



universität
wien

MASTERARBEIT

Titel der Masterarbeit

Fluid assisted cataclastic deformation in quartzitic rocks
(Portizuelo Antiforme, Luarca, NW Spain)

Verfasser

Richard Laner, Bakk.

angestrebter akademischer Grad

Master of Science (Msc)

Wien, 2010

Studienkennzahl lt. Studienblatt: A 066 815

Studienrichtung lt. Studienblatt: Erdwissenschaften

Betreuerin / Betreuer: Univ. Prof. Mag. Dr. Bernhard Grasemann

CONTENTS

Acknowledgements	4
Abstract	5
Zusammenfassung	6
1. Introduction	7
1.1. Aims of the Master thesis	7
1.2. Geographic Overview	8
1.3. Previous Geological Works	9
2. Geological Overview of Iberia	10
2.1. the Iberian Massif	10
2.1.1. Introduction	10
2.1.2. Subdivision	11
2.1.3. Tectonic Terranes of the Iberian Massif	12
2.1.4. Pre-variscan tectonic evolution of the Iberian Autochthon Block	13
2.1.5. The variscan tectonic evolution of the Iberian Autochthon	14
3. The Westasturian-Leonese Zone (WALZ)	17
3.1. Introduction	17
3.2. Subdivision	17
3.3. Stratigraphic succession (Fig. 8)	19
3.3.1. Pre-cambrian rocks	19
3.3.2. Paleozoic rocks	20
3.4. Structure	23
3.4.1. Deformation Phase 1	24

3.4.2.	Deformation phase 2	24
3.4.3.	Deformation phase 3	24
3.5.	Metamorphism of the WALZ	26
3.6.	Alpine Stage	27
4.	The Portizuelo Antiforme	29
4.1.	The Investigation Site	29
4.2.	Stratigraphic Succession and Sedimentary Structures.....	30
5.	Structures observed in the field.....	33
5.1.	Structures related to Hercynian Deformation.....	33
5.1.1.	Geodynamic setting	36
5.2.	Late hercynian brittle deformation	37
5.3.	Alpine brittle structures	38
5.3.1.	Geodynamic setting	44
5.4.	Summary of the structural data.....	45
5.5.	Synoptic 3-dimensional sketch of the outcrop	45
6.	Microstructural investigations.....	47
6.1.	Methods	47
6.2.	Slates of the Transition Zone.....	48
6.3.	Quartzites of the Cabos Series.....	52
6.3.1.	Mineralogical composition	52
6.3.2.	Deformation	53
6.4.	Cataclastic rocks	59
6.4.1.	Components of the cataclasites.....	59
6.4.2.	Matrix	60
6.4.3.	Ultracataclasite	61

6.4.4. Protocataclasite	63
6.4.5. Fluidized cataclasite	64
7. Grain Size Analysis (GSA) of fault rocks.....	65
7.1. Fractal relation D	65
7.2. Procedure	66
7.3. Parameters	67
7.4. Results of the single samples	68
7.4.1. P5	68
7.4.2. P11	69
7.4.3. P8	71
7.4.4. P10	72
7.4.5. P17	72
7.5. Summary of the results.....	74
8. Conclusion	75
References	76
Appendix	82
I. Shape preferred orientation of quartzite	82
II. Spreadsheet for GSA Calculations.....	83

ACKNOWLEDGEMENTS

I started to work on this thesis in 2009 during my stay in Oviedo, Asturias and finished it in July 2010 in Vienna, where I am living at the moment. So there are two groups of people I want to thank for helping and advising me.

Among the people in Oviedo, a special thanks goes to Andres Cuesta, who was my unofficial reference person on the University of Oviedo. He helped me preparing the samples and gave me loads of advice and information. Furthermore I want to thank Fernando Bastida for information on the outcrop, Josep Poblet and Alberto Marcos for their impressive teaching in structural geology and Maria Angeles Fernandez Gonzales for her help in all administrative matters.

On the University of Vienna, great thanks go to my advisor for this thesis, Bernhard Grasemann. He assisted me in the field and on the university and taught me what structural geology really is about. Anytime he was disposable for advice and help. It was a great benefit for me to work with him.

I want to thank the whole Structural Processes Group, which I was part of during my time in Vienna. The discussions with research assistants and study colleagues contributed a lot to this thesis. Among them special thanks go to Jonas Weil, Ulrike Exner, Cigdem Erkmen, Alexander Rath and Norbert Kohlmayer. Also great thanks to Hugh Rice, who had the power of the SEM.

For the help with the Cathodoluminescence microscope I want to thank Dieter Mader. Further thanks to Susanne Gier who did the X-ray diffraction Analysis for me.

My family always supported me during my study time and encouraged me to do good work and to finish my thesis. I owe them great thanks, especially my beloved wife Gabi. She accompanied me during all the time I wrote this thesis. She assisted me in the field, rated my interpretations and checked my English. She is the main reason why one can read this thesis now.

ABSTRACT

The outcrop at the Portizuelo Beach in Western Asturias presents an antiformal bulge of the transition zone between siliciclastic and marine sediments. The core of the antiform comprises of pure, rigid and resistive quartzitic rocks, severely damaged by brittle deformation and cataclasis. Two large transform faults with a particular thrust component can be found in the hinge area. They are clearly in charge of the damage of the surrounding rocks. The faults accommodate the main part of the deformation, but also sub-parallel cataclastic bands show evidence for lateral movement. Originating from the fault planes, fluidized cataclasites pervade the rock mass, leading to further fracturing. Obviously the fracturing ceases with increasing distance from the transform faults. The fluidized material tends to use preexisting planes, such as bed interfaces, joints or veins for its intrusion. Additionally the fluids are responsible for the cementation of the cataclastic zones, generated during incremental strike slip deformation.

Crests of cemented material, cropping out in the surf zone, are linear structures that can be mapped with differential GPS to reveal their spatial distribution. Besides the main bands of cataclastic material also veins showing Riedel-like geometries appear. Furthermore there exists a network-like system that connects the bands. Based on cross cutting relationships a syn-alpine, coseismic formation of the brittle faults and the related cataclasites is suggested.

Microstructural investigations exhibit multiple generations of cataclastic deformation and fluidization events and yield the coherence between them. Grain Size Analysis of binary Back-Scattered Electron and Cathodoluminescence images of cataclastic material clearly shows differences between fault gouges and fluidized cataclasites.

ZUSAMMENFASSUNG

Der Aufschluss am Strand von Portizuelo in Westasturien zeigt eine antiklinale Aufwölbung einer Übergangszone von rein siziliklastischen und voll-marinen Sedimenten. Der Kern der Falte besteht aus sehr reinen, harten Quartziten, die intensiv kataklastisch zerlegt sind. Im Scheitelpunkt treten zwei große Transformstörungen mit Aufschiebungskomponente auf, die eindeutig für die kataklastische, spröde Deformation verantwortlich sind. Durch die Störungen ist ein Großteil der Deformation kompensiert, es existieren aber auch sub-parallele kataklastische Bänder, in denen laterale Bewegung aufgenommen wurde. Von den Kernzonen der Störungen ausgehend, intrudieren kataklastische Dikes in das quartzitische Nebengestein, welche wiederum für weitere Zerlegung des Gesteins sorgen. Das fluidisierte intrudierende Material nutzt vorgegebene Schwächezonen, wie Veinflächen oder stratigrafische Schichtgrenzen, für die Verbreitung. Zusätzlich sind die Fluids verantwortlich für die Zementierung der ansonsten inkohesiven Kataklastite.

Härtlingsrippen aus zementiertem kataklastischem Material, die in der Brandungszone hervortreten, wurden mittels differentieller GPS vermessen, um ihre räumliche Verteilung zu studieren. Neben den Hauptbändern, die dike-ähnlich im Nebengestein stecken und subparallel zu den Seitenverschiebungen verlaufen, tauchen Bänder mit einer Riedel-Geometrie auf. Zusätzlich existiert ein System von kataklastischen Bändern, die quer zu den anderen verlaufen und kinematisch nicht an das Störungssystem gebunden sind. Sämtliche spröden Deformationsstrukturen im Arbeitsgebiet sind auf die alpine Gebirgsbildung zurückzuführen. Weiters stehen die Störungen mit seismischen Ereignissen in Zusammenhang.

Mikrostrukturelle Untersuchungen zeigen mehrere Generationen von kataklastischer Deformation und Fluidisierung. Korngrößenanalyse (KGA) von Rasterelektronen- und Cathodolumineszenz-aufnahmen ergeben deutliche Unterschiede für verschiedene kataklastische Zonen. Protokataklastisches Material kann mittels KGA von ultrakataklastischem oder fluidisiertem Material unterschieden werden.

1. INTRODUCTION

1.1. AIMS OF THE MASTER THESIS

The aim of the thesis is to investigate geological features of the Portizuela Antiform with regard to brittle faults and related cataclastic zones. The antiform is cut by two significant transform faults with thrusting characteristics that are encovered by cataclasitic deformed and heavily jointed quartzitic rocks. Further on, joints filled with cataclastic material crosscut the rock mass up to a certain distance of the faults. Investigation of failure structures will reveal the hitherto unknown origin and formation of the cataclastic zones and the cataclastic veins and their relation to the transform faults.

Macroscopic studies of the rocks in the field, including detailed description of geometry together with a detailed regional mapping of linear failure structures, will give rise to geomechanical considerations. Microscopic examinations of selected samples in thin sections show mineral composition, their relation to the deformation events and the type of deformation. Additionally, the cataclasites are investigated with Cathodoluminescence Microscope and Scanning Electron Microscope (with Back Scatter Electron (BSE) Detector) to reveal internal structures of grains and the chemical composition of minerals. Selected images from both optical microscope and BSE-photographs are used for Grain Size Analysis (GSA), showing whether different cataclastic events have different Grain Size Distribution (GSD) and how they can be related to GSD's given in literature. Another tool that I used is the X-ray powder diffraction analysis. This was carried out to reveal the mineral composition of fine grained material from the fault gauge, from stratigraphic layers and from core zones of cataclastic veins. All together these methods gather information on the failure mechanism in this particular area.

The investigations are made at the geological faculty of the University of Vienna, Austria as well as at the geological faculty of the University of Oviedo, Spain.

1.2. GEOGRAPHIC OVERVIEW

The Portizuelo Antiforme, so called from the beach where it crops out, is situated near the city of Lluarca on the Cantabrian Coast of Asturias in north eastern Spain. The beach of Portizuelo stretches for 1 kilometre, starting approximately 1 kilometre east from the lighthouse of Lluarca. The working area is located at the western part of the beach, where the studied structures crop out in a 40 meter high wall. The geographic coordinates of the centre of the area are 43°32'58" N and 6°31'14" W. Generally, the shoreline in this part of the Cantabrian Coast is made up of several dozen meter high cliffs, embayments and estuaries where rivers disemboque as well as dunes and sand beaches. The remarkable relief originates in the uplift of an old marine platform at the Cantabrian margin with subsequent littoral and fluvial erosion. The local climate is dominated by tempered atlantic domain and therefore mild and humid (Arce, 1997). The mountains of the Cantabrian range act as a meteorological barrier and lead to constant rainfall all over the year. Together with perpetual gales from NE, the coastal erosion rate is considerably high (Fernández Pereiro, 1992). Further information on the development on the relief is given in chapter III.VI.

Politically contemplated, the area is in the vicinity of the city of Lluarca in the judicial district of Valdés which is part of the autonomic region of the Principality of Asturias in northern Spain.

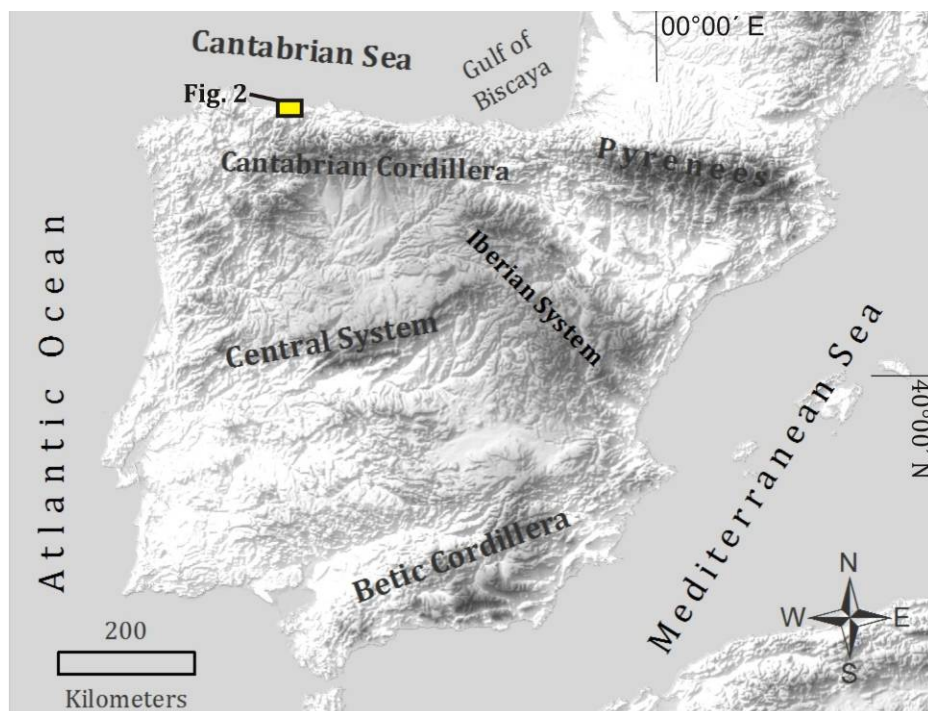


Figure 1: Elevation model of the Iberian Peninsula with the approximate situation of the working area (Based on DEM).

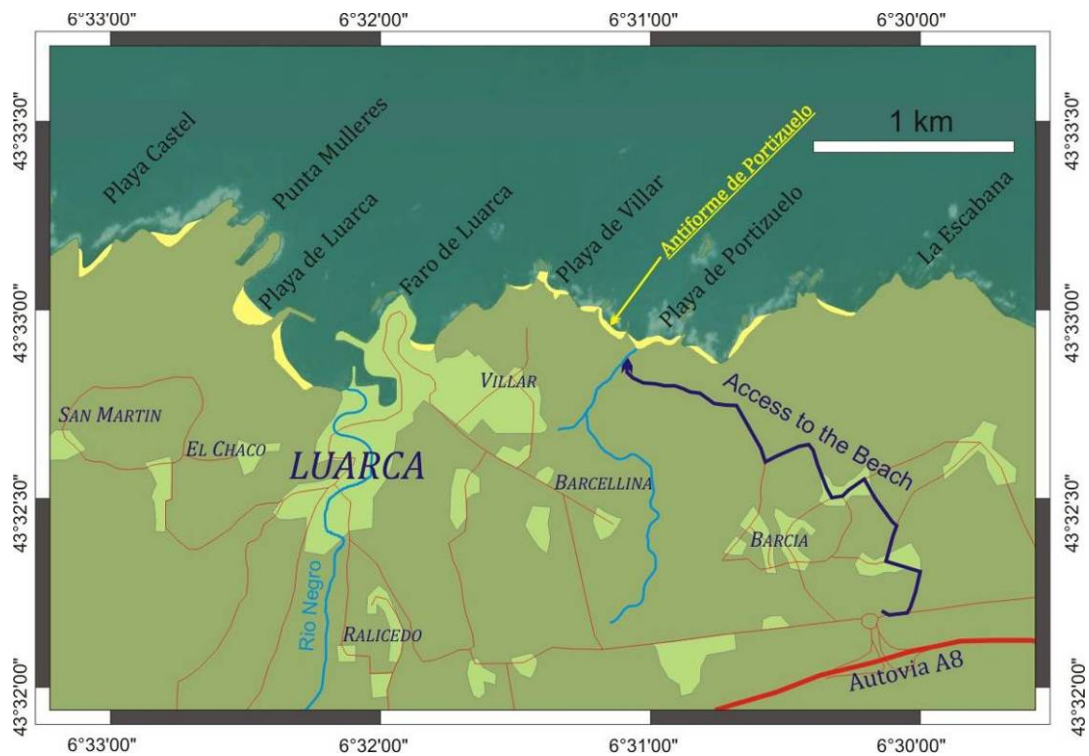


Figure 2: Proximity of the working area (Based on Google Earth).

