post-tectonic granitoids yield K-Ar ages between 73 and 148 Ma with a dominance of ages between 80 and 85 Ma (Suárez and Emparan, 1997). Fossils found in the Fm. Nacientes del Biobío indicate an upper Pliensbachian to Kimmeridgian age (De la Cruz and Suárez, 1997). Consequently, folding is constrained to have occurred in the Early Cretaceous.
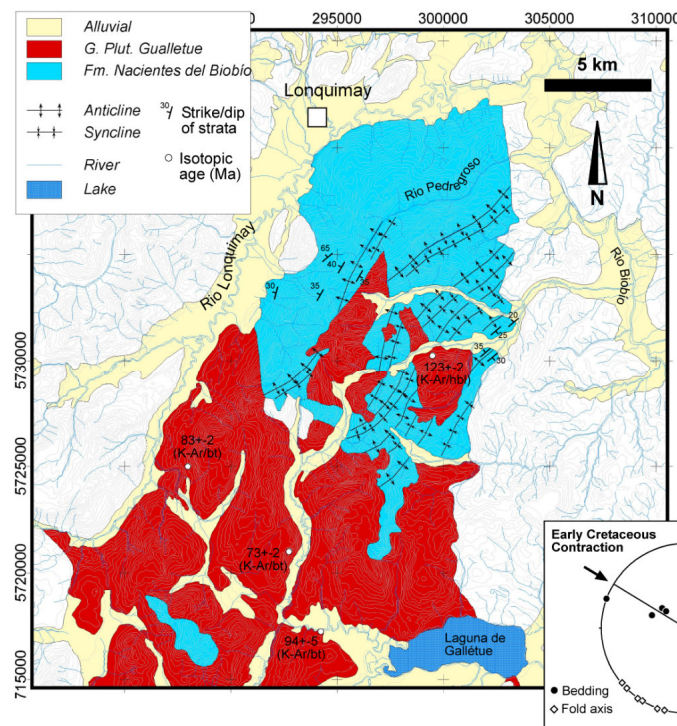

Fig. 4.2: *Map of Early Cretaceous structures in the Cordillera Lonquimay (modified from Suárez and Emparan (1997), UTM 19 S projection, PSAD 1956 ellipsoid, inset plot: lower hemisphere, equal area).*

4.2.2 Inversion of the Cura-Mallín basin

Rocks of the Eocene to Miocene Cura-Mallín basin (Fm. Cura-Mallín, Fm. Trapa-Trapa, and Fm. Mitrauquén, De la Cruz and Suárez, 1997) have their main outcrops in the intra-arc zone between ca. 37 and 38°S (Fig. 4.3).

In the northern part of the Lonquimay area, the strata is folded with hundreds of meters to kilometers wavelength along subhorizontal, N-S striking fold axes. Folds are generally open to tight and cylindrical (Fig. 4.4). The bedding is tilted up to vertical position or is even overturned. Some of the tighter folds show a westward vergence.
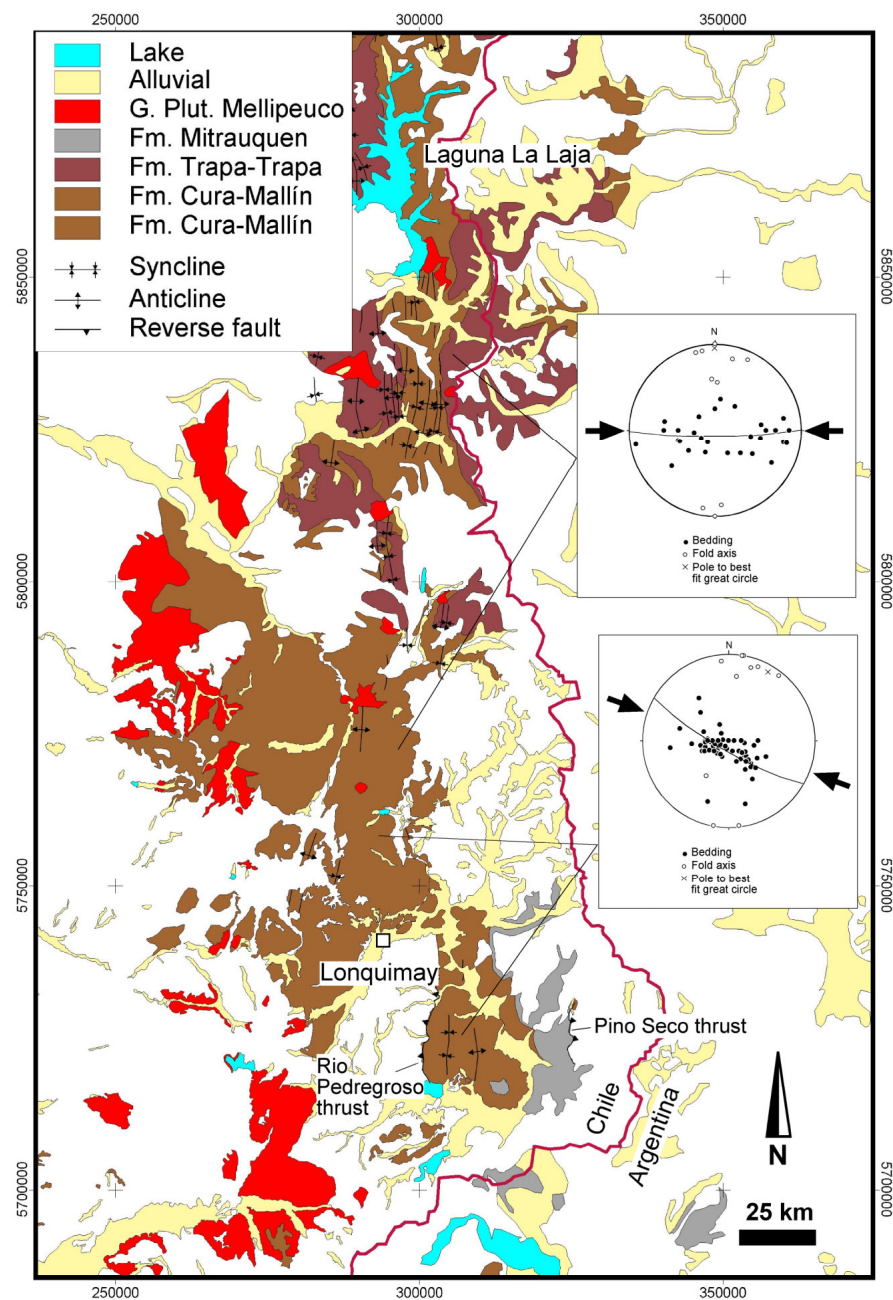

Fig. 4.3: *Map of Late Miocene structures in the Lonquimay area (modified from Suárez and Emparan (1997), UTM 19 S projection, PSAD 1956 ellipsoid, inset plot: lower hemisphere, equal area).*
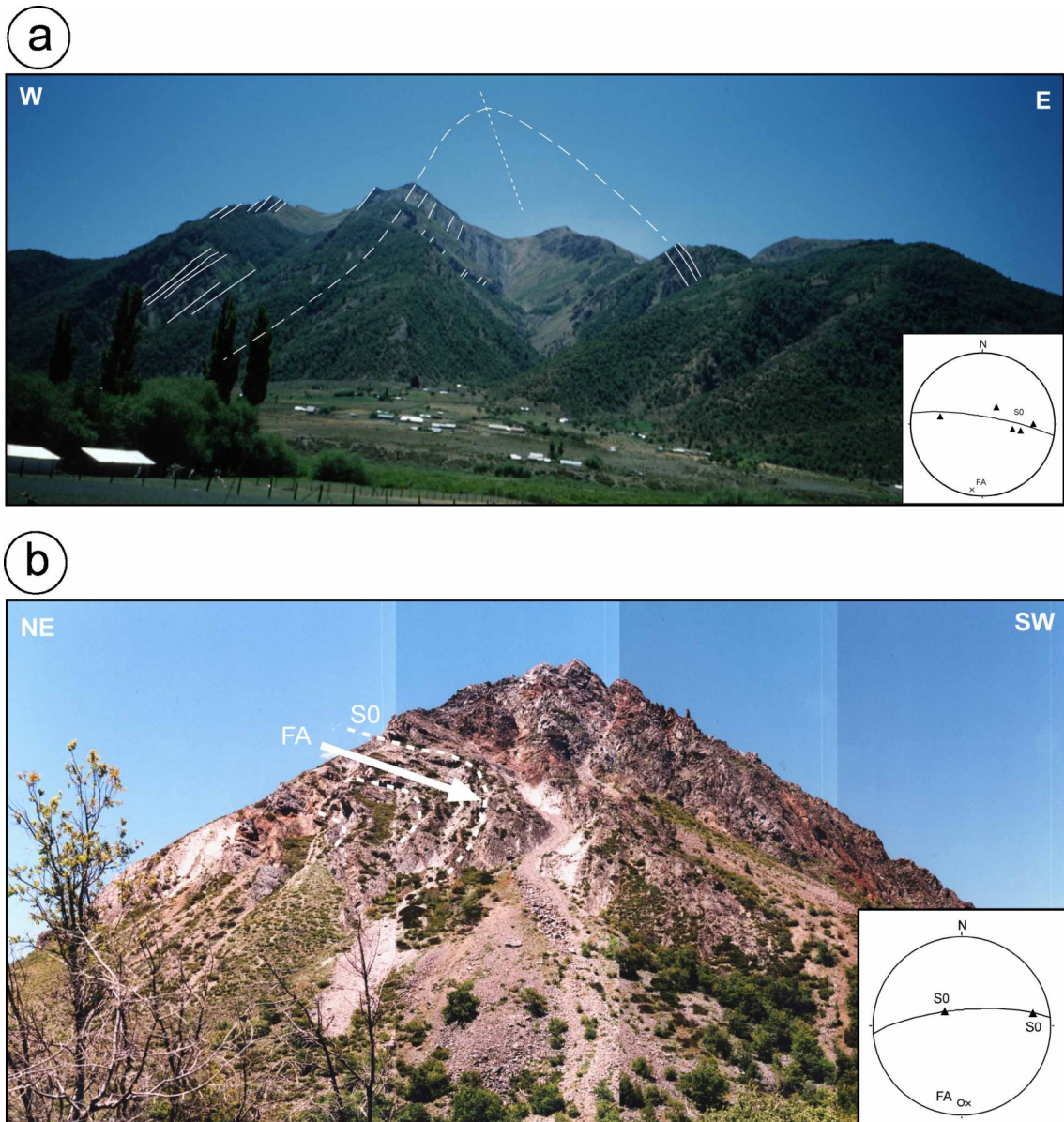
Fig. 4.4: *Late Miocene folds in the northern part of the Lonquimay area: a) Queuco river valley, b) south of Callaqui volcano.*

In the southern part of the Lonquimay area, in contrast, the strata shows only very gentle folding. Folds here are very open with limbs inclined generally less than 20-30° and with corresponding subhorizontal, N-S striking fold axes (Figs. 4.5). Two N-S striking thrusts are observed: the Pino Seco thrust delimiting Miocene strata to the east (Fig. 4.6a, b) and the Pedregroso thrust delimiting Miocene strata to the west of the Biobío valley (Fig. 4.6c). The Pino Seco thrust is inclined 20°-30° to the east and juxtaposes Jurassic strata upon ignimbrites and conglomerates of the Late Miocene Mitrauquén formation. The uppermost ignimbrites of this formation are dated as 8 Ma (Suárez and Emparan, 1997). The thrust is sealed by Early Pliocene Lavas of the Association Volcanico Precordillera Oriental (Fig. 4.6b), the oldest of which yielded an age of 5 Ma (Niemeyer and Munoz, 1983, Suárez and Emparan, 1997). Therefore,

thrusting along the Pino Seco thrust must have occurred between 5 and 8 Ma. The Pedregroso thrust has not been observed directly but is inferred from the observations that (1) Jurassic rocks are exposed at higher topographic levels west of the presumed fault trace and (2) that Tertiary strata east of it is inclined 10-20° to the east and can be interpreted as drag folding associated with eastward thrusting.
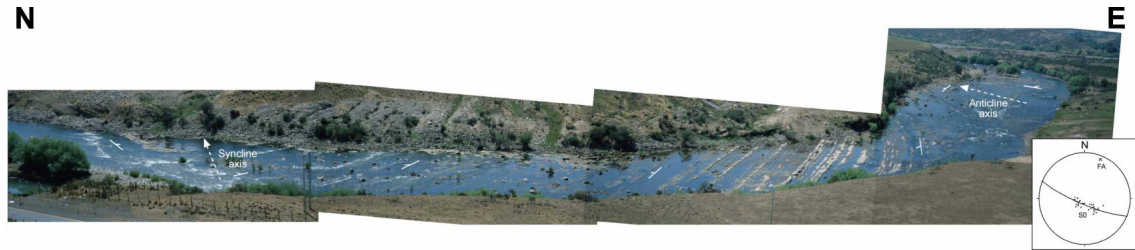

Fig. 4.5: *Late Miocene folds in the southern part of the Lonquimay area (view to the NE).*
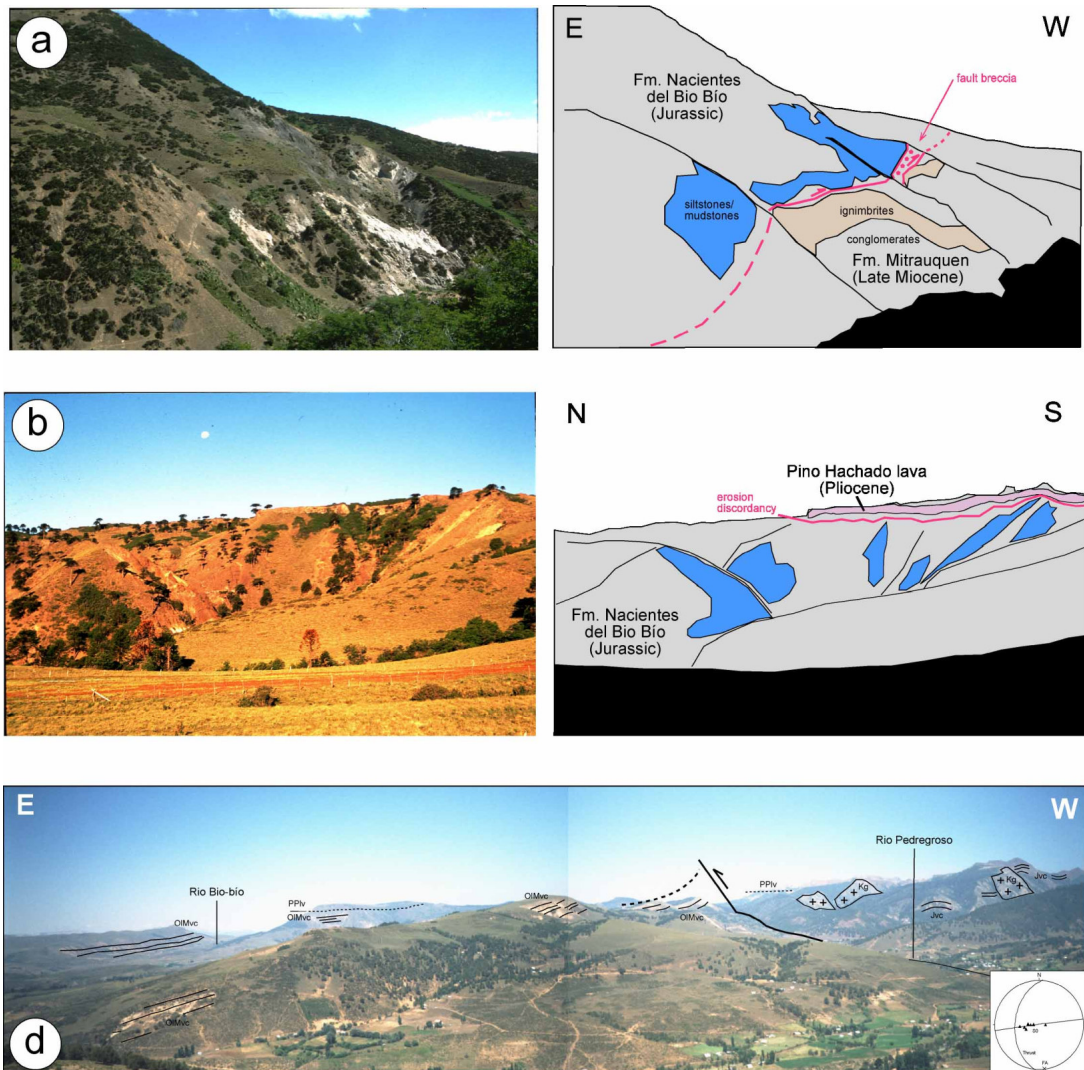

Fig. 4.6: *Late Miocene thrusts in the Lonquimay area: a) Pino Seco thrust, b) Early Pliocene lava unconformly overlies overthrusted and eroded Jurassic rocks, d) Pedregroso thrust.*

Early Pliocene lavas (Niemeyer and Munoz, 1983, Suárez and Emparan, 1997) unconformly overlie folded and eroded Tertiary strata in the Lonquimay area indicating that regional contraction ceased before the beginning of the Pliocene. However, the widespread occurrence of locally posttectonic granitoid plutons which yielded ages between 15 and 7 Ma (Suárez and Emparan, 1997) may indicate that contraction occurred not only during the 5 - 8 Ma interval when thrusting along the Pino Seco thrust took place, but also earlier during the Late Miocene. This is consistent with the observation of a local unconformity between the Cura-Mallín and the Trapa-Trapa formation (ca. 15-18 Ma, Suárez and Emparan, 1997) and the occurrence of out-of-sequence thrusting northeast of the study area (Folguera et al., 2002). It may be concluded, that intra-arc contraction started already during the Middle Miocene. Consequently, deposition of the youngest members of the Fm. Trapa-Trapa, and the Fm. Mitrauquén has to be considered as syntectonic. The Late Miocene Mitrauquén formation shows clear indications for syntectonic deposition. In the east, where the Mitrauquén formation is overthrusted by the Jurassic succession, it consists of an interlayering of ignimbrites and conglomerates of tens of meters thickness including up to decimeter large clasts. To the west, this sequence thins out rapidly indicating a wedge-shaped geometry with the main depocenter close to the thrust and the supply area to the east.