1.3. PREVIOUS GEOLOGICAL WORKS

The Cantabrian coast, especially its western part, has already been the subject of many geological works. The given relief provides perfect outcrop situations for studying all kinds of geological features. The good infrastructural system permits easy access to the outcrops. But over all counts the fact that, following the coast from west to east, provides a clear section through the internal and external zones of an orogenic belt with all its changes in metamorphism, deformation and stratigraphy. Therefore it is a perfect spot for geoscientists to study and understand the dynamics and geometry of an orogen.

First geological works in the area were carried out in the beginning of the 19th century by the German-born Guillermo Schulz. As mining engineer and commissioner of mining in Spain, he did a lot of work especially around the carbon reservoirs in Asturias and Galicia. In 1858 he published the first geological map of Asturias (see references Schulz, 1858). In the following decades geological work was driven by the industrial interest in resource exploration. In 1959 the actual faculty of geology at the University of Oviedo was founded. Henceforth modern geological investigation began. Important publications concerning the stratigraphic succession in the western part of Asturias are Marcos (1973), Julivert & Truyol (1983) and Pérez Estaún et al. (1992). Details on structural geology, like deformation phases, folding, thrusting and metamorphism, especially in the area around Luarca is described by Bastida & Pulgar (1978) and Bastida (1982). The brittle faults and cataclastic zones in the Antiforme of Portizuelo are mentioned

in Bastida (1982) but have not been described exactly yet. The presented master thesis investigates all structures occurring in the outcrop and concentrates especially on the cataclastic zones.

2. GEOLOGICAL OVERVIEW OF IBERIA

Concerning the main geological zones, the Iberian Peninsula may be divided into three different parts. These are the variscan Orogen, the alpine Orogen and younger basin sediments lying in-between them.

The Iberian Massif is made up of proterozoic and paleozoic rocks, deformed by the hercynian (variscan) orogeny, and incorporations of magmatic intrusions. The actual relief is clearly represented by the Spanish Meseta in Castilia, which is a widespread flat plateau made up of granitic and gneissic rocks. The younger alpine rocks are found in the mountain ranges of the Pyrenees and the Betic Cordillera. Also the Cantabrian Cordillera, although its core is hercynian, achieved its elevation during alpine orogeny. In between these crystalline zones, cenozoic rocks cover intramontaneous basins and rifting zones. Mesozoic sediments can be found among others in the Lusitanian basin in the most western part of the peninsula. Furthermore the continental margins, the Atlantic and the Mediterranean, are considered and treated as stand-alone geological zones.

2.1. THE IBERIAN MASSIF

2.1.1. INTRODUCTION

The Iberian Massif, or also denominated Hesperian Massif, represents in its actual position the most western limb of the European Variscan orogenic belt, resulting from the collision of two continental masses and the formation of the supercontinent Pangea. Laurasia in the north and Gondwana in the south hit each other. As a consequence of this, microcontinents were welded to the main mass and the ocean basins between them were consumed. The convergence in the Variscan belt lasted for more than 150 Ma (between 450 Ma and 300 Ma) and the post-collisional intracontinental tectonothermal events lasted from 380 Ma to 280 Ma (Matte, 1991). The collision between the continents started in the west with the Alleghenian orogeny, reviving the Appalachian system. The west- and central European areas were formed in the hercynian/Variscan orogeny s.s., dominated by accretion of microterranes. As a last consequence of the formation of Pangea, the Ural orogenic belt developed at the edge of Carboniferous/Permian.

2.1.2. SUBDIVISION

The term Iberian Massif is used for the enormous outcrop of proterozoic and paleozoic rocks on the Iberian Peninsula, affected by variscan tectonics (Quesada, 1992). First subdivision in zones was done by Lotze (1945) and further on modified and enhanced by Bard (1969) and Julivert (1971). The actual denomination and the borders of major subzones are shown in Fig. 3.

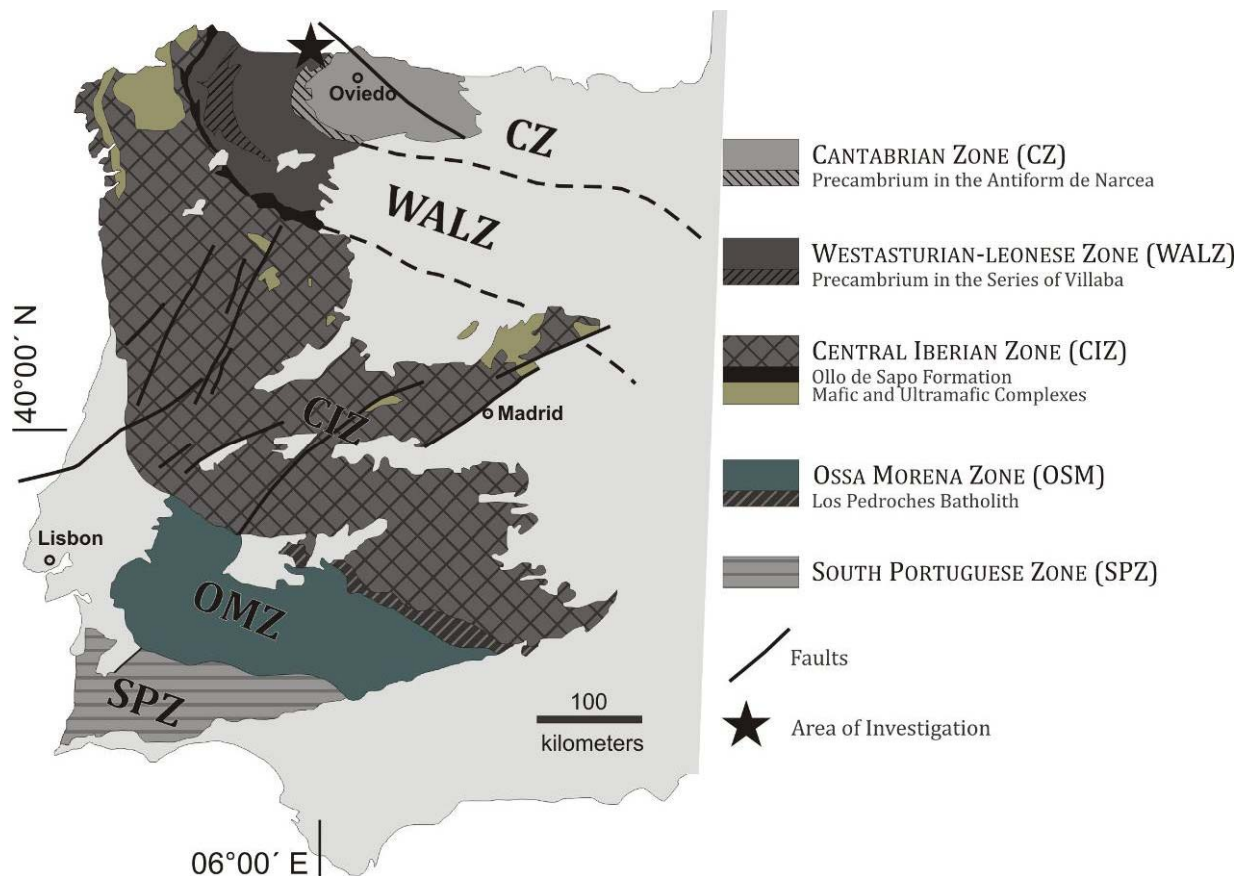


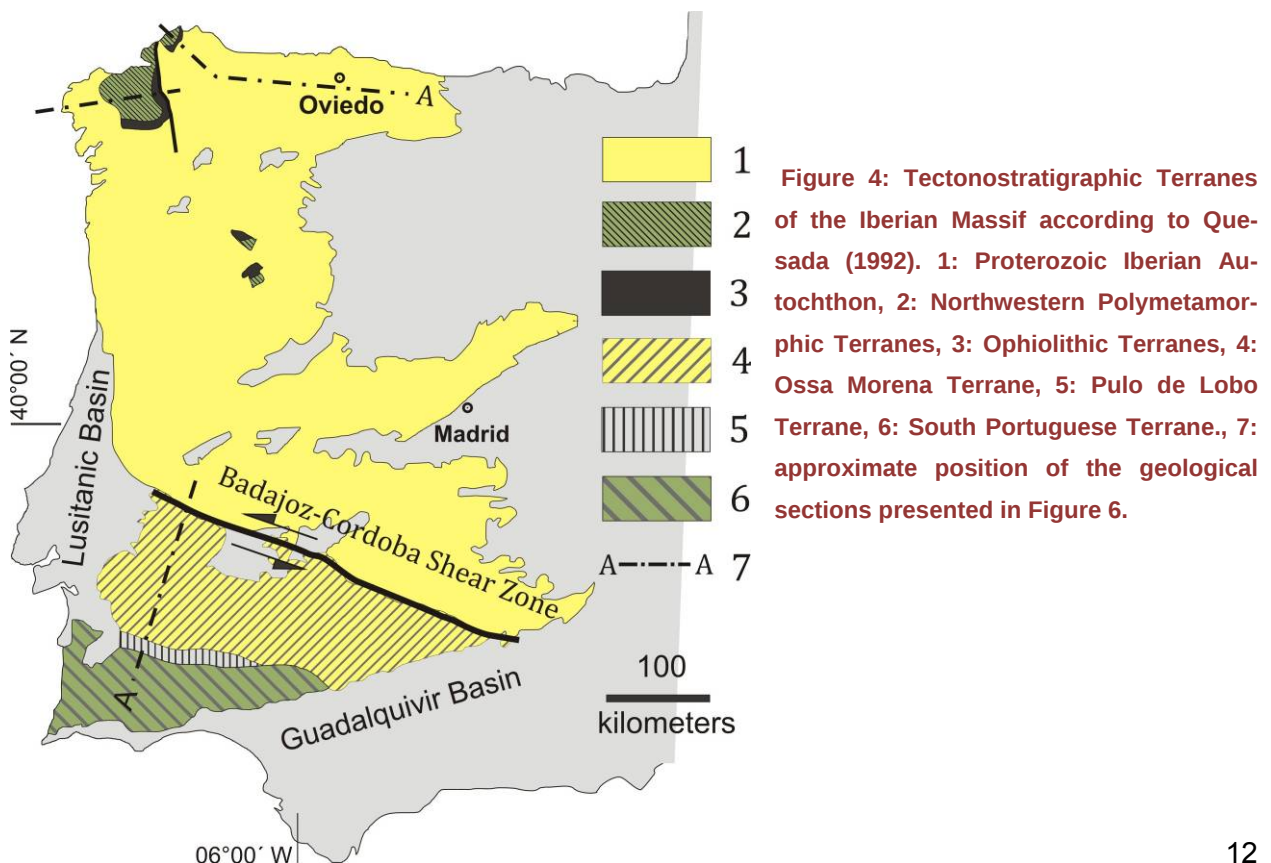
Figure 3: The Iberian Massif and its subdivision in zones according to Julivert et al. (1972). Modified after Bastida & Aller (1995).

Correlation of the variscan zones of the Iberian Massif with that of central Europe, defined by Kossmat (1927), can be done by a clockwise rotation of the peninsula by 37° to the pre-mesozoic position prior to the opening of the bay of Biscaya (García-Mondéja, 1996) and a subsequent connection of the zones. Thus the CIZ and the WALZ can be put on a level with the Moldanubian Zone. The South Portuguese Zone corresponds to the Rhenohercynian Zone. The CZ, considered as the southern fold and thrust belt on Gondwana, can be traced to the Montagne Noir in France. More to the east this zone vanishes underneath the thrusting front of the Alpine Orogen. So after pre-mesozoic reconstruction, the Ibero-Armorican Arc can be con-

sidered as the western syntax of the Variscan Orogen, whereas the variscan areas in central Europe belong to the central orogenic belt s.s. (Matte, 1991).

2.1.3. TECTONIC TERRANES OF THE IBERIAN MASSIF

Apart from the subdivision in zones given in Fig.3, the Iberian Massif can be regarded as a mass of amalgamated tectonic terranes, whereby terranes are considered as geological units separated from each other by tectonic contacts, such as faults (Coney et al., 1980). According to Quesada (1992) the CIZ, the WALZ and the CZ are part of the so-called Iberian-Autochthon Block, considering this terrane as the reference block on which other masses are welded. Another important constituent is the Ossa Morena terrane, tantamount to the OMZ, whose contact to the Autochthon is defined by the Badajoz-Cordoba Shear Zone. The mafic and ultramafic complexes in Fig.3 also are distinct terranes, made up of oceanic crust in the case of the Northwestern Ophiolitic Terranes (Arenas et al., 1986) and multiply deformed metasediments in the case of the superimposed Northwestern Polymetamorphic Terranes, supposed to a Volcanic arc (Ribeiro et al., 1989). The Galicia-Trás-Os-Montes Allochthon is considered as the unity of the Ophiolitic and Polymetamorphic Terranes in Galicia. Further terranes are the Pulo de Lobo Terrane, its mix of sedimentary and ophiolitic fragments originate from an accretionary prisma (Quesada, 1991) and the South Portuguese Terrane. The rocks of this terrane are exclusively of upper paleozoic age and are correlated to formations in the Rhenohercynian Zone. A graphic overview of the mentioned terranes is given in Figure 4.



2.1.4. PRE-VARISCAN TECTONIC EVOLUTION OF THE IBERIAN AUTOCHTHON BLOCK

The Iberian Autochthon is made up of continental crust, formerly part of the West African craton. Evidence for this is given by Cadomic rocks of the upper Proterozoic in some formations, which are related to the Panafrican orogeny (Quesada, 1992). The exact age of the deepest basement units cannot be revealed easily. Some authors measured the ages of zircons obtained from proterozoic granites, supposed to be anatectites from lower basal crust. By this, Schäfer et al. (1988), for example, gained U-Pb ages of about 2 Ga which also correlates with data received from west-african granites.