4.3   The Lonquimay area as an example of shallow crustal tectonics

4.3.1 Mechanisms and kinematics

The Lonquimay area shows paleotectonics representative for shallow crustal levels (< ca. 3 km) during the Early Cretaceous and Late Miocene increments of deformation. The Late Miocene structural inventory of the Lonquimay area includes N-S trending folds and inverse faults both associated with erosional unconformities. Syntectonic deposits associated with the growth of folds (growth deposits) have not been observed because of the lack of adequate outcrop and/or preservation. Folds in the Lonquimay area are generally open to tight and cylindrical. The presence of slickensides on the bedding planes oriented perpendicular to the fold axes indicate that flexural-slip was the dominant folding mechanism, i.e. that parallel folds were produced by slip of layers over each other. There is no regional tendency in the direction of vergence of both folds and faults.

Although in the Lonquimay area is no opportunity to observe the relationship between folds and faults directly, it is plausible that folds in the study area are related to blind inverse faults. This is consistent with findings in adjacent areas which are based on field (structural and sedimentological) and seismic evidences:

- To the north of the Lonquimay area, laterally extensive thrust faults are clearly related to folds within Cenozoic deposits (e.g. Charrier et al., 2002, Radic et al., 2002, Godoy et al., 1999)

- To the west of the study area, in the Central Valley around Lago Llanquihue (40° - 41°S, 73°W), seismic data suggest that moderate-angle reverse faults are associated with folds within Cenozoic deposits (Jordan et al., 2001, and references therein)

- To the northeast of the Lonquimay area, in the argentine Andocollo area (ca. 37°S, 71°W), folds within Cenozoic deposits are probably related to the inverse eastern Andean boundary fault (Jordan et al., 2001)

- To the southeast of the study area, in south Central-West Argentina (40° – 42°S, 71°W), folds are associated with basement ramps (Spaletti and Dalla Salda, 1996, Diraison et al., 1998)

Most authors agree, that these inverse faults are inverted Oligo-Miocene basin-bounding normal faults (Vergani et al., 1995, Jordan et al., 2001, and references therein). No detachment faults has been proposed so far for the Southern Andes, however, gently folded strata observed in the area is likely to reflect shortening above such a detachment horizon. Because folds generally form parallel to the finite strain ellipsoid (i.e., fold axis is parallel to finite extension, Tikoff and Peterson, 1998), the regional consistent trend of fold axis subparallel to the volcanic arc (which is presumably parallel to the shear plane) indicate either large amounts of passive rotation (> 30 – 40°) of initially oblique fold axis into parallelism with the shear plane or, more likely, the dominance of the pure shear.

4.3.2 Amount of shortening

To the author's knowledge, there has been no attempt up to now to quantify the amount of cross-arc shortening in the study area. This is probably due to the patchy outcrop conditions and the lack of seismic lines which could be used to construct a reliable

balanced cross-section of this part of the Southern American active margin. However, some constraints of the amount of shortening accommodated during the Late Miocene increment of deformation may be given using a simplified two-dimensional kinematic model (Fig. 4.7) which has the following conditions and parameters:

- Shortening accommodated by detachment faults is mirrored by gently dipping strata (10 - 20°). Thus, shortening accommodated by folding above a detachment here is generally about 1.5 - 6 % (Fig. 4.7 b).

- Uplift of pre-Tertiary basement occurred along inverted normal faults. The vertical throw of such faults is constraint to the maximum by (1) the thickness of the basin infill (before inversion) and (2) the depth of exhumation in the hanging block (after inversion). The maximum basin infill in the study area is around 3 km (Jordan et al., 2001). The maximum depth of exhumation of pre-Tertiary granitoid basement in the Lonquimay area is around 3 km (Seifert et al., in press). Thus the vertical throw is ca. 3 - 6 km. Assuming a dip of more than 60° for the inverted normal-faults and a simple planar fault geometry (i.e. no listric fault plane), a vertical throw of 3 - 6 km corresponds to a theoretical horizontal displacement of 1.7 - 3.5 km (Fig. 4.7 c). In the Lonquimay area, two of those faults exist (the Rio Pedregroso and Pino Seco faults) along a profile of ca. 100 km width. Thus the amount of shortening accommodated by faulting in the Lonquimay area is probably about 3.5 - 7 %.

- Outcrop-scale strain and ductile microstrain can be neglected for the purpose of a simplified quantification. For sure, they cannot in a balanced cross section, because they may represent 1/6 to 1/4 of total shortening (Hogan and Dunne, 2001).

According to this kinematic model, the relative amount of cross-arc shortening in the Lonquimay area is in the order of 5 - 13 %. If the Lonquimay area is taken as representative and considering that shortening affected the whole continental margin, from the fore-arc to the back-arc region, in total a zone 200 – 300 km wide, then up to 10 – 39 km of cross-arc shortening may have occurred during the Late Miocene in the Southern Andes.
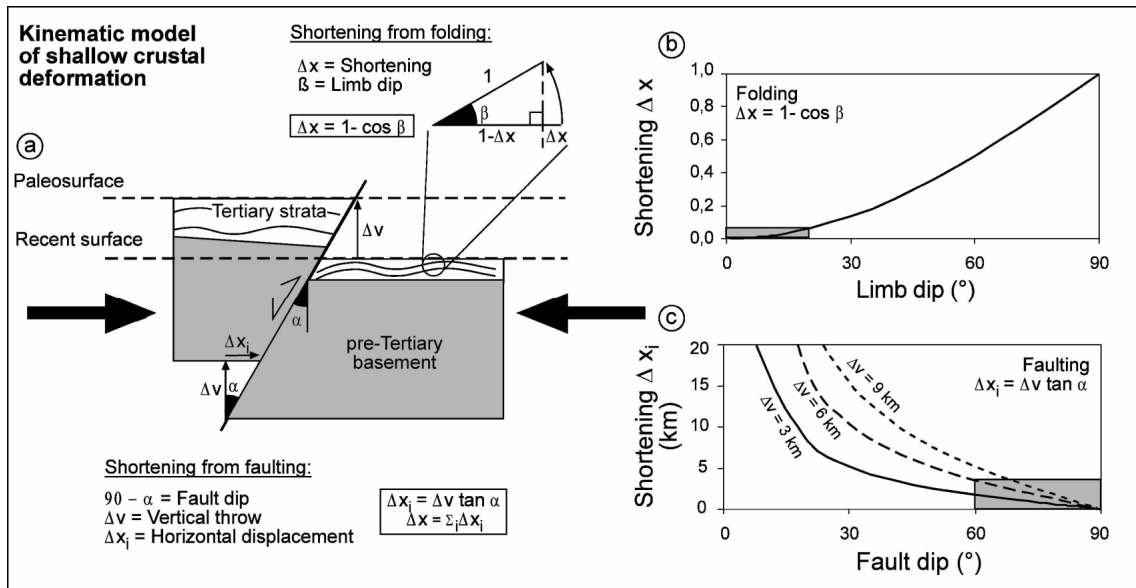
Fig. 4.7: *2-D kinematic model of Late Miocene deformation in the Lonquimay area: (a) Concept of the model including inverted normal faults and folds; (b) Relative amounts of shortening from folding; (c) Absolute amount of shortening from a single fault.*

Although, this is a highly speculative shortening balance, it demonstrates that upper crustal cross-margin shortening in the Southern Andes is about one order of magnitude less than in the Central Andes. In the Arica region (19°S), Neogene cross-continental margin shortening amounts are highest and range between 210 and 530 km (Sheffels, 1990, Schmitz, 1994, McQuarrie, 2002a, b) and decreases systematically to about 20 km at 27°S (Kley and Monaldi, 1998). Thus the here proposed model is conform with the widely accepted model of formation of the Andean plateau and Bolivian orocline due to differential back-arc shortening (Kley and Monaldo, 1998, Kley, 1999, Hindle et al, 2002, Klosko et al., 2002). Most authors agree that shortening in the Central Andes accelerated in the Late Miocene during the main plateau building deformational increment known as the "Quechua orogenic phase" (Malumián and Ramos, 1984) largely coeval to basin inversion in the Southern Andes.

4.4   Paleotectonic structures in the Liquiñe area (40°S)

In contrast to the Lonquimay area, pre-Pliocene structures in the Liquiñe area are subvertical inverse faults cutting dominantly granitoid basement and leading to a block-like tectonic segmentation. Plutonic rocks in the Liquiñe area are dominantly hornblende-biotite granodiorites and granites of Carboniferous to Miocene age which intruded at mid-crustal levels (5 - 20 km, Seifert et al., in press). In places, they are unconformly overlain by Triassic to Tertiary volcano-sedimentary rocks. Three NNE-SSW trending subvertical faults, from west to east the Maihue – Neltume fault (MNF), Pilmaiquen – Liquiñe fault (PLF), and Reigolil – Pirihueico fault (RPF), separate four basement blocks with different exhumation histories (Fig. 4.8):

(1) The fore-arc, bound to the east by the Maihue and Neltume faults, which had been exhumed to the surface prior to the Late Triassic

(2) The "Neltume block", named after Lago Neltume situated in this block, between the Maihue and Neltume faults to the west and the Pilmaiquen and Liquiñe faults to the east, which had been exhumed during the Late Miocene

(3) The "Pirihueico block", named after Lago Pirihueico situated in this block, between the Pilmaiquen and Liquiñe faults to the west and the Reigolil-Pirihueico fault to the east, which had been exhumed during the Late Cretaceous/Early Tertiary

(4) The back-arc which had been exhumed probably during the Early Triassic

The geology of the different blocks together with the discordancy relationships constraining their exhumation history is described in detail below.

4.4.1 Fore-arc

West of the Maihue fault (Campos et al., 1998) and its assumed northern prolongation, the Neltume fault (referred to as the Maihue – Neltume fault, MNF), pre-Andean basement consisting of dominantly Permocarboniferous granitoids (Panguipulli batholith, Martin et al., 1999) crop out. Late Paleozoic metasedimentary rocks in this area (Fm. Trafún) are correlated with rocks of the Eastern Series (Martin et al., 1999). They were intruded by granitoids which were emplaced at depths of 15-20 km (Seifert et al., in press) during the Permocarboniferous (Martin et al., 1999). This basement

shows a discordant erosional contact with Late Triassic sedimentary rocks of the Fm. Panguipulli (Martin et al., 1999) indicating that they had been exhumed to the surface prior to Late Triassic times. The Fm. Panguipulli records subgreenschist facies metamorphism (Martin et al., 1999) suggesting that the fore-arc has not experienced significant vertical movements since Early Mesozoic times. This is supported by shallow level, Miocene granitoids which intruded Tertiary volcanosedimentary deposits (Fm. Ranco, Campos et al., 1998) overlying pre-Andean basement in the southern part of the Liquiñe area.
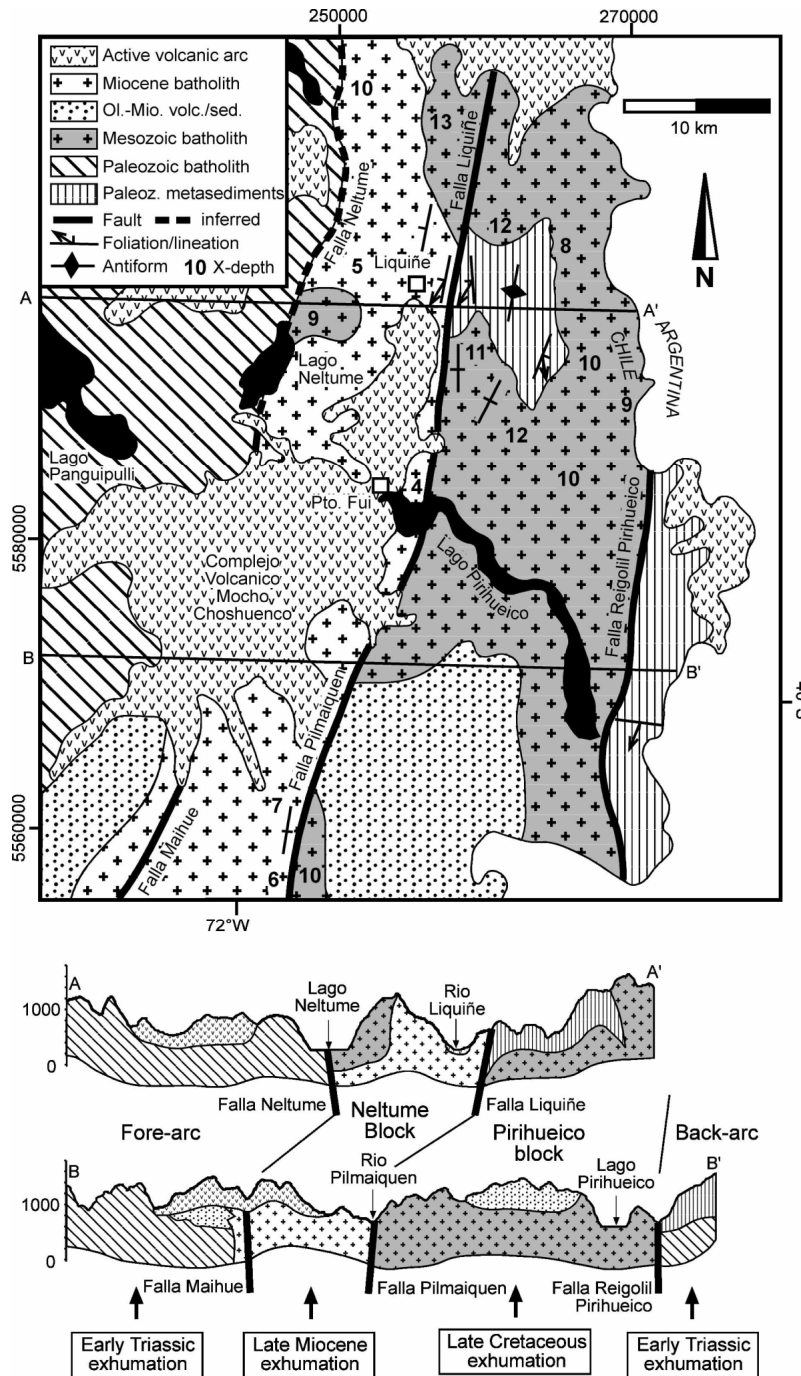

Fig. 4.8: *Geological map and profiles of the Liquiñe area (modified from Lara and Moreno (1998), UTM 19 S projection, PSAD 1956 ellipsoid). Crystallization depths (X-depths) are derived from Al-in-hbl geothermobarometry (Seifert et al., in press)*

4.4.2   Neltume Block

Between the MNF and the Pilmaiquen fault and its northern prolongation, the Liquiñe fault (referred to as the Pilmaiquen – Liquiñe fault, PLF), mid-crustal (4-10 km, Seifert et al., in press) granitoids of dominantly Miocene age crop out (Fig. 4.8). Published isotopic ages (K-Ar data of biotite and amphibole, Lara and Moreno, 1998, and references therein) for these granitoids range between 6 and 15 Ma. These ages may represent either intrusion ages or cooling ages. East and west of Neltume block, unmetamorphic Oligocene - Middle Miocene volcano-sedimentary strata of the Fm. Ranco (Campos et al., 1998) crops out. This strongly suggests that the MNF and PLF were active as inverse faults after the Middle Miocene causing uplift and exhumation of the Neltume block with respect to the adjacent blocks. Late Pliocene - Pleistocene (Lara et al., 2001) lavas of the active volcanic arc unconformably overlie Miocene granitoids and seal the MNF indicating that uplift and exhumation of the Neltume block occurred during the Late Miocene/Early Pliocene.

The only place where the PLF can be observed directly is along the road from Liquiñe to the Argentine border, ca. 2 km east of the town of Liquiñe. The road passes the Liquiñe fault which juxtaposes a Miocene granodiorite onto Cretaceous migmatites. A ca. 10 m wide zone of lower greenschist facies mylonites is developed from the generally undeformed Miocene granitoid. These mylonites show a well developed foliation dipping steeply to the west. Kinematic indicators (SC-, CC´-structures, etc.) suggest a dextral sense of shear along a weakly developed, subhorizontal lineation. In thin section (Fig. 4.9), one can observe features indicative of crystal plasticity in quartz (undulose extinction, deformation bands, subgrain formation). Chlorite recrystallizes as fine grained crystals along the mylonitic folation and along shear bands. Dynamic recrystallization of quartz is consistent with syntectonic lower greenschist facies metamorphic conditions.
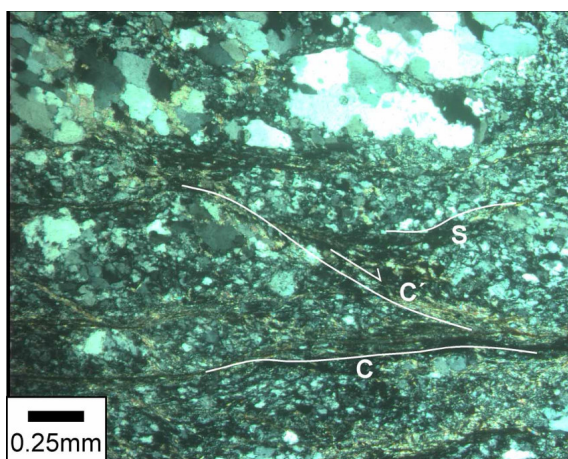

Fig. 4.9: Micrograph of greenschist facies mylonites in the Miocene granitoid near Liquiñe (section normal to foliation, parallel to lineation, crossed nicols).