The Pre-cadomic evolution was dominated by the tectonic regime of a passive continental margin (Quesada, 1992). The rocks of those times are mainly terrigenous sediments with pelitic-aluminous characteristics. During the Cadomic orogeny (650-550 Ma) the Ossa Morena Terrane was amalgamated and the passive margin characteristics changed to back-arc basin conditions (Quesada, 1989). The turbiditic series in the basal units of the Complejo Escabioso Grauváquico in the CIZ represent the associated syn-tectonic sediments (Quesada, 1990). In the following Paleozoic era, prior to the hercynian collision, the Iberian autochthon block embodied a vast continental platform affected by several extensional events. Rifting events, beginning at the edge of Precambrian/Cambrian, gave rise to a Horst and Graben structure in an epicontinental sea. Actual zonation of the autochthon block (see Fig.3) is due to the segmentation in basins and ranges whilst this rifting event (see Fig. 5).

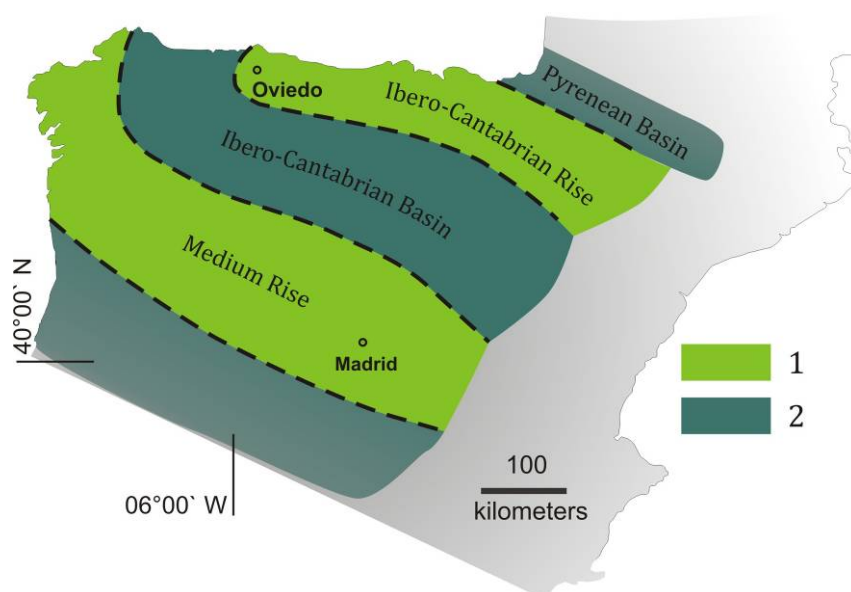


Figure 5: Paleogeographic reconstruction of the northern Iberian peninsula in the Lower Ordovician. The arch-shaped form developed in the Carboniferous, during the Variscan orogeny, and is denominated as Asturian Arc or also Ibero-Armorican Arc.

1: Emerged land, 2: Epicontinental Sea (sedimentary basins). Modified after Aramburu et al. (1992).

The WALZ, for example, represents a deepening basin with cambrian and ordovician syn-rift sediments of notable thickness, whereas the CZ acted as rise and contains lower paleozoic sediments of only minor importance (Aramburu et al., 1992). Material transport from the Ibero-Cantabrian Rise into the adjacent basins is proved by the deposited sediments. The existence of the Medium Rise is still uncertain. In marginal areas of the continent, especially near the recent amalgamated Ossa Morena Terrane, extensional bimodal magmatism was active during the lower Paleozoic. Some authors mention the possibility of the formation of an ocean basin due to proceeding thinning of the crust at this time (Quesada, 1991). The epicontinental platform conditions endured for more than 200 million years with only little variation. After the relative unstable rifting stage in the Cambrian, the continent margin became more stable in the Ordovician. Terrigenous sedimentation was predominant and traces of upper Ordovician glaciomarine rocks prove the subpolar, perigondwana position (Quesada, 1992). Extensional tectonics set back in the Silurian and affected again the marginal areas, especially the schistose domain of the Galicia-Trás-Os-Monte Zone. Once more persistent thinning of crust led to magmatism and even to the formation of tholeiitic rocks, regarded as evidence for the initial formation of oceanic crust (Ribeiro, 1987). During the lower Devonian a perpetual uplift of the continental platform, associated with the approaching hercynian orogen-wave, reduced the epicontinental sedimentation and finally stopped in middle/upper Devonian period with the initiation of the orogeny (Quesada, 1992).

2.1.5. THE VARISCAN TECTONIC EVOLUTION OF THE IBERIAN AUTOCHTHON

As already mentioned, the Iberian Massif is a melange of different continental masses, welded together during the hercynian orogeny, i.e. the approach and final collision of Gondwana and Laurussia (Matte, 1991). The in between lying oceanic basins, partly formed during early-paleozoic rifting, or already existing since the Proterozoic, were consumed by subduction or, in special cases, obducted on the continental plates.

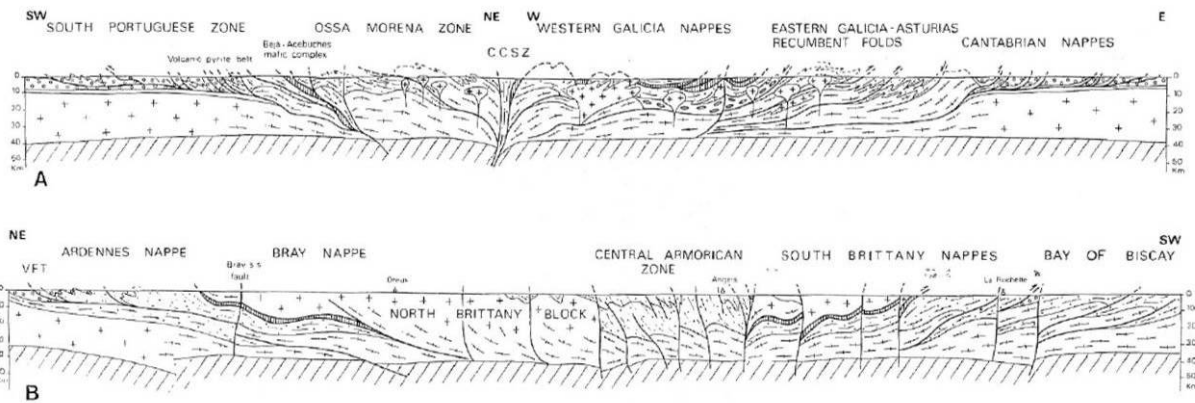


Figure 6: Section through the Iberian Massif after Matte (1991). The fan-like form in the section is typical for the Variscan belt and is considered as evidence for orogeny by lithospheric subduction. The approximate position of the section is shown in Fig. 4.

Two suture zones, evidences for subducted crust, are exposed in the Iberian Massif, although their timing and polarity differ significantly (Quesada, 1992). The Badajoz-Cordoba Shear Zone is a wide, complex, intracontinental shear zone with a sinistral offset of more than 200 km. It represents a major suture between Ossa Morena and the Autochthon block (Matte, 1991). The southern suture can be found in the Beja-Acebuches ophiolite-arc complex (or also Ossa Morena suture), thrust on the Polo de Lobo Terrane between the Ossa Morena Zone and the South Portuguese Zone. The polarity of the northern suture is nearly vertical to south dipping and therefore marks the Autochthon as foot wall block, i.e. the subduction direction is to the south. So the active continental margin was at the side of the northern exotic terranes (Galicia-Trás-Os-Monte Allochthon). The southern suture, made up of a thick sequence of ophiolitic rocks, dips to the north, indicating subduction in the opposite direction. The oceanic crust was consumed under the Autochthon (see Fig. 6). The suture zones differ also in their timing. The continental margin of the northern exotic terranes became active in Ordovician (Peucat et al., 1990), whereas the subduction of the southern ocean began much later in the Lower Devonian (Andrade et al., 1991). The accretionary history of the northern part of the Iberian Massif (without Ossa Morena and South Portuguese Zone) can be described as follows:

- During approach of the Iberian Autochthon, as a part of Gondwana, to an active continental margin of a volcanic arc (NW-Polymetamorphic Terrane), oceanic crust was obducted (Ophiolitic terrane). The age of the initial continent collision and obduction is about 380 Ma (Matte, 1991).
- The ongoing collision tendency transported the Allochthon nappes far over the continent. The orogenetic wedge gained height through tearing of marginal basal units of the Autochthon (Galicia-Trás-Os-Monte Zone and Westasturian-Leonese Zone) (Quesada, 1992).

- The external zones of the passive continental margin transformed in a back-arc basin (Cantabrian Zone) after the rise of the continental margin.
- The arc-shaped form was developed in the Carboniferous due to the irregular shape of the colliding continents. A promontory of Gondwana, namely Iberia, hit the continent and by bilateral escape of crust, the actual curvilinear shape was formed.
- Implantation of S-type Granites occurred in the time from 350 to 280 Ma.
- The subduction ceased finally and led to a collapse of the thickened crust and the formation of late carboniferous (Estefian) to permian intracontinental basins (Matte, 1991).
- Latest granitic intrusions took place as consequence of the thinning crust from 290 to 280 Ma.

3. THE WESTASTURIAN-LEONESE ZONE (WALZ)

3.1. INTRODUCTION

As seen in Fig. 2, the north-western part of the Iberian Peninsula is made up of three main geological zones (see also Section in Fig. 6):

- The Cantabrian Zone (CZ), representing the most external zone with the development of mainly non-metamorphic upper Paleozoic sediments. The tectonic deformation only affected the sedimentary cover in upper crustal levels (i.e. “thin-skinned tectonics”) and no schistosity is developed.
- The Westasturian-Leonese Zone (WALZ) comprises of a thick sequence of lower paleozoic rocks affected by regional hercynian metamorphism which is overlying the pre-cambrian non-schistose rocks with unconformity. Orogeny affects the sedimentary cover as well as the underlying basement (i.e. “thick-skinned tectonics”). Metamorphism augments from east to west from epizonal to mesazonal (sub greenschist to amphibolit fazies). Additionally granitic rocks appear and tectonic schistosity is well developed.
- The Central Iberian Zone (ZIC), as most internal zone of the hercynian cordillera, is characterized by the abundant emplacement of syn-orogenic granites. Metamorphism locally reaches catazonal levels (granulit and eclogit fazies).

3.2. SUBDIVISION

The definition of the Westasturian-Leonese Zone was first established by Lotze (1945) as an arc-formed band with a more or less complete succession of lower Paleozoic sediments, showing the equal tectonic deformation and a reduced presence of granitic rocks compared to the CIZ. Later, Matte (1968) and Julivert et al. (1972) modified the definition slightly and finally Marcos (1973) introduced its subdivision into 3 domains with notable stratigraphic differences (outlined in Fig. 7). The boundaries to the adjacent zones of the Iberian Massif are roughly constituted by two big anticlinoria. The Narcea Antiform in the east and the Ollo de Sapo Anticlinorium in the west. The exact pathway of the boundary in the Narcea Antiform is defined by a within lying, dominant thrust fault. In the case of the Ollo de Sapo Formation the border follows a line, given by the Viveiro fault and the Sil- Truchas Synclines (see Fig. 7). The WALZ covers territory in western Asturias, north-eastern Galicia and in the north-western part of the province of Leon,

where it also vanishes eastward underneath the paleogen cover. More to the east it crops out again in Burgos and Aragón, where it is only of minor importance (Pérez-Estaún et al. 1992).

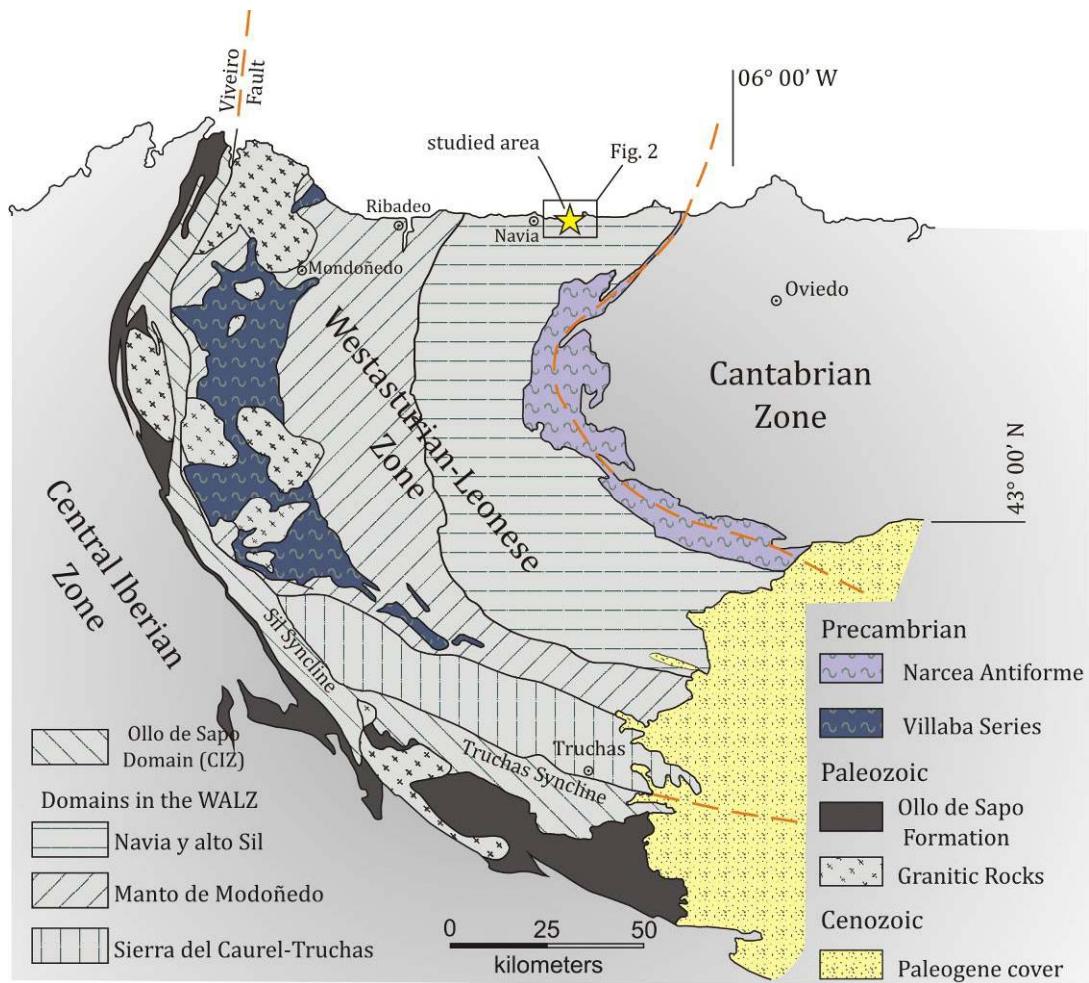


Figure 7: Subdivision of the WALZ in domains after Marcos (1973). Western boundary after Pérez Estaún et al. (1992).

The domains roughly distinguish themselves from each other by the following characteristics (Marcos, 1973):

- Navia y alto Sil Domain: Big lower Paleozoic depression with complete stratigraphic sequence from lower Cambrian to upper Ordovician rocks.
- Manto de Mondoñedo Domain: Represents an allochthon sheet ("Manto" en castellano) thrust on the latter mentioned domain. The lower Paleozoic sequence is less potent and not as complete either.
- Caurel-Truchas Domain: The most internal part of the WALZ. It differs from the latter domain by Cambrian rocks of even minor importance but the existence of lower Devonian rocks in an important recumbent fold (Caurel Syncline).