4.4.3 Pirihueico block

Between the PLF and the Pirihueico-Reigolil fault (PRF, Lara and Moreno, 1998), dominantly Jurassic to Cretaceous granitoids, emplaced at mid-crustal depths, (8-12 km, Seifert et al., in press) crop out (Fig. 4.8). Published isotopic ages (K-Ar data of biotite and amphibole, Lara and Moreno, 1998) range between 184 and 74 Ma. 7 of 12 age determinations give ages between 100 and 110 Ma suggesting that large portions of the Mesozoic batholith emplaced during the mid-Cretaceous. South of the Lago Pirihueico, unmetamorphic deposits of the Oligocene-Middle Miocene Fm. Ranco (Campos et al., 1998) unconformly overlie the granitoid basement indicating that the Pirihueico block had been exhumed during the Late Cretaceous - Early Tertiary.

East of the town of Liquiñe, Cretaceous migmatites (100 - 105 Ma, see Ch. 5.1 for detailed petrography, chemistry, and geochronology of these rocks) represent a domal-like roof pendant (further referred to as the Liquiñe migmatite dome, Fig. 4.10) of the mid-Cretaceous batholith. Migmatites are developed from metasedimentay rocks (as indicated by the presence of rounded zircons) and contain quartz, feldspar, biotite, and muscovite as the main mineral phases. The layering within the migmatites is defined by muscovite and biotite and has variable orientations (see upper right stereoplot in Fig. 4.10) in the central part of the dome. Occasionally, decimeter-scale folding has been observed. At the flanks of the dome, the migmatitic layering is overprinted by a late-syn- to postmigmatitic mylonitic foliation (80 - 105 Ma, see Ch. 5.1) which is defined mainly by muscovite. The mylonitic foliation dips steeply to the west on the western flank and to the southeast on the eastern flank, a weakly developed mineral stretching lineation defined by muscovite varies in orientation from subhorizontal to vertical (see lower left and lower right stereoplots, respectively).

Microscale observations indicate greenschist facies metamorphic conditions during deformation (Fig. 4.11). The main mineral phases which recrystallized during deformation are muscovite (along C and C´ planes) and quartz. Kinematic indicators indicate east-up on the western side and west-up on the eastern side of the dome. This is consistent with uplift of the central part of the dome most probably due to intrusion of the pluton underneath.
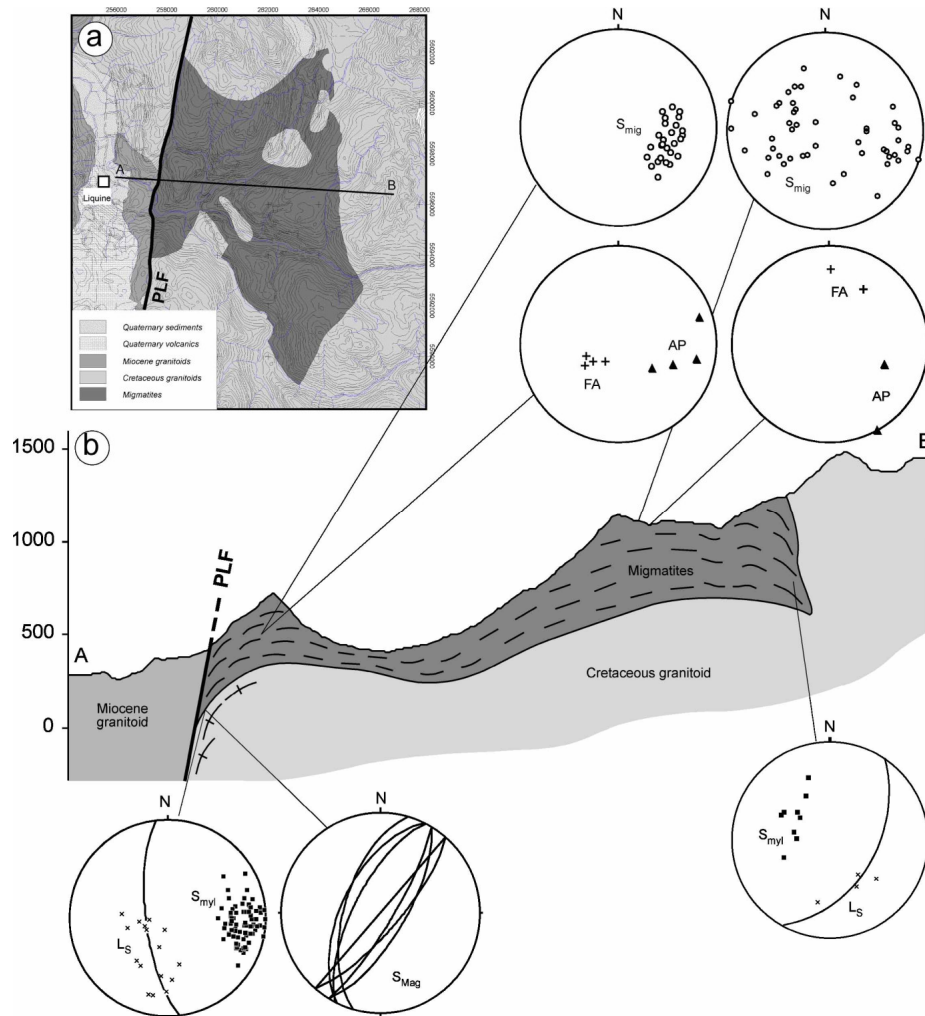
Fig. 4.10: *Geological map (a, UTM 19 S, PSAD 1956 ellipsoid) and profile (b) through the Liquiñe migmatite dome. Structures are shown as lower hemisphere, equal area stereoplots. PLF = Pilmaiquen-Liquiñe fault.*
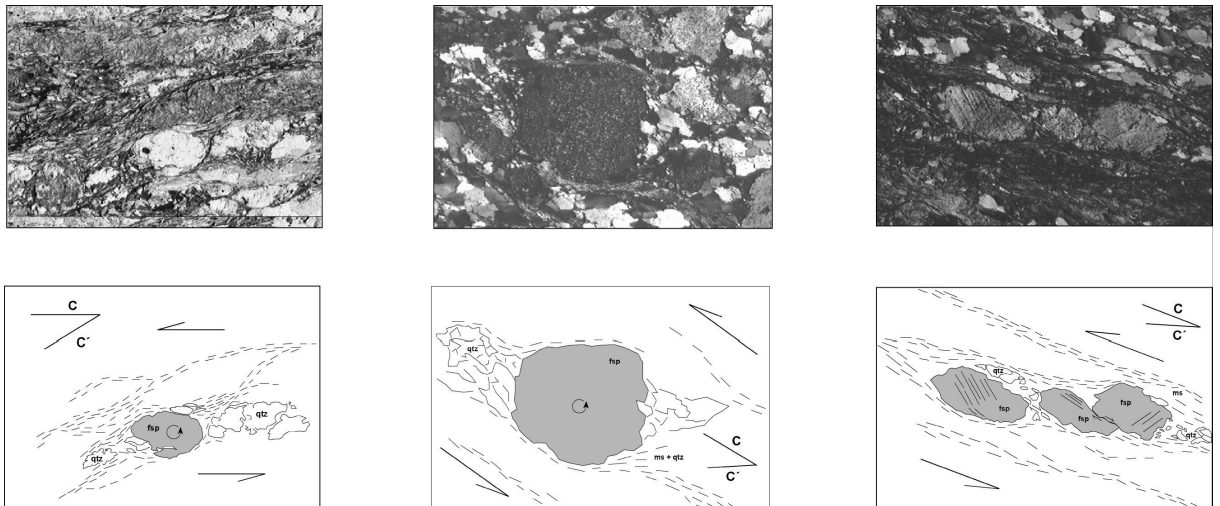

Fig. 4.11: *Micrographs and line drawings of greenschist facies mylonites on the eastern flank of the Liquiñe migmatite dome (sections normal to foliation, parallel to lineation). Microfabrics indicate sinistral sense of shear along a generally SW-dipping mineral stretching lineation, i.e. east-up. Note the contrasting rheological behaviour of quartz and mica (ductile deformation, dynamic recrystallization) with respect to big crystals of feldspar (brittle deformation, passive rotation): (**a**, transmitted light) s-clast within shear bands, (**b**, crossed polars) d-clast rotating synthetically within the shear plane, (**c**, crossed polars) antithetic bookshelf sliding and boudinage within shear bands.*

4.4.4 Back-arc

East of the PRF, Permian low-grade metapelites (phyllites) and Carboniferous granitoids (see Ch. 8.3 for detailed petrography, chemistry, and chronology of these rocks) crop out. The phyllites bear a south-dipping crenulation cleavage S_C and an associated lineation L_C (Fig. 4.12). I interpret these rocks to be related with rocks of the Eastern Series and to share a common PT-History (see Ch. 5.2). If so, they are the eastern pendant of rocks of the fore-arc described above and probably exhumed also during the Early Triassic.
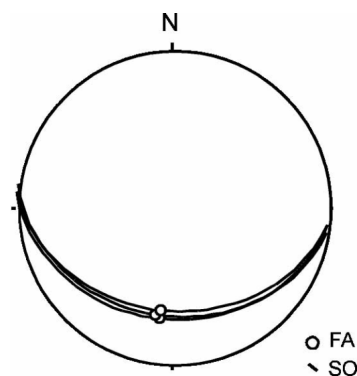

Fig. 4.12: *Pre-Andean structures of the back-arc basement (lower hemisphere, equal area stereoplot).*

4.5   Paleotectonic structures in the Hornopirén area (42°S)

In the Hornopirén area (Fig. 4.13), plutonic rocks (hornblende-biotite tonalite and granodiorite) mainly of Late Miocene ages (Cembrano et al., 2000) crop out. They were emplaced at depths of around 8 - 15 km (Seifert et al., in press). These granitoids are ductily deformed within a N-S striking, ca. 5 km wide zone bound by presumably tectonic contacts (Hornopirén fault E and W) with weak to non-metamorphic Tertiary volcanic rocks to the west and greenschist facies metavolcanics of presumably Tertiary age to the east. The Miocene granitoids are hosted by greenschist to amphibolite facies metapelites and metabasites belonging to the Paleozoic accretionary wedge which makes up the fore-arc region (Western Series, Pankhurst et al., 1992, Martin et al., 1999). Deformation and contact metamorphism within the metamorphic host rock are related to intrusion of the Late Miocene granitoids (Cembrano et al., 2000). Published isotopic ages (Cembrano et al., 2000) from deformed Miocene granitoids range between 3.6-6.6 Ma for biotite samples (Ar-Ar), and 8.7-13.1 Ma for hornblende samples (Ar-Ar). Deformation of the country rock within the deformation zone occurred contemporaneously with plutonism during the Late Miocene as indicated by Ar-Ar ages ranging between 7.2 Ma (biotite) and 10.6 Ma (muscovite).
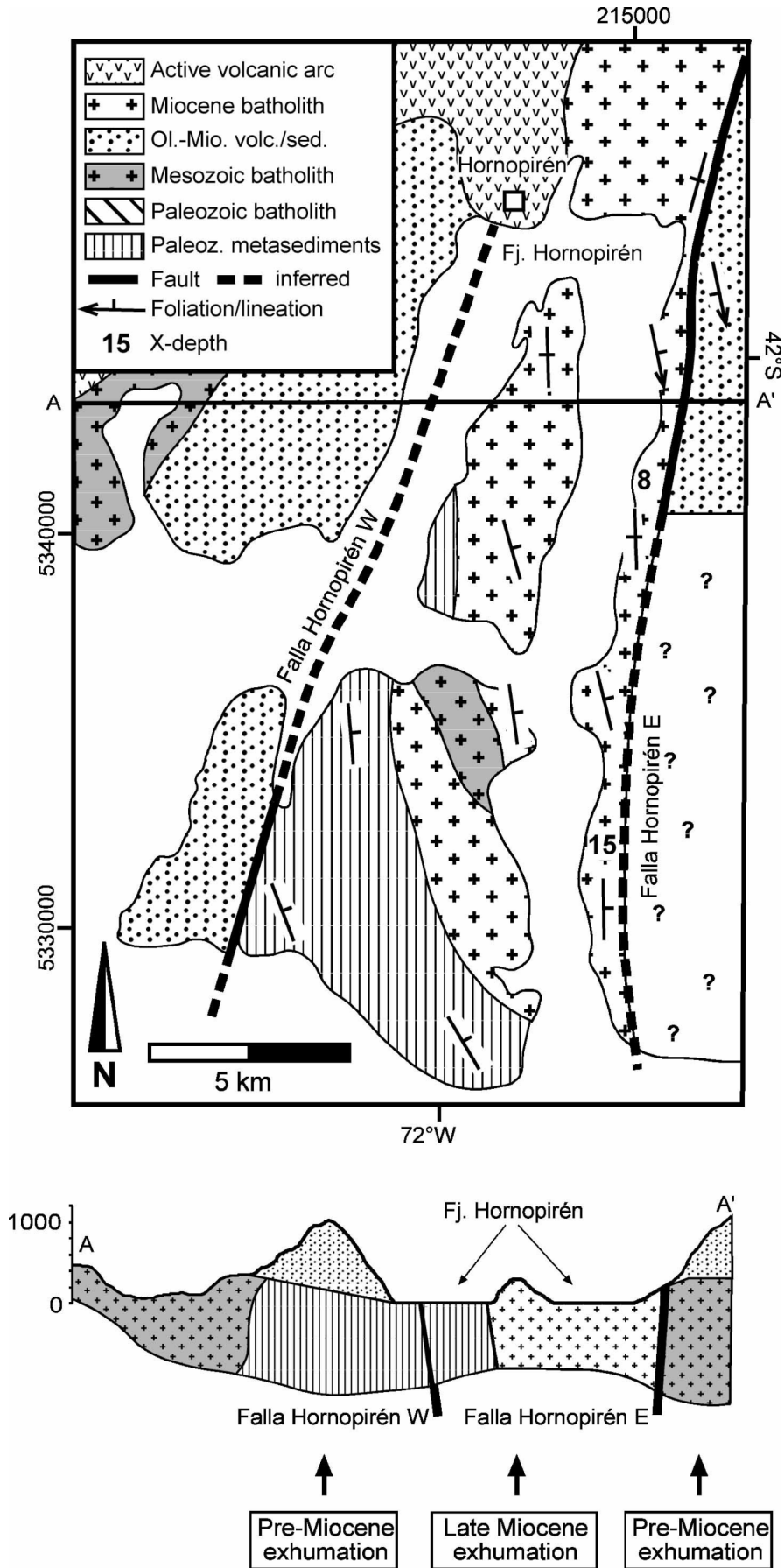
Fig. 4.13: *Geological map (modified from Cembrano et al. (2000), UTM 19 S projection, PSAD 1956 ellipsoid) and profile of the Hornopirén area.*

In the northeastern part of the Hornopirén fjord, the tectonic contact between the Miocene granitoid and greenschist facies metavolcanics (Hornopirén fault E) can be mapped. The fault strikes NNE-SSW, is subvertical, and follows the eastern flank of the Hornopirén fjord. Along the fault, chlorite-muscovite-bearing mylonites developed from a probably volcanic protolith, which may correspond to Miocene volcano-sedimentary deposits east of the Hornopirén fault W. The greenschist facies mylonites show a penetrative NNW-SSE striking, subvertical foliation bearing a subhorizontal mineral stretching lineation (S_{myl} and L_S, respectively, Fig. 4.14). At macroscale, they appear as fine grained green schists. In thin section, chlorite and muscovite can be recognized as the main mineral phase recrystallized during deformation beside quartz along the mylonitic foliation (Fig. 4.15). Quartz shows features of crystal plasticity like undulose extinction, deformation bands and subgrain formation indicative of dynamic recrystallization. Feldspar show only minor evidences for crystal plastic behaviour like local recrystallization. Kinematic indicators observed in thin sections oriented parallel to the lineation (SC-structures, CC'-structures, mica fishes, rotated clasts) indicate a dextral sense of shear. The stability of chlorite and muscovite during deformation, ductile behaviour of quartz, and dominantly brittle behaviour of feldspar suggest that deformation occurred at greenschist facies metamorphic conditions. Abundant calcite in the mylonites suggests a calcium rich protolith consistent with a calc-alkaline volcanic (andesitic) protolith.
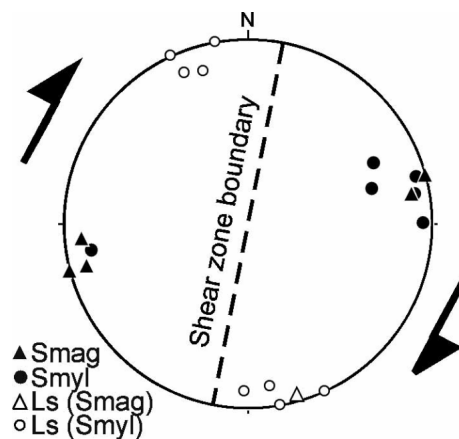