3.3. STRATIGRAPHIC SUCCESSION (FIG. 8)

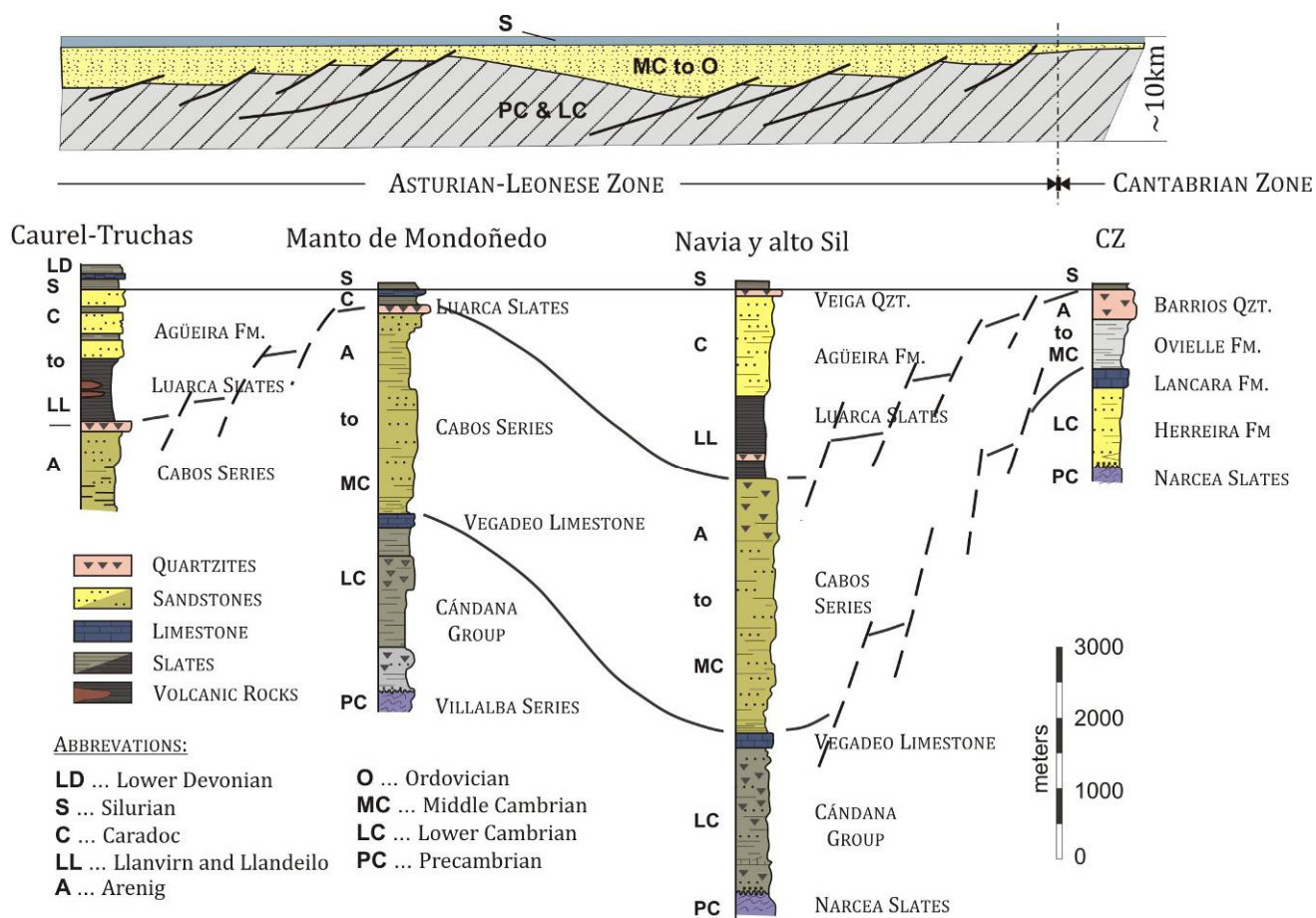


Figure 8: Stratigraphic succession and idealized E-W-section of the "Asturian Arc Basin before hercynian deformation"; modified and redrawn after Pérez Estaún et al. (1992)

3.3.1. PRE-CAMBRIAN ROCKS

Upper Proterozoic rocks can be found in the Narcea Antiforme and in the Villaba Series. As their development is very similar, they are described together in the following. The Pre-cambrian rocks are mainly weak metamorphic, greenish sandstones and slates (Pérez Estaun & Martínez, 1978). Frequent slump structures, flute marks and the occurrence of Bouma-sequences indicate the turbiditic character of the sediments (Pérez Estaun, 1973). Additionally concordant intercalations of igneous rocks of volcanic and volcanoclastic nature can be found (Matte, 1968). Microfossils, mostly cyanobacterias, yield Upper Precambrian ages. The Pre-cambrian sequence is separated from the upper Cambrian series by a remarkable angular unconformity which can be compared to the asyntic (Cadomian) discordance (Aramburu, 1995). The deformation, however, produced large -scale recumbent folds without the generation of any tectonic foliation. Neither pre-cambrian metamorphism affected the rocks (Pérez Estaun, 1973).

3.3.2. *PALEOZOIC ROCKS*

The Paleozoic succession comprises of a complete sequence from the Cambrian to Ordovician, in some areas even up to lower Devonian. Over it, the postorogenic Stephanian is located with an angular unconformity (Pérez Estaun et al., 1992). The sequence in the Navia y Alto Sil Domain (NASD) represents the most complete succession. As the working area is situated in this area, only this succession will be described in detail.

The Cándana Group

This formation is made up of feldspatic sandstones alternating with greenish and reddish slates. The base of the group, just atop the discordance with the Precambrian, comprises of conglomerates or in some cases of dolomitic rocks (Marcos, 1973). Microconglomerates appear repeatedly towards the upper part of the group (Pérez Estaun et al., 1992). The sedimentary environment is supposed to be a shallow-water marine facies in continental environment. The thickness reaches up to 2000 meters in the syncline of Cabo Vido and also in the Gistral Tectonic Window in the Manto de Mondoñedo domain. Paleontological studies on trilobites and archeocyathides from the top of the group reveal ages of Early-Lower Cambrian (Sduzy, 1961). Ichnofossils from the basal part indicate lower cambrian to even precambrian ages (Crimes, 1987).

The Vegadeo Limestone

The contact of the latter group to the Vegadeo limestone is gradually distributed in a transition zone in the upper part of the Cándana group. The Vegadeo limestone comprises of dolomitic and carbonatic beds, which arose from sedimentation in a tidal flat facies or shallow-water facies (Pérez Estaun et al., 1992). During the hercynian orogeny the rocks recrystallized almost completely in order that the primary components, i.e. fossils, cannot be traced anymore. However, the possible age was narrowed down by fossils from the units next to the lower and upper limits of the formation. Basal beds, together with the Cándana transition beds are of early-lower Cambrian and the top beds are supposed to be of middle-Cambrian age (Pérez Estaun et al., 1992).

The Cabos Series

The cambro-ordovician aged Cabos Series is composed of a thick sequence (more than 4000 m) of shallow marine clastic sediments (Baldwin, 1975). Lotze (1957) introduced the denomination "Serie de los Cabos", according to the abundant appearance of this formation around Capes (=Cabos) along the Cantabrian coast. This is due to the relatively high resistance of the comprising rocks against weathering. Outcrops of this series can be found all over the Westasturian-Leonese Zone, but anyhow the best sequence can be contemplated in the section between Cadavedo and Luarca in the Navia y alto Sil Domain (Färber & Jaritz, 1964). The stratigraphic succession will be described for this area.

Armorican Quartzite in the Variscan Cordillera. The Cabos Series in the vicinity of Luarca, where parts of the structural investigations for this work have been carried out, is represented by idem Barayo-Quartzite.

The Luarca Slates (“Pizarras de Luarca”)

The extensional regime in the passive continental margin continued during Ordovician, leading to depressions filled with marine deposits with open shelf facies. The siliclastic sediments of the Cabos Series are followed by homogeneous pyrit-rich black slates, deposited in euxinic environment.

These slates are denominated Luarca Slates (Barrois, 1882) and reach the maximum thickness of about 1200 m in the Navia y alto Sil Domain (Marcos, 1973). In the proximity of Luarca the series can be subdivided into three parts, as proposed by Marcos (1973) (see Fig. 9). The Lower Member and the Upper Member comprise of lustrous black slates with regular intercalations of quartzitic and Fe-bearing layers. The Middle Member is made up of white quartzites.

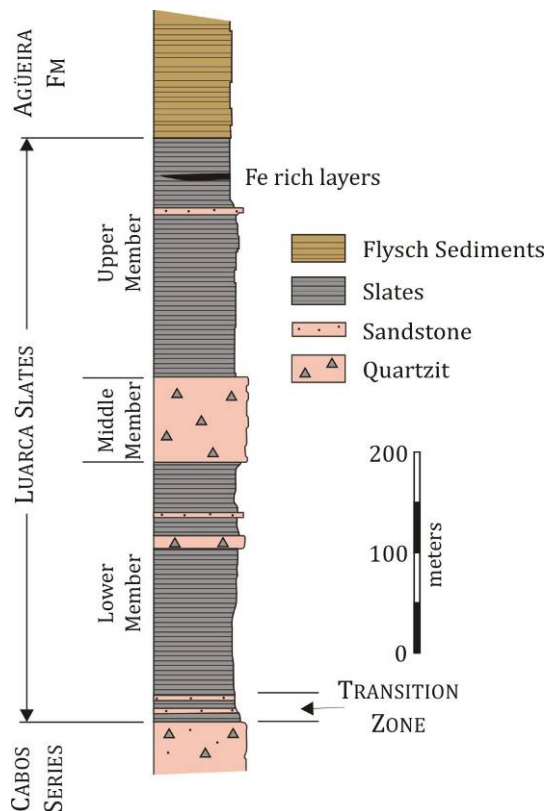


Figure 10: Stratigraphic subdivision of the Luarca Slates Series in Upper, Middle and Lower Member. The stratigraphic profile is from the coastline around Luarca. The Middle Member quartzite is denominated Sabugo Quartzite, according to the beach of Sabugo in the east of Luarca. Redrawn and modified after MARCOS (1973).

The facies of the Luarca slates (black slates characterized by little amounts of clastic material, richness in organic matter and the lack of carbonate) proposes deposition in euxinic media. The occurrence of well conserved graptolites indicates calm waters. Consequently the sedimentation took place in a deepening basin bounded to the open ocean by a submarine barrier, providing calm and anoxic deposition conditions. Evidence for the great profundity of the basin comes from the fact that the superposing units of the Agüeira Formation already have turbiditic characteristics (Marcos, 1973).

Paleontological data from the lower units of the Luarca slates yields lower Ilanvirn ages and the Upper Member yield ages up to Landeilo (Pérez Estaun et al., 1992). The transition between the Luarca Slates and the Cabos Series is gradual. Mostly areniscas appear which alternate

with finely laminated slates. The outcropping units at the beach of Portizuelo are part of the Lueca Slates, the Transition Zone and the Cabos Series.

The Agüeira Formation

In the Middle Ordovician the central part of the WALZ suffers an abrupt increase in subsidence. Consequently a basin was formed that was further on filled with syn-orogenic sediments of the Agüeira Formation. Metasandstones and slates are the main constituents of this series, which are very often found in turbiditic sequences. The middle part of the formation was dated with trilobites and brachiopods to caradocian age (Pérez Estuán et al., 1992).

Silurian rocks

After the pre-orogenic ordovician deposition, the advancing orogenic wave hampers further subsidence. The depositional cycle ends with the sedimentation of Silurian black metapelites. Their thickness is significantly lower than in the underlying sediments. It reaches only about 700 m at its maximum (Marcos, 1973). The general situation after silurian sedimentation is shown in the section of the Asturian Arc Basin in figure 8.

The syn- and post-orogenic sediments are not mentioned in this work, because they do not crop out in eastern and central Asturias. Additionally their regional appearance in the WALZ is only of minor importance, whereas mesozoic and cenozoic deposits play a major role in the geology of the Cantabrian Zone.

3.4. STRUCTURE

In the beginning of the Cambrian the paleozoic sediments were folded, fractured and have undergone metamorphism in the course of the hercynian deformation. The consequently generated structures can be ordered in 3 groups which indicate 3 different stages of deformation. All these stages are considered to have lower carboniferous age (Perez Estuán, 1973). The dating of micas, which are growing under metamorphic conditions in the cleavage planes, yields the latter mentioned ages. Lower carboniferous syn-orogenic flysh can be found as lower limit and post-orogenic upper westfalian B deposits provide an upper marker (Martinez Catalan et al., 1990).

The earliest deformation structures are large scale east-vergent recumbent folds. In a second stage single sheets were transported along thrust faults to the foreland basin. Finally large open folds with sub-vertical axial planes and N-S-striking fold axes were developed (Martinez Catalan et al., 1990).

3.4.1. DEFORMATION PHASE 1

As a first consequence of the continental collision, the paleozoic sediments were folded to large recumbent folds with limbs of more than 30 km length. A primary cleavage S1 was developed as a consequence of this folding. Martínez et al. (1990) distinguish two types of major F1 folds. One set with constant wavelength and minor scale, appearing mainly in the NASD and the other set of folds with huge wavelength to which belong the Mondoñedo-Lugo-Sarria Anticline and the Villadrid syncline. Secondary folds that occur are generated by buckling, the major folds are affected by a superimposed flattening that increases to the W (Bastida, 1980). The associated cleavage S1 is a fine-grained slaty cleavage in pelitic rocks. In more competent lithologies it is a rough spaced cleavage. The shape of the major and the minor folds proposes a generation mechanism of the whole F1 structures similar to that of the morcles nappe proposed by Ramsay et al. (1983): The base of a thick multilayered sequence is affected by a generalized horizontal shear. Some local inclined shear zones evolve from the base and produce long-limbed recumbent folds. Bastida et al. (1986) propose at least two of such local shear bands in the WALZ.

3.4.2. DEFORMATION PHASE 2

The second deformation phase F2 is considered as the continuation of the basal shear zone, developed during the first phase. Major structures generated are thrust faults. The most important fault is the Mondoñedo basal thrust (see fig. 11) that juxtaposes the lower units of the MD with the upper units of the NASD. Minor structures related to the thrust faults are fault rocks of all types. In the basal shear zone minor folds and a S2 cleavage appears (Martínez Catalán et al., 1990).

3.4.3. DEFORMATION PHASE 3

Open, large parallel folds are the consequence of the third deformation phase F3. The Fold axis is similar to F1 folds and strike NE-SW. The superimposing of F1 and F3 deformation produces a type 3 refolding structure on the major scale and crenulation cleavage on the minor scale. Crenulation cleavage is developed in pelitic rocks and a rough cleavage in psammitic rocks. An important structures are in the MD is the Lugo Dome, which is a interference structure of F1, F2 and F3 deformation (see fig. 11).