Fig. 4.14: *Late Miocene structures in the Hornopirén area (lower hemisphere, equal area stereoplot). Abbreviations: S_{mag} = magmatic foliation developed within granitoids, S_{myl} = mylonitic foliation developed within greenschist facies mylonites, L_s = mineral stretching lineation. The geometrical relationship of the foliation and lineation with the shear zone boundary indicate a dextral sense of shear.*
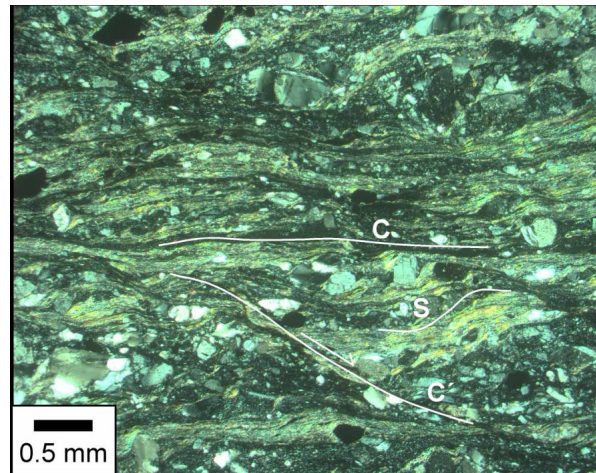
Fig. 4.15: *Micrograph of greenschist facies mylonites in the Hornopirén area (section normal to foliation, parallel to lineation, crossed nicols). SCC'-fabrics indicate dextral sense of shear.*

Miocene granitoids along the eastern side of the Hornopirén fjord show a weakly to well developed magmatic foliation S_{mag} defined by oriented feldspar and biotite which is coplanar with the mylonitic foliation in the greenschist facies mylonites (Fig. 4.14). Mafic (andesitic to basaltic) dikes crosscut the foliated granitoids at a high angle to the foliation (NW-SE) and contain felsic enclaves. Vice versa, granitoids contain mafic enclaves elongate parallel to the foliation (Fig. 4.16). This indicates that magma mingling processes occurred during and shortly after formation of the foliation which is therefore interpreted as a magmatic foliation and underlining the syntectonic character of the intrusion.
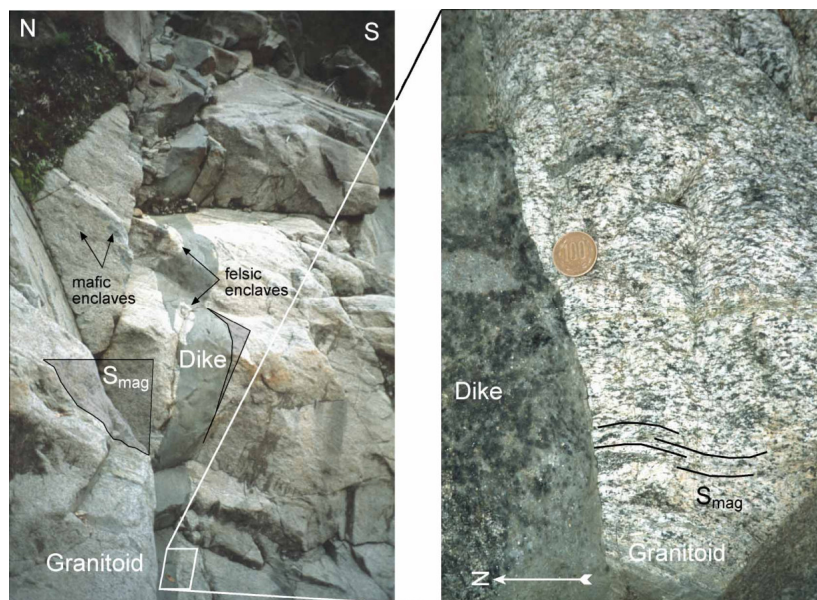

Fig. 4.16: *Syntectonic granitoids in the Hornopirén area. Granitoid host rock shows a N-S striking subvertical magmatic foliation (Smag) which is crosscut by a dioritic dike at a late-magmatic stage. Magma mingling phenomena (felsic/mafic enclaves within, respectively, mafic/felsic host) and the lack of chilled margins at the dike-host contact indicate, respectively, a small rheologic and thermal contrast between dioritic and granitic melts.*

Occasionally, cm-wide, biotite-rich shear zones in the granitoid show a subhorizontal lineation parallel to mineral stretching lineations in the greenschist facies mylonites (Fig. 4.14). In thin section, quartz show features of dynamic recrystallization along those shearzones (Fig. 4.17). This suggests some degree of postmagmatic sub-solidus deformation to have occurred.
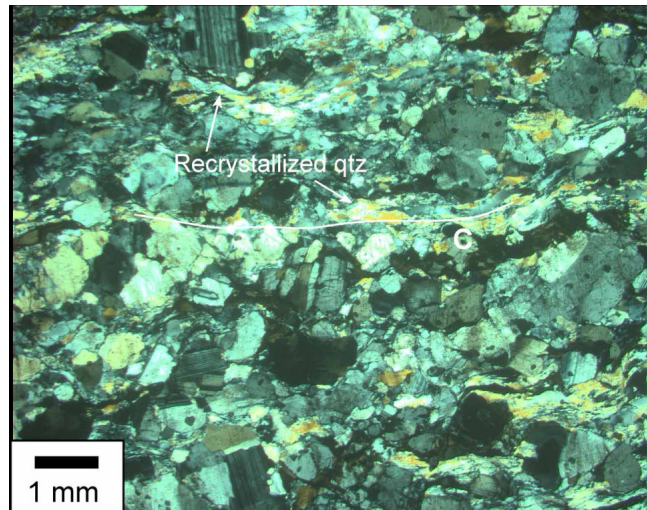

Fig. 4.17: *Micrograph of a submagmatic shear zone in syntectonic granitoids of the Hornopirén area (section normal to foliation, parallel to lineation, crossed nicols)*

4.6   The Liquiñe and Hornopirén areas as examples of mid-crustal tectonics

4.6.1   Mechanisms and kinematics

In contrast to the shallow crustal level tectonics of the Lonquimay area, the Liquiñe and Hornopirén areas show paleotectonics representative for deformation at mid-crustal levels (5 – 15 km). Here, the structural inventory includes up to greenschist facies metamorphosed and ductily deformed country rocks of syntectonic granitoid plutons and subvertical, arc-parallel (N-S trending) brittle-ductile basement faults, and ductile shear zones. The latter occur within Late Miocene granitoids at several scales from mm – km and display mostly dextral shear senses along sub-vertical mineral stretching lineations. Crystal plastic behavior of quartz and dominantly brittle behavior of feldspar during deformation indicate that deformation occurred at greenschist facies conditions (ca. 300 – 450°C, Tullis et al., 2000) which is consistent with crystallization depths of

syntectonic granitoid plutons ranging between 5 - 15 km (Seifert et al., in press) if assuming a geothermal gradient of 30 – 50°C/km.

Brittle ductile faults cutting the basement in an northerly direction (e.g., the Hornopirén faults) are subparallel to the aforementioned ductile shear zones and are thus interpreted as their late-stage, brittle equivalent, which developed at lower temperatures (< 300°C) or higher strain rates. Although the kinematics of these basement faults could not be determined directly, the observation that they bound blocks with different exhumation histories requires a vertical slip component.

The association of dextral shear and rock uplift indicates that transpression, i.e. the combination of pure and simple shear (Sanderson and Marchini, 1984), occurred at mid-crustal levels during the Late Miocene ("Quechuic") increment of active margin deformation in the Southern Andes.