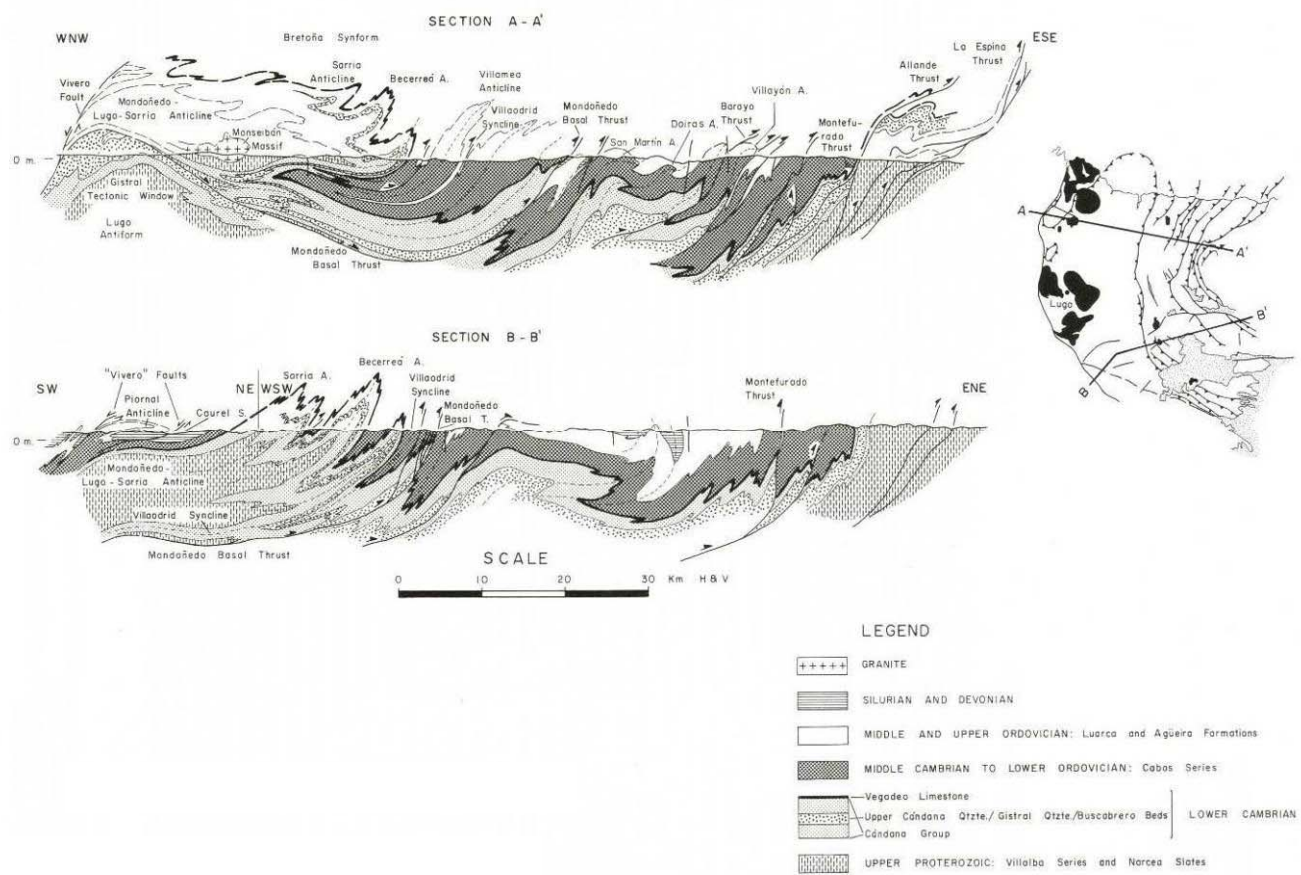


Figure 11: Geological sections through the WALZ; Martínez Catalán et al. 1990.

3.5. METAMORPHISM OF THE WALZ

Travelling through Asturias from east to west, one passes the Variscan orogen from its internal to its external zones. Consequently all orogeny-related processes cease from east to west (as already mentioned in prior chapters). Metamorphism, of course, shares this relation too. The two main units of the WALZ, the Manto de Mondoñedo and the Navia y Alto Sil Domain, differ significantly in their metamorphic overprint. In the east, respectively in the MD, medium to high metamorphic conditions are achieved, whereas in the NASD the rocks are only affected by low grade metamorphism (Suárez et al. 1990). Three metamorphic belts cut through the WALZ from north to south. They define local zones with distinct higher grades in metamorphism. An overview of the metamorphic zones is given in Figure 12 below.

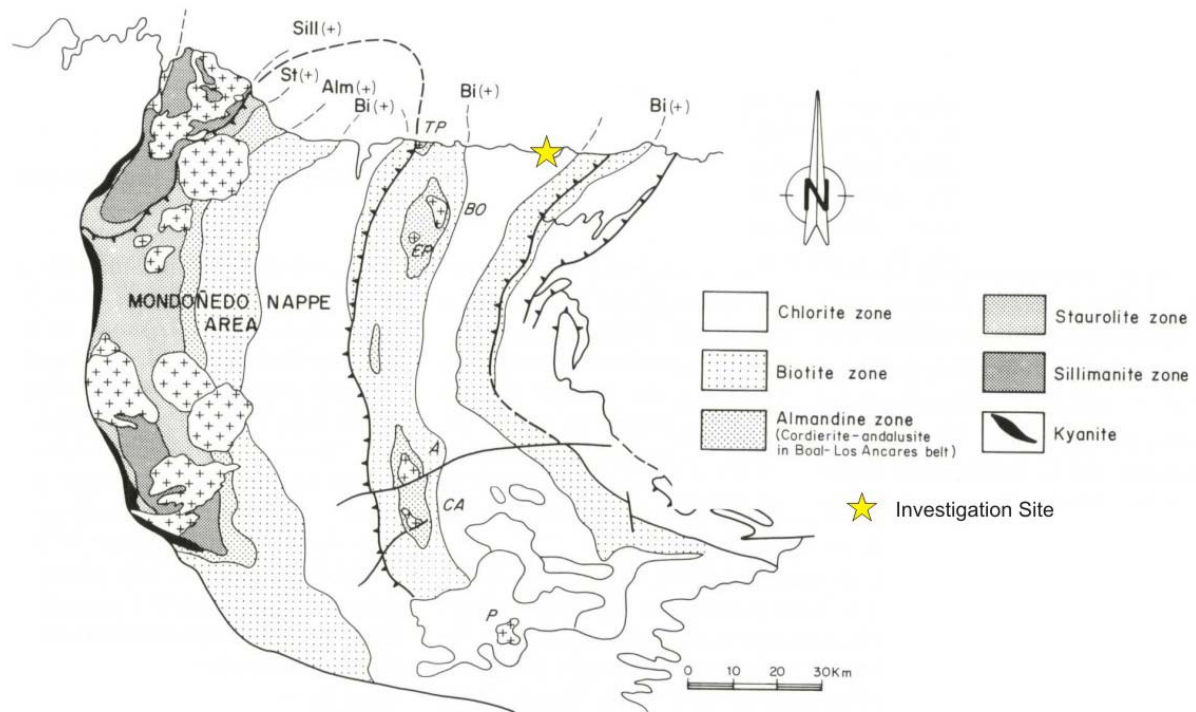


Figure 12: Distribution of the metamorphic zones in the WALZ (modified after Suárez et al, 1990).

The Chlorite zone is mineralogically defined by the paragenesis of muscovite and chlorite, which have grown syn-tectonically in the F1 phase. Furthermore, chloritoid appears occasionally, forming porphyroblasts which show syn- and post-tectonic growth related to F1 (Marcos, 1973). The Biotite zone is best developed in the pelitic precambrian rocks of the Novellana-Pola de Allende-Degaña belt. The shape of the belt is defined by a N-S striking F3 antiformal structure. Biotite grows mimetically in respect with S1-cleavage and additionally muscovite, chlorite and recrystallized quartz occur. They are also related to the first phase of deformation (F1) (Suárez et al., 1990). Higher metamorphic grades are achieved in the Boal-Los Ancares belt and in the Vivero-Lugo-Sarria Metamorphic Belt in the West. In these belts, several characteristic

paragenesis can be distinguished, one of them even including kyanit which indicates a high pressure metamorphism (Suárez et al. 1990). Metamorphism in these belts takes place during a time period that includes all three hercynian deformation phases. Furthermore late variscan plutonic rocks occur in the metamorphic belts. Within their proximity the host rock suffered contact metamorphism resulting in the F3 syn-tectonic growth biotite and garnet.

The metamorphism was limited to a minimum age (Sakmarian, 285 Ma) by measuring the age of the granitic intrusions in the Boal-Los Ancares Belt (Suárez et al., 1978). A maximum age of Mid-Devonian (Emsian, ~ 400 Ma) is proposed by Drot and Matte (1967), taking into account folded and S1-foliated Emsian stratas in the Mondoñedo Nappe.

3.6. ALPINE STAGE

In the course of the alpine orogeny in the Eocene, the long since eroded variscan mountains experienced a general uplift. Along an E-W-striking fault system, the variscan units have been segmented and elevated relatively to each other. This segmentation provides the basis for the actual relief of the Cantabrian Cordillera (Farias & Marquínez, 1992). The Cantabrian Cordillera itself is regarded as the continuation of the Pyrenean mountain range to the west, arisen from the alpine continental collision (Quesada, 1992).

In an N-S section of Asturias and the adjoining northern parts of Leon, three blocks can be determined, which are separated by dominant alpine thrust faults (Fig. 13). The Cantabrian Mountains, defining the major block, are thrust on the Iberian Meseta and are now juxtaposed to mesozoic molasse sediments and non-deformed cenozoic sediments of the Duero Basin (Anton et al., 2010). More to the north a minor block is thrust on the cordillera block, giving rise to the Medium Depression which is also filled with syn-orogenic sediments. The minor block constitutes of the prelitoral mountain range, the coastal plain and the actual marine platform.

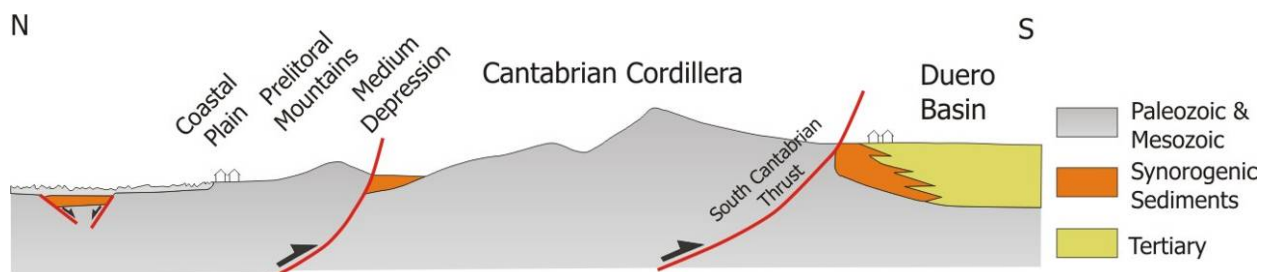


Figure 13: Sketch of variscan blocks segmented by alpine faults in an N-S section through Asturias. Modified and redrawn after Farias & Marquínez (1992).

Figure 14 shows the tectonic pattern of the major alpine structures in northern Spain. Thrust faults generally strike E-W. Within them the main part of internal deformation during alpine N-S shortening is accommodated (Anton et al, 2010). Additionally, alpine folding is recorded in sev-

eral areas, especially in the calcareous mesozoic sediments of Cantabria and the Basque Region. A dominant system of strike slip faults crosscuts the thrust fault system and lead to further segmentation and lateral escape of tectonic blocks (Farias & Marquínez, 1992). Altogether, a late alpine E-W extensional regime is indicated.

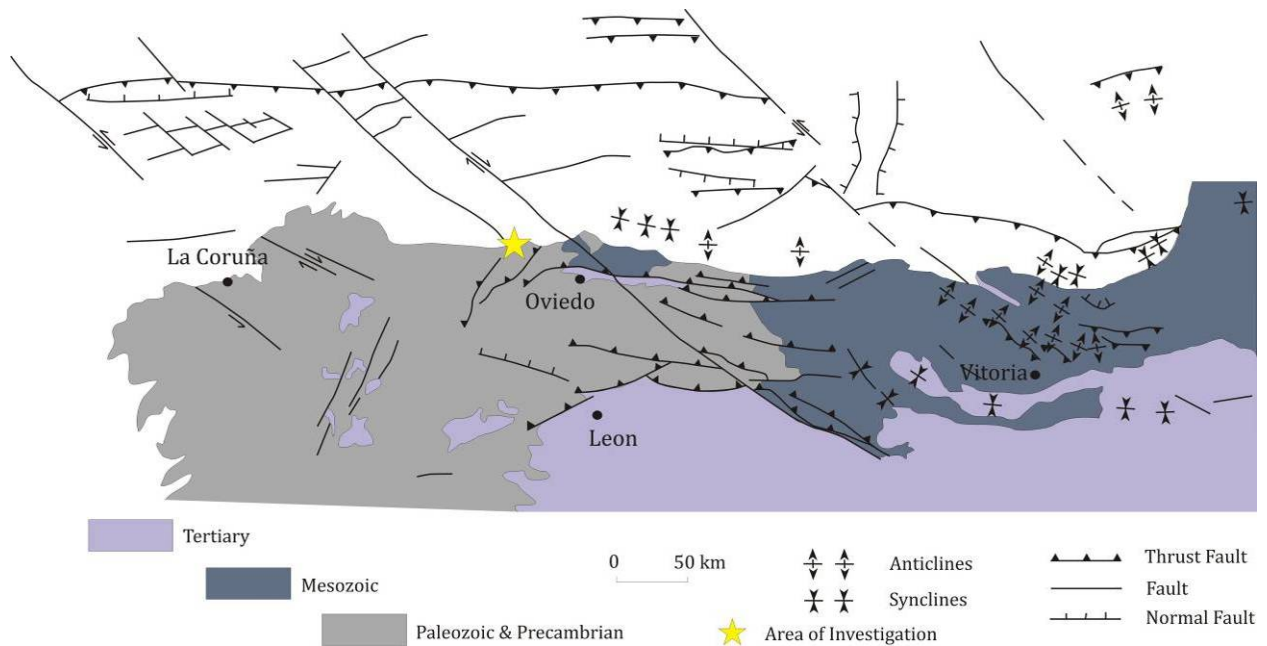


Figure 14: Alpine structures in northern Spain (modified and redrawn after Farias & Marquínez, 1992).

4. THE PORTIZUELO ANTIFORME

4.1. THE INVESTIGATION SITE

All information given in the chapters above should provide an outline of the regional geology of the working area, because the over-regional facts are fundamental for the understanding of the local setting. In the following chapter the outcrop will be presented and geologically interpreted.

The Portizuelo Antiforme is situated in the west of the Portizuelo Beach where a small nameless race joins the sea. At low tide, it is possible to walk far out into the flat abrasive platform and the structures of the cliffs at the shore side can be observed. The antiforme is a very good example of the broad, upright, parallel folds, described as F3 structures in literature (see chapter 3.4). At its highest point the wall rises up to 30 meters over the beach shingle and strikes NE-SW, which is perpendicular to the fold axis. Its orientation makes the wall particularly suitable for structural studies. From a frontal view, the slight asymmetry of the fold attracts attention. The eastern limb dips relative steeply whereas the western limb is less inclined. The asymmetry results from a shift in the verticality of the axial plane which steeply dips (~75°) to the east. The core of the fold comprises of bright, yellow quartzites from the upper Cabos Series. The top of the succession is made up of dark, fine layered slates of the Luarca Formation. In-between the Luarca Slates and the Cabos Series lies the Transition Zone. Also visible at first sight are two faults that moved the stratigraphic layers relative to one another. More accurate investigations define them as sinistral strike slip faults with a thrusting component. The striation dips about 10° to NE. With this data and the measured vertical offset of 20 meters of the small grey layer in the upper part of the quartzites, a total offset of about 130 meters can be calculated for the main fault. Apart from the strike slip faults a conjugated set of normal faults crops out in the eastern part of the wall.

Detailed consideration of the fold-core reveals the intense cataclastic deformation that the quartzitic rocks suffered from. The slaty layers of both the Transition Zone and the Luarca Formation show significantly less brittle deformation. Loose quartzitic material is constantly dropping out of the wall, forming small piles of debris spreaded along the wall foot by the strong tidal activities. Crushing is most intense in the areas adjoining the main faults and ceases rapidly with distance from them. The terms 'core zone' and 'damage zone' are further on used to describe the heavily crushed areas directly next to the faults (core zone), respectively the more distant areas that also suffered certain cataclastic deformation (damage zone).

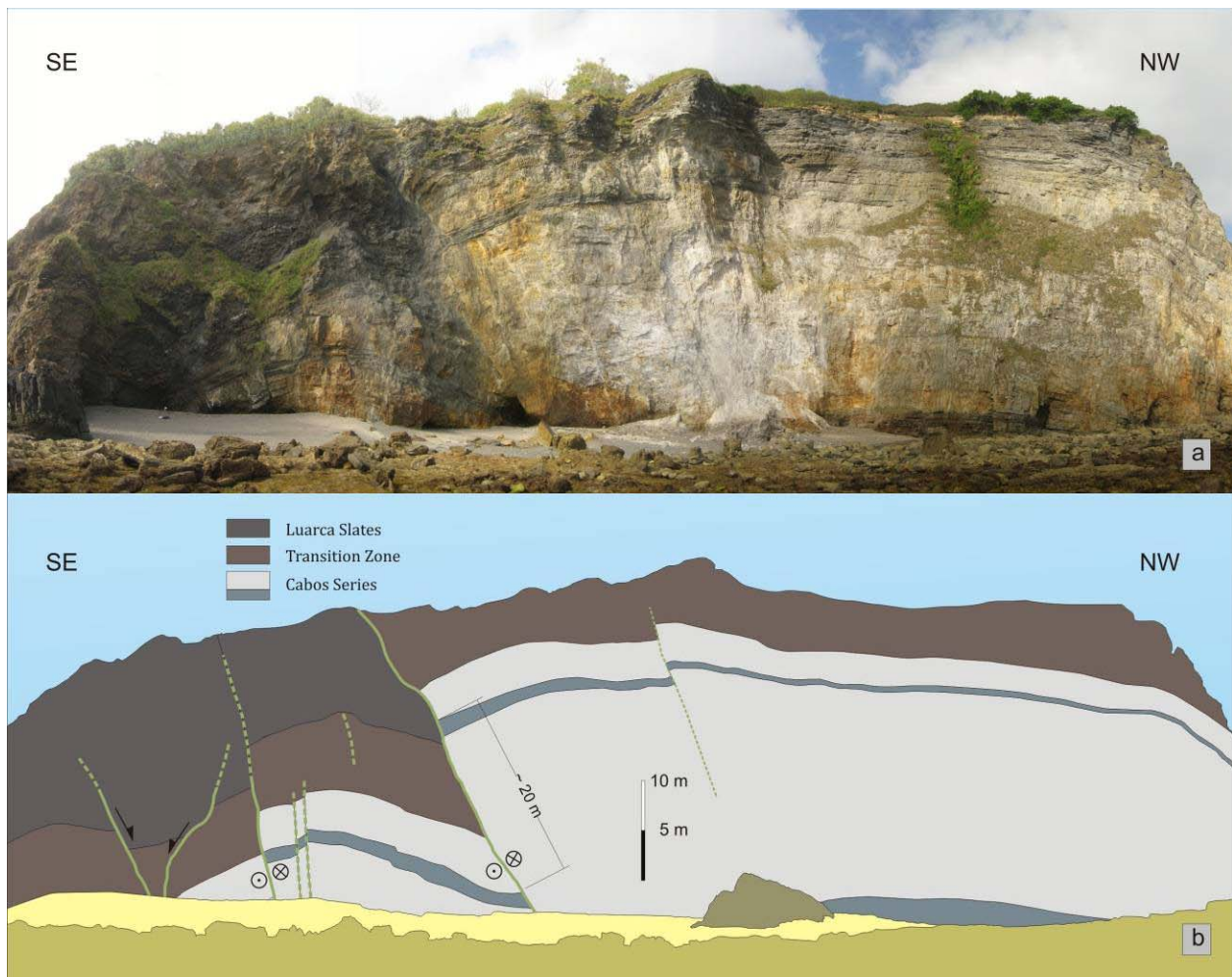


Figure 15: Portizuelo Antiforme in frontal view. a) Photograph, b) Geological interpretation.

4.2. STRATIGRAPHIC SUCCESSION AND SEDIMENTARY STRUCTURES

The gradual change from siliciclastic to marine sedimentation without any carbonatic input has a most striking appearance in the outcrop. A stratigraphic column of the encountered lithologies is presented in Figure 16.

The siliciclastic sediments are represented by the very hard, brittle and bright quartzites of the upper Cabos Series. Fäber and Jaritz (1964) call these units the Barayo Quartzite. In the outcrop a thin band of fine-laminated silty sandstone separates the Barayo Quartzite in two parts. The lower part is about 25 meters thick. The rocks are of a bright yellowish to grey colour and show no sedimentary structures. The about 4 meters thick upper part is almost similar but of slightly darker colour. The quartzites are very pure, this means that they mainly constitute of quartz and hardly comprise other minerals. But also little variations in mineralogical composition

are macroscopically visible through different colours. Only microscopic investigations supply detailed information on this (see chapter 6.3).

Below the massive quartzites a series of quartzites, sandstones and siltstones is located whose layers show thicknesses of only a few decimetres. In between, layers of fine clayey materials with a certain content of graphite show up. In the sandstones and in some of the quartzites a cross-stratification can still be noticed, which encloses an angle of about 15° with the stratification.

The Transition Zone starts just above the Barayo Quartzite. Its thickness cannot be distinguished exactly, because the change to the non-clastic sediments is gradual. However, in this case the thickness is set to about 10 meters. Above these 10 meters, the sediments show clearly less siliciclastic intercalations, and are therefore supposed to belong to the Luarca Formation. Rocks of the Transition Zone are sandstones and siltstones. Sometimes even finer material is deposited in laminated layers. The sandy and silty layers occasionally show sedimentary structures, like cross-stratification and ripples. The fine layers are rich in mica and iron, whose oxidized state gives them a reddish colour.

The Luarca Slates on top of the stratigraphic succession are defined by a high content in fine marine material that contain a certain amount of iron and organic matter but lack any carbonate (Marcos, 1973). In the outcrop the dark, slaty layers alternate with brighter, coarser material such as siltstones, which might be still of terrigenous-clastic origin. The fine laminated pelitic layers (i.e. the slates) present cleavage planes covered with mica.

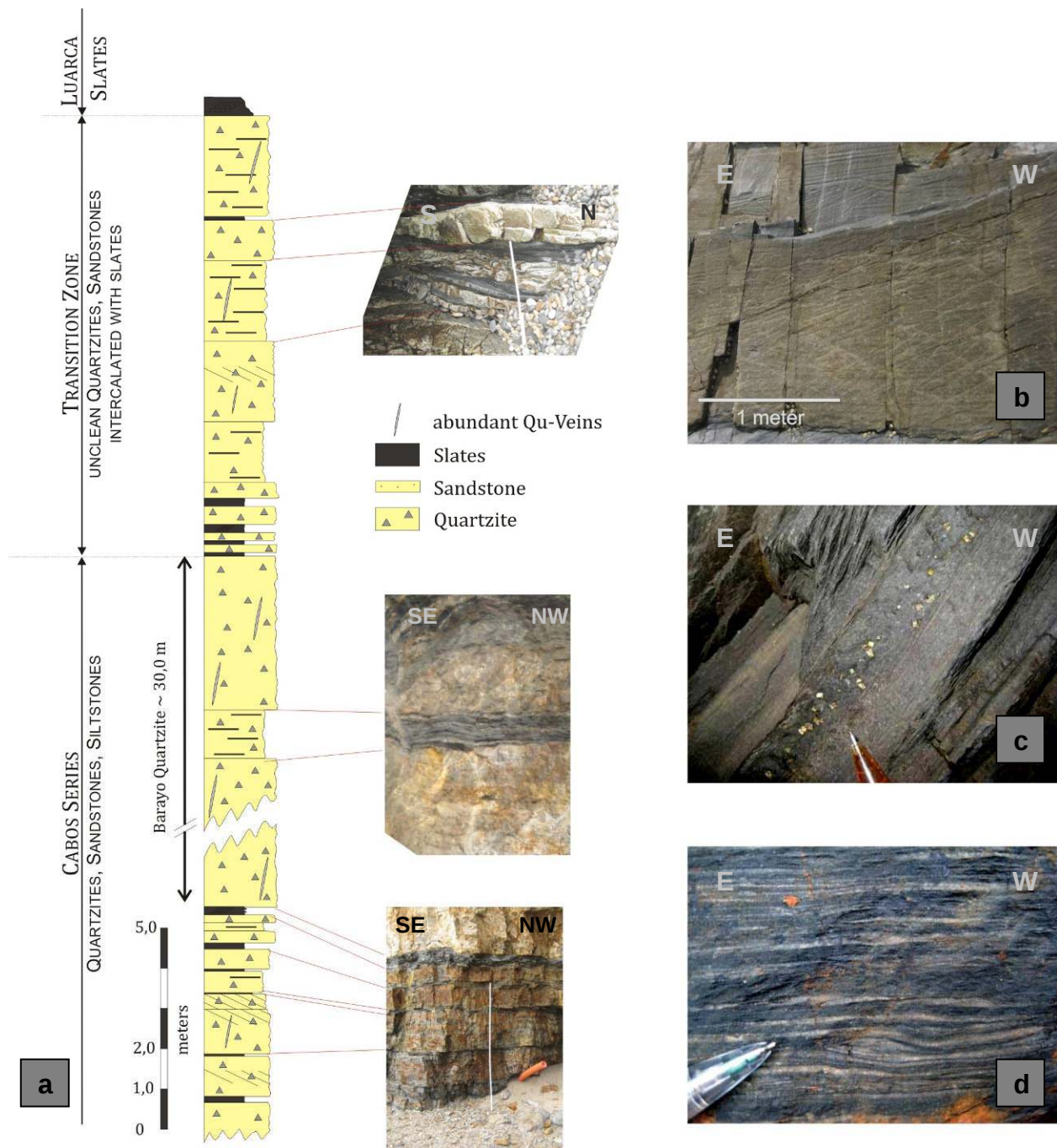


Figure 16: Stratigraphic and sedimentary characteristics of the rocks: a)-Stratigraphic succession in the investigation area with photos of interesting regions. b) Hummocky cross stratification in sandstones of the Cabos Series. c) Pyrit crystals in the Luarca Slates. d) Lamination of silty and clayey material in the Luarca Formation.

Idiomorphous pyrit crystals grow stratiform in some layers. A certain amount of graphit, as an alteration product of organic matter, is supposed to be incorporated in the clayey sequences but could not be explicitly proved by x-ray powder diffractometry. The peak of graphite in the spec-

tral profiles of the x-ray powder diffractometry is very close to the quartz and muscovite peak and cannot be separated from them.

Concerning the rheology of the rocks, it can be generalized that the sediments get more 'soft' in the upper regions because of the increasing content in silt and clay. On the contrary, the Barayo Quartzite is very rigid and brittle because it lacks any fine material.

5. STRUCTURES OBSERVED IN THE FIELD

5.1. STRUCTURES RELATED TO HERCYNIAN DEFORMATION

The observed structures in the course of my work are sorted in accordance to the generally accepted deformation phases introduced by other authors (Bastida & Pulgar, 1978; Marcos, 1973). According to chapter 3.4, the terms F1, F2 and F3 are used. Only two of the described hercynian deformation phases are recorded in the working area. The F1 recumbent folding is preserved as a penetrative, rough to slaty cleavage or primary schistosity as it is developed over a primary sedimentary texture (Bastida, 1982). Mostly it is found in the black slates of the Luarca Formation but it can be occasionally found in fine-grained layers of the Cabos Series too. Frequent S1 cleavage penetrates the rocks in the more inclined parts of the limbs in the east and west of the anticline. Right there also isoclinal folded layers of sandstones and slates with NE-SW fold axis can be regarded as minor F1 folds. Layer parallel pressure solution horizons in the quartzites of the Cabos Series is also related to the F1-shortening event (see Fig 17a) and developed together with overburden stress induced compaction. In the Luarca Formation, especially eastward of the antiform, evidence for pressure solution in horizons parallel to layering can be found frequently. The peak of pressure solution deformation is supposed to be accompanied by phase 1, but may already have started before.

The S1 cleavage is sub parallel to the layering and is determined by the shape preferred orientation of micas, derived by mechanical rotation of detrital grains perpendicular to the pressure direction (Marcos, 1973). Stress induced mineral growth also occurs but is only of minor importance. Nevertheless the major mechanism for generation of S1 cleavage is pressure solution. Minerals, such as quartz and feldspars, are dissolved in a field of high differential stress and precipitate in low differential stress sites, like gashes and strain shadows.



Figure 17: F1 related structures. a) Notable decrease of volume due to pressure solution in silty layers of the Luarca Formation. b) Layering parallel stylolites in quartzites of the Cabos Series. c) Layering-parallel slaty cleavage, note: F1 folded Qtz-Veins. d) Isoclinal folded layer of siltstone. The darker bended lines are originally cross stratification horizons.