4.6.2 Amount of shortening

To determine semi-quantitatively the amount of cross-arc shortening accommodated by mid-crustal transpression, a simplified two-dimensional model (Fig. 4.18) with the following assumptions has been used:

- All mid-crustal cross-arc shortening is accommodated within the magmatically weakend arc by a divergent pop-up structure, which is bound by two subvertical (80-90° dip), planar inverse faults trending orthogonal to the direction of shortening, i.e. in the direction of the medium strain axis (e_2)

- Plane strain

- Rock uplift is localized along those inverse faults, i.e. no distributed pure shear

Assuming a fault dip of 80° and about 10 km of rock uplift (as inferred from topographic analysis), an amount of horizontal displacement of 3.5 km can be calculated (Fig. 4.18 b) which is the maximum amount of cross-arc shortening accommodated at mid-crustal levels during the Late Miocene increment of deformation. This is considerably less than in the Lonquimay area and suggests that the amount of cross-arc shortening at mid-crustal levels is circa one order of magnitude smaller than at shallow crustal levels.
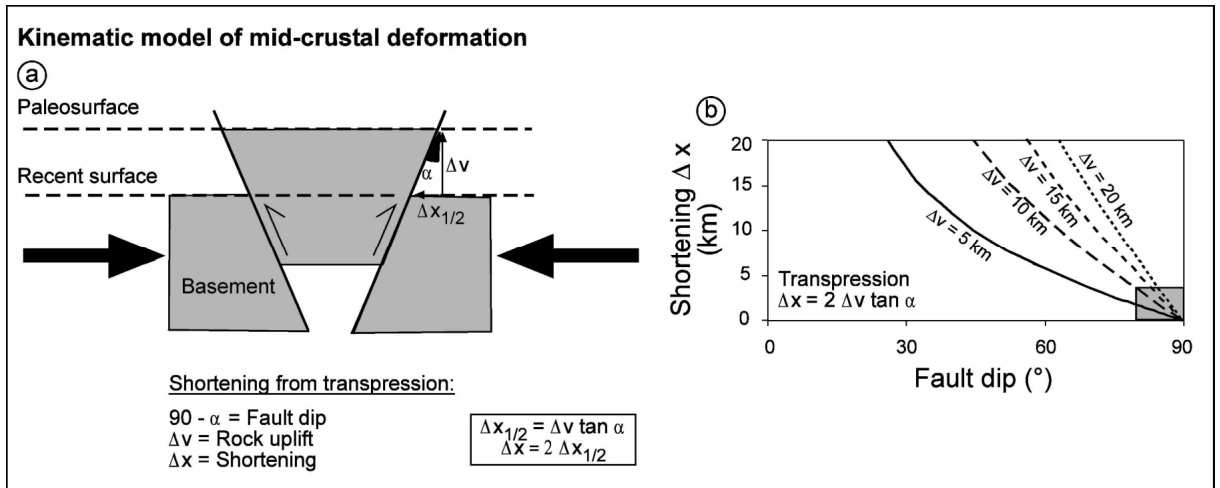
Fig. 4.18: *2-D kinematic model of Late Miocene deformation in the Liquiñe and Hornopirén areas: (a) Concept of the kinematic model; (b) Absolute shortening amount from transpression.*

4.7   Discussion

4.7.1   Transpression at different crustal levels

The Southern Andes provide the opportunity to study continental deformation at different crustal levels. Due to differential post-Miocene and exhumation, structures related to the Late Miocene increment of deformation are exposed at shallow (< 3 km) to mid-crustal (10 – 15 km) levels along strike of the study area (Seifert et al., in press). Tectonics vary accordingly from broad zones of folding and thrusting to brittle-ductile deformation localized within narrow subvertical shearzones (Fig. 4.19 a).
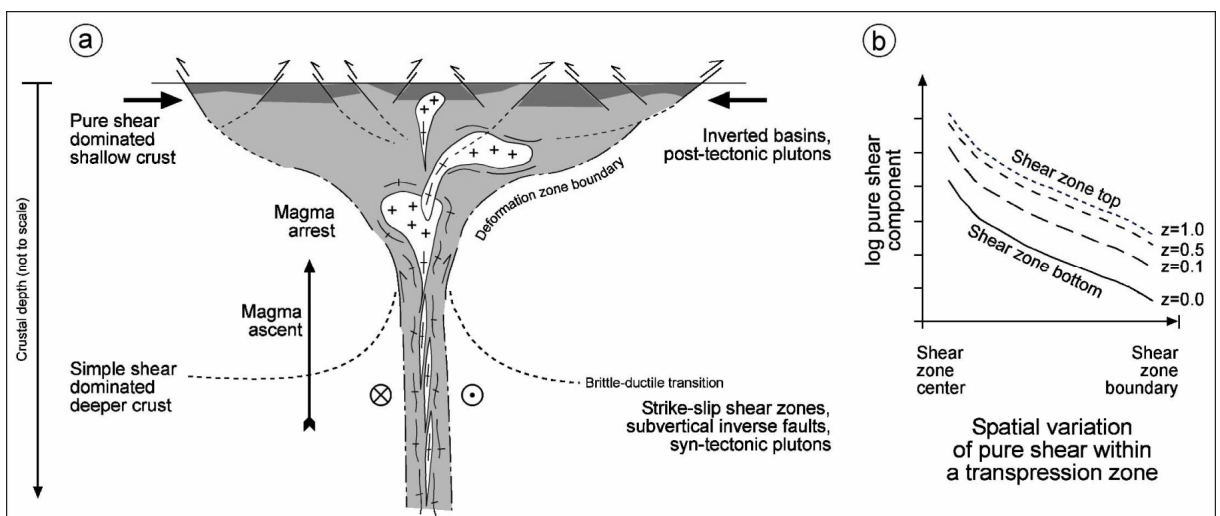

Fig. 4.19: *Paleotectonic model: (a) Geometry and kinematics of a transpression zone, (b) Finite strain within a transpressional shear zone.*

Not only the tectonic regimes, mechanisms, and geometry of deformation vary, also does the kinematics: Whereas pure shear cross-arc shortening dominates at shallow crustal levels, arc-parallel shear with an order of magnitude smaller amounts of cross-arc shortening (i.e. simple shear dominated transpression, Sanderson and Marchini, 1984) occurs at mid-crustal levels (Fig. 4.19 a). This indicates that deformation is heterogeneous and partitioned vertically at a crustal scale. The vertical variability of kinematics observed in the field is consistent with mathematically predicted heterogeneous instantaneous (Robin and Cruden, 1994) and finite (Dutton, 1997) strain distributions during transpression. Fig. 4.19 b shows the results of a simplified model applying equation (1a) of Dutton (1997): Accordingly, in the simplest case of transpression, where the boundaries are vertical and converge at a constant rate, the pure shear component (proportional to the vertical velocity within the transpression zone) is proportional to the height within the transpression zone and proportional to the square of the distance from the boundary. The pure shear component thus increases both towards the center and the top of transpression zones ranging over five orders of magnitude (Fig. 4.19 b).

4.7.2 Synorogenic lateral escape

The change from upper crustal pure shear to lower crustal simple during transpression has an important effect on the crustal scale mass transfer pattern: the change from vertical mass transfer at lower crustal levels to horizontal mass transfer subparallel to the shear zone at lower crustal levels. Because melt flow is driven primarily by pressure gradients generated by buoyancy forces and tectonic stresses (Brown and Solar, 1998 b), deformation not only control tectonic mass transfer but also the mass transfer by magmatic flow. There are several field and examples which show that melt is channelled through actively deforming shear zones (e.g. Tobisch and Cruden, 1995, Brown and Solar, 1999, Druguet and Hutton, 1998, Handy et al., 2001, Mulch et al., 2002, Pawley et al., 2002).

Brown and Solar (1998 a) proposed a conceptual tectonomagmatic feedback model in which melt is pumped through the deformation zone parallel to the finite extension direction. Experimental studies showed that melt pathways may be determined primarily by melt pressure gradients between melt source and melt sink rather than by the orientation of the finite extension direction: Whereas during pure shear the grain scale melt flow direction is subparallel to the finite extension direction (Rosenberg and Handy, 2001), melt migrates oblique to the shear zone boundary and subparallel to the incremental shortening direction during simple shear (Rosenberg and Handy, 2000). Anyhow, tectonic as well as magmatic mass transfer (or flow) may switch from vertical to lateral with increasing crustal depth. Manifestations of such lateral flow in the study

area are syntectonic granitoids of the Hornopirén area. Consequently, it is proposed that upper crustal mountain building is balanced by lower crustal lateral escape.

4.7.3 Internal versus external trigger of subduction orogeny

The tectonic history of the Southern Andes show longlasting periods (tens of millions of years) of extension and basin formation during the Jurassic/Early Cretaceous and Oligo/Miocene alternating with relatively short (several million years) intervals of cross-arc shortening and mountain building (e.g. Kemnitz et al., in press, and references therein). At least during the Cenozoic there is no clear relation ship between convergence parameters, and deformation of the overriding plate. In contrast, phases of extension are coeval with periods of fast orthogonal convergence. This history is inconsistent with geodynamic models linking high rates of convergence to strong compressional coupling (Jordan et al., 2001). Moreover, periods of cross-arc shortening correlate with magmatic pulses during late stages of basin development. This strongly suggests that subduction orogeny may be triggered by magmatic weakening and crustal thinning, i.e. by internal processes, rather than by external forces imposed by plate kinematics. This is consistent with theoretical dynamic models (e.g., Robin and Cruden, 1994, Brown and Solar, 1998 a, b, Saint Blanquat et al., 1998) which suggest that the development of regional scale transpression zones is favored by high T, high melt pressures, and low viscosities, i.e. by magmatic weakening, and also by a high width-to-depth ratio of the deformation zone, i.e. by a wide zone of thinned crust.