The most remarkable structure in the working area is the big Antiforme of the Portizuelo beach, which is clearly related to the deformation phase F3. The fold is slightly asymmetric with a steeper limb in the west, and therefore possesses a non-vertical axial plane dipping about 75° to the east. The stratigraphic layers persist parallel and isogons remain perpendicular to the stratigraphic boundaries. Therefore the folds are classified as 1B parallel folds (according to Ramsay & Huber, 1987). Secondary F3 folds only appear in the eastern limb in the Luarca Slates. As already mentioned in Chapter 3.4, F3 and F1 folds generate an interference pattern of type 3 refolded structures (Ramsay & Huber, 1987) which cannot be seen in the scale of an outcrop but only in overregional geological sections (e.g. see Fig 11 in chapter 3.4). The foliation S3 generated by F3 is somewhat similar to S1 but not as penetrative and rougher, with greater distances between cleavage domains. Generally, S3 cleavage is scarce in the outcrop and can be found only in the slates of the Luarca Formation. In this case a certain angularity with the S1 cleavage reveals its different character and, together with S1, crenulation cleavage is produced. The cleavage planes are parallel to the axial planes. In areas with more inclined

limbs, where the angle between S1 and S3 gets small, no new cleavage planes are formed, but S1 cleavage planes are reactivated. The deformation of the layers itself depends clearly on the composition and thickness of the layers. In the Luarca Slates siltstones, slates and quartzites alternate in small steps. Thinner layers show abundant secondary folding and strain is concentrated between the layer boundaries, giving rise to flexural folds. Movement striae on slaty surfaces beyond more competent layers indicate the flexural slip along these boundaries. As another consequence of the flexural folding, quartz-filled extension gashes appear in some heavily deformed layers in the Luarca Formation. They are sometimes further deformed, rotated and cross-cut by younger gashes. When the layers reach a certain thickness, especially the massif banks of quartzite in the Cabos Series, strain distribution within the layers leads to tensile cracks in the hinge domain of the folds. These cracks usually come up with secondary folds of minor wavelengths. Other F3 related structures are thrust faults that emerge from stratigraphic boundaries and then criss-cross the layers, leading to a flat-ramp geometry of the fault-plane. One of these faults crops out on the adjacent beach in the west of the antiforme. The total offset along the stratigraphic boundaries is about 3 meters, diminishing to zero at the tip of the fault. The layers are bended in the course of the deformation (i.e. Fault Propagation Fold). Hercynian quartz veins are frequent, but their separation in different generation seems difficult. A set of veins that is folded by the F1 and the F3 event may even be pre-orogenic, whereas only slightly folded veins are syn- or post-F3. In this case only their metamorphic aspect infers hercynian ages.

In respect of metamorphism, a greenschist facies for the area around Luarca is reported (Suárez et al. 1990). The quartzites of the Cabos Series are very bright and are well cemented. Veins in the quartzites, related to the F1 and F3 phase, are of a dull whitish color and their borders are not distinct but diffuse. The slates, siltstones and clays show abundant micas (phengite and muscovite) in their cleavage planes, presuming a low temperature metamorphism. Better evidence for metamorphic conditions was derived from deformation mechanisms identified in thin sections (see chapter 6.2 & 6.3).



Figure 18: F3 related structures: a) Flexural slip lineation between two layers. b) Quartz rods in slaty layer affiliated to flexural slip. c + d) F3 secondary folds which refold F1 isoclinal folds.

5.1.1. GEODYNAMIC SETTING

The deformation phases F1 and F3 both have the same tectonic background. The shortening direction is NE-SW, indicated by the fold axis of F1 and F3 folds. Tectonic vergence is towards west in both cases. F3 axial planes, measured directly from the folds and from the S3 axial plane cleavage dip steeply towards east. F1 axial planes are refolded by F3 and usually enclose a small angle with the stratification. Stretching lineations in pelitic layers have identical orientations as the F3 fold axis. Crenulation cleavage (S1 folded by S3) is rare and only occurs in the hinge area of some minor F3 folds. The produced crenulation lineation is once more parallel to the F3 fold axis. A dominant set of faintly deformed quartz veins can be posted as late Hercynian. They opened in a late stage of the variscan constriction and run perpendicular to the F3 axial planes. In a later deformation phase (see next chapter) some of these veins are reactivated as sinistral strike slip faults.

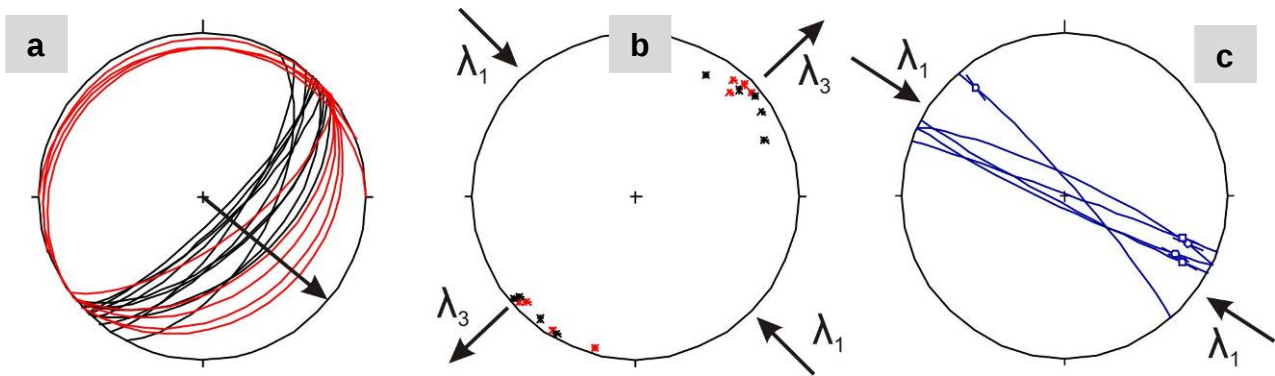


Figure 19: Stereoplots of hercynian structures orientated to the north. a) Axial surfaces of F1 folds (red) and F3 folds (black). Arrow indicates tectonic transport direction towards SE. b) Fold axis of F1 and F3 folds. Extension direction parallel to axis, shortening is perpendicular. c) Late hercynian Qtz-veins, indicating ~NW-SE shortening. Some planes are reactivated as slip planes in a later phase.

5.2. LATE HERCYNIAN BRITTLE DEFORMATION

The most remarkable, late hercynian, brittle structures are of course the strike slip faults and the cataclastic zones. Besides these, also a conjugated set of normal faults, quartz veins and a widespread joint system exists.

After the variscan constriction ceased, a phase of extension began (Quesada, 1992). Early structures in this phase are **en echelon quartz veins** that exhibit the extensional regime towards E-W. These veins crosscut the older hercynian structures clearly and are less affected by metamorphism, indicated by their clear colour and sharp boundaries. Extensional characteristics can also be interpreted through a **conjugated set of normal faults**, cropping out in the southeastern part of the wall. There, a layer-offset of about 2 meters can be noted and cataclastic material shows up in the direct contact with the slip plane. The eastern fault surface dips 50° to WSW and has a straight form. The western fault dips 40° to SE and has a sinuous form. Thus a ramp flat geometry results from the small angle between fault and stratification. In the core zone of this fault, cataclastic deformation is frequent and furthermore the strong internal deformation leads to chaotic microfolding of certain layers.

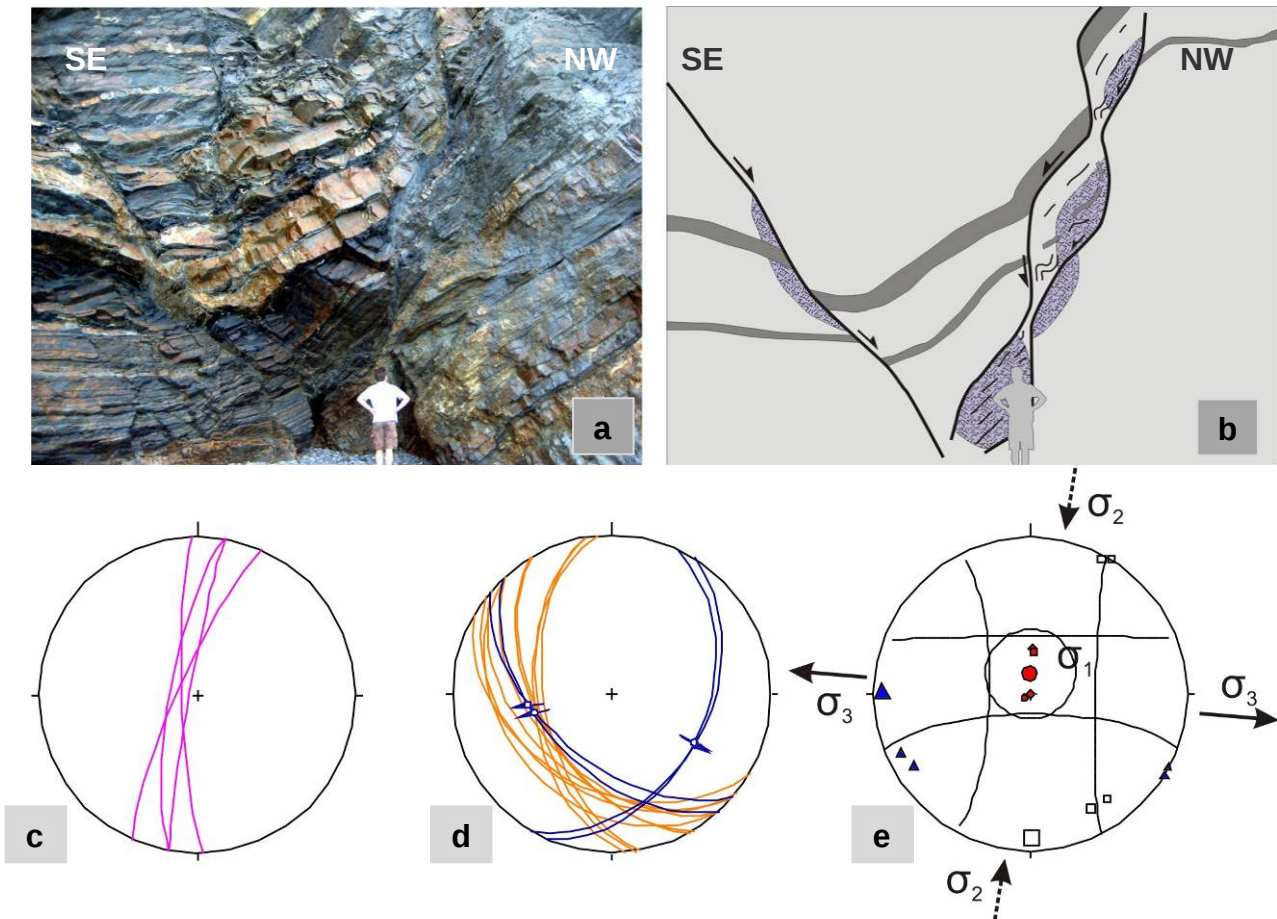


Figure 20: Structures and stereoplots of an extensional regime. a) Conjugated fault system. b) Interpretation; western faultplane has a wavy shape; Blue areas are zones with enforced cataclasis. c) Stereoplot of the enechelon Qtz-veins. d) Normal faults. e) Fault plane analysis results of the PT-Axis Method (Turner, 1953) showing the kinematics of this deformation phase.

5.3. ALPINE BRITTLE STRUCTURES

Two main transform faults crop out in the antiformal, displacing the layers of the Cabos Series, the Transition Zone and the Luarca Formation. The lineation on the fault planes dips slightly ($\sim 10^\circ$) to NE. A total sinistral offset of 120 meters in case of the western fault (FII) has already been stated in chapter 4.1. Shear sense indicators are difficult to find, because of the highly altered surfaces. Best evidence for sinistral movement is the offset of a marker horizon in the Cabos Series, which is relatively lifted in the W of the largest fault (FII). The smaller FI fault, more in the east, has the same movement striations and also the same shear sense. More evidences for shear sense come from sheared quartzitic blocks. There, quartzite-chips are spalled off in the direction of the movement (see fig. 21). The core zones of the faults are made up of several cm thick fault gauge. X-ray powder diffraction measurements reveal a mineralogical composition of quartz and muscovite and traces of kaolinit and gypsum. The surrounding rocks of the Cabos Series have similar composition; Kaolinit and gypsum are alteration products of

feldspar minerals from the Transition Zone or the upper lying Luarca Slates. That already indicates to a certain movement and the distribution of the fault material. Adjacent to the fault gauge material the rock is heavily cracked and cataclastically deformed. The hanging wall suffers stronger deformation than the footwall rocks, where the cataclastic zone already ends after several centimeters. The width of the damage zone, meaning the area next to the core zone where the rock is still mechanically damaged by the fault, reaches several meters in the hanging walls (up to 50 meters of the FII fault). The damage zone of the hanging wall from the FI fault even intersects with the damage zone of the footwall of the FII fault. To study the quantity of deformation in the rock, a Fracture Density Index (FDI) profile was made along the wall. The FDI is simply the number of fractures/joints that cross through a certain thickness of the rock mass per unit of length. The result shows the rapid increase in cracks, when approaching the main faults and also the abrupt decrease when reaching 'softer' lithologies (see fig. 24). This means that the brittle, hard quartzite of the Cabos Series is severely damaged by the faults, because no internal plastic deformation is possible. In contrast to that, in the underlying silty sediments of the Cabos Series and the upper-lying Transition zone, deformation can be accommodated by the fine grained layers by dissolution precipitation creep without development of distinct failure planes.

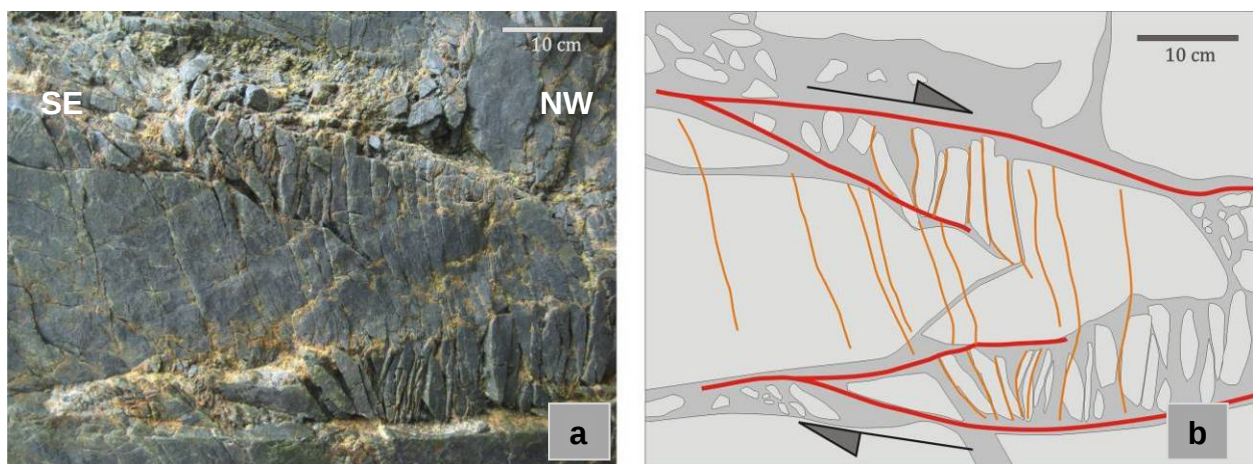


Figure 21: Damage zone near fault FI. Chips are breaking off from the quartzite, indicating a shearing deformation and also show the shear direction (here dextral). Picture taken from bottom to top; therefore the real shear sense is sinistral.

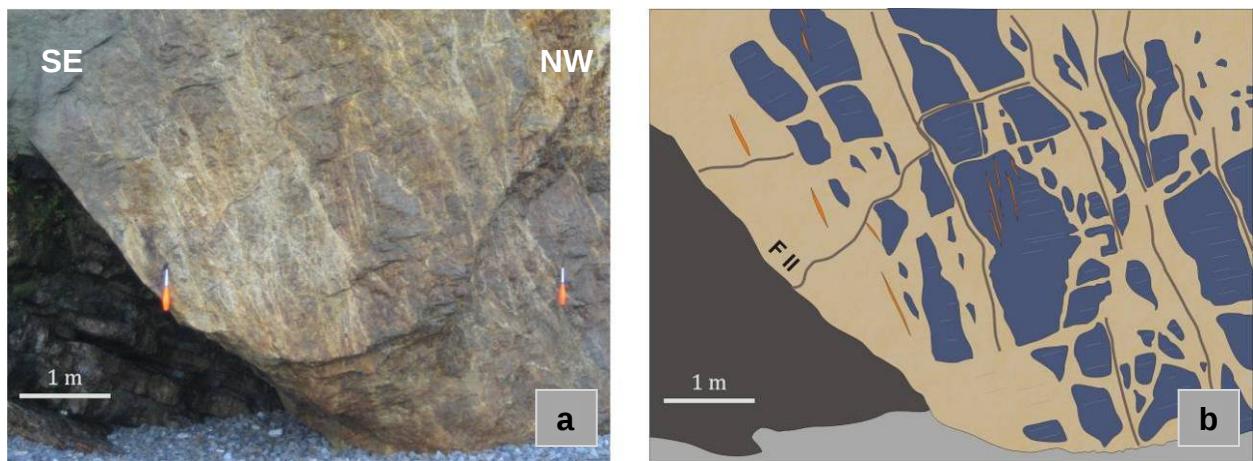


Figure 22: Core zone next to fault FII. Some areas are nearly undeformed, whereas the main part is severely damaged by cataclastic deformation. Several cataclastic veins cut through the rock mass, nearly parallel to the fault plane.

As already mentioned in a previous paragraph, there exist some vein planes, whose orientation favors a reactivation as lateral slip planes through this deformation event. Even conjugated fault planes on these veins were found which indicate dextral movement. Inside the core zone and damage zone several bands of cataclastic veins are visible. They penetrate, sub-parallel to the strike slip faults, the already cataclastically deformed rocks and can be even traced in the abrasional platform. These bands will be described in the following.

Entering the beach from the east, crests of more competent material that run out to the sea immediately attract attention. Upon closer examination the crests turn out to be made of cemented cataclastic material, mostly quartzite, and crop out as several dm-wide bands. These bands are more resistant against tidal erosion than the heavily jointed and fractured quartzitic host rock and therefore stick out morphologically. At low tide several dozens of these bands can be observed, cropping out from the abrasional platform. They can be tracked up to 50 meters out in the sea. By the aid of differential GPS, the cataclastic crests, the lithological contacts and the wall shape were mapped (see fig. 23). According to the spatial orientation, three different sets of cataclastic bands can be distinguished:

- The main set strikes more or less perpendicular to the wall and contains the bands with the greatest thickness (up to 1 meter). Here it is denominated as **D-structures** (according to their dike-like appearance). Cracked quartzitic material of both rounded and angular particles is cemented by a fine grained matrix. The quartzitic matrix has multiple colors, from white, yellow to brown-reddish, depending on the state of alteration by iron-hydroxide bearing fluids. In the center of these D-cataclasites, frequently a thin line of very fine grained, grey material shows up. Microscopic examinations reveal them as ultracataclastic material with grain sizes $<10\ \mu\text{m}$. Macroscopically, there does not seem to be a particular abundance of grain size reduction with distance from this 'core zones'

and generally no evidence for slip can be found. Therefore lateral movement cannot be proved. On the other hand slip planes with movement striations sometimes exist inside cataclastic D-structures, where ultrafine material is produced 'in-situ' as a process of comminution. In the other case parts of the fine cataclastic material must be transported to its actual position by the aid of fluids. This mechanism is terminated 'fluidization'. Better evidence for fluidization is attained by microscopic investigations (see chapter 6.4) and Grain Size Distribution (GSD, see chapter 7).

- A second system shows some kind of Riedel geometry in respect with latter mentioned D-structures and therefore these structures are called **R-structures**. The thickness of these crests is 20 to 50 cm, their length reached up to 20 meters. The constituting material is similar to those of the D-ridges. Slip planes are not found; the Riedel like geometry, however, proposes some lateral movement.
- The third set comprises of cataclastic veins which are joining the other sets to an overall network-like structure. The spatial orientation of this set varies over a wide range, but mostly the veins run diagonally to perpendicular to the latter mentioned structures, therefore they are called **Q-structures** in this work ("Quer" as German term for crosswise). The thickness varies between some centimeters to zones of about 1 meter width. The network-like characteristic is also visible in the structure itself. Fine bands of white quartzitic material cut through yellowish course 'protocataclastic' quartzite. Absolutely no slip evidence can be found.

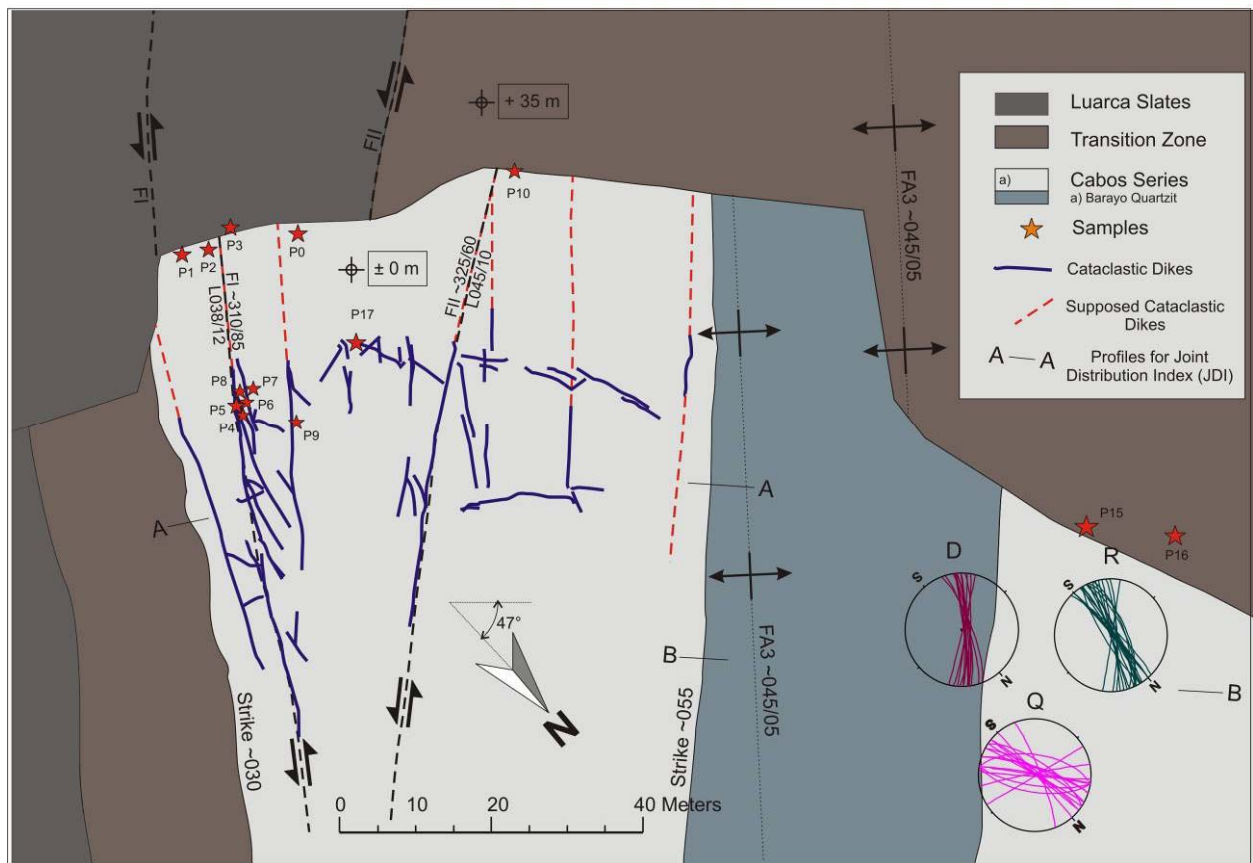


Figure 23: Geological map of the investigation area. Three distinct sets of cataclastic bands cropping out in the abrasional platform exist, but only occur in the hard Barayo Quartzite of the Cabos Series. Lithological contacts converge towards NE, because of the small plunge of the fold axis.

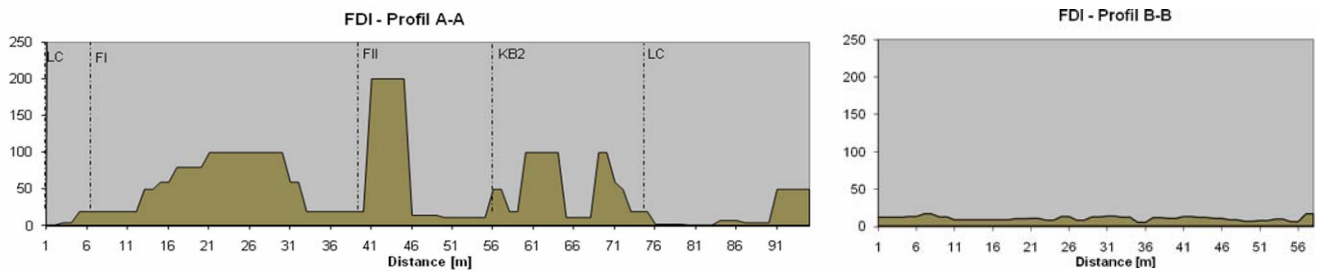


Figure 24: Fracture Density Index (Fractures/meter) for sections marked in Fig.23. Profil A-A: Damage Zone. LC...Lithological Contact; FI, FII and KB2 are faults. Peak just next to FII is the core zone of this fault, broad peak between FI and FII is because of interaction of both damage zones. Other peaks are smaller local faults. Profile B-B: no fault induced fracturing.

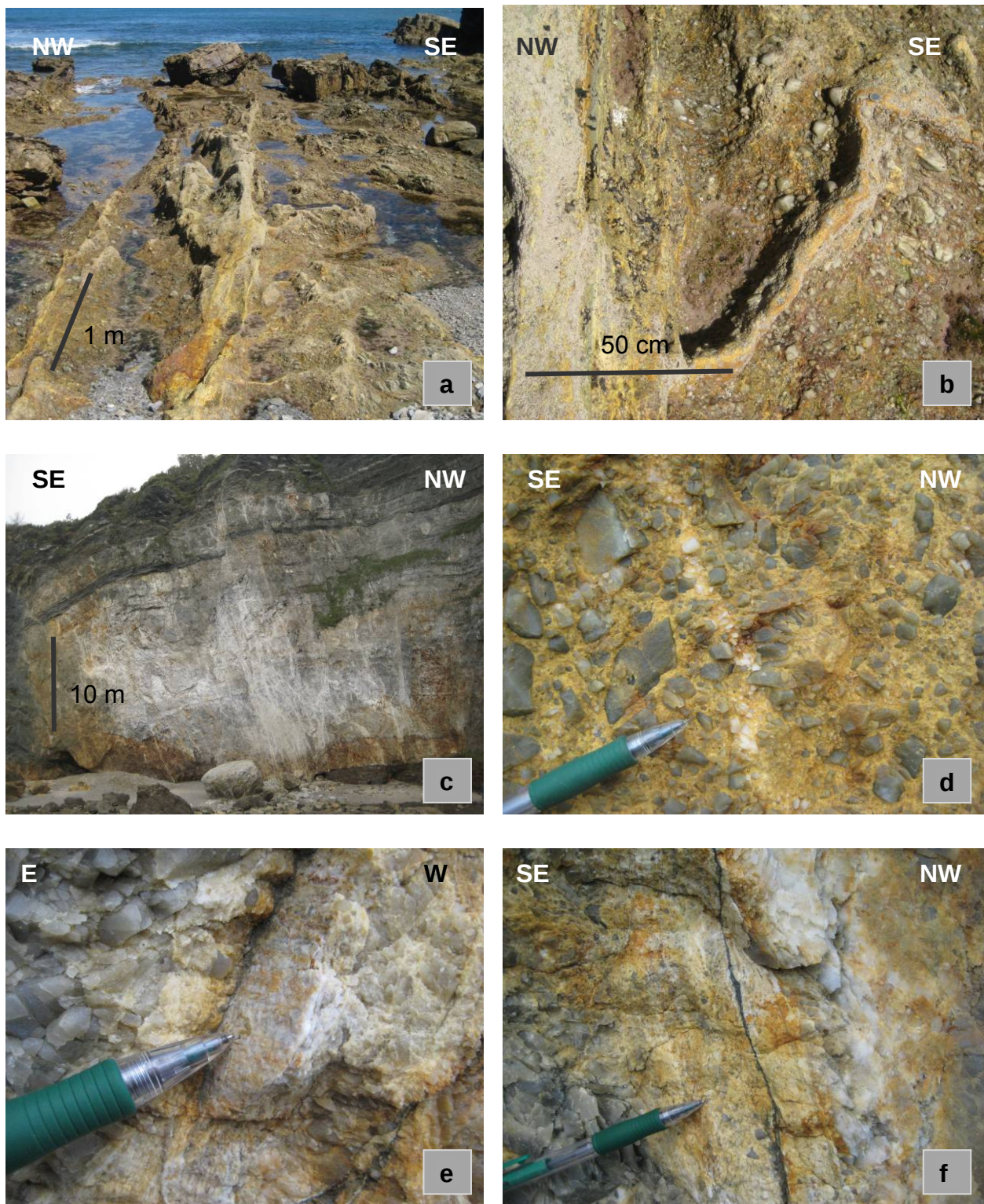


Figure 25: Structures related to the alpine brittle deformation. a) Crest of cemented cataclasites in the abrasional platform (D-structure). b) Branch of cataclastic vein that intrudes in the already fractures quartzite. c) Damage Zone west of FII. White lines are cataclastic bands in the quartzites which usually end at the contacts with surrounding lithologies (darker layers). d) Damage zone, close up. The quartzites and the veins are cracked but still in their original position. The yellow matrix consists mainly of cataclastic quartz, affected by fluidization. e) Slip planes in cataclastic bands, having the same orientation like the strike slip faults. f) 'Core zone' of a cataclastic band, consisting of ultracataclastic quartz.

5.3.1. GEODYNAMIC SETTING

As already described in chapter 3.6, the alpine orogeny comes along with an early N-S shortening regime in northern Iberia (Quesada, 1992). The Pyrenees and its western extension, the Cantabrian Cordillera, are elevated by the means of thrust faults and its crustal blocks are further on segmented by a system of strike slip faults (Farias and Marquínez, 1992). NE-SW striking left-lateral faults exist in the variscan basement westward of the Duero Basin (Vilariça Fault or Regua Fault). They are classified as structures emerged from a cenozoic N-S paleostress field (Anton et al., 2010). During this alpine phase, the main part of deformation was applied, induced by collision along the Cantabrian-Pyrenean border (Anton et al., 2010). The strike slip faults in the Portizuelo Antiforme have similar orientations and shear sense. Fault plane analysis of slip planes from the faults together with slip planes from cataclastic veins, indicate N-S shortening too. The fault system as well as the cataclastic veins may therefore be classified as Cenozoic-alpine structures.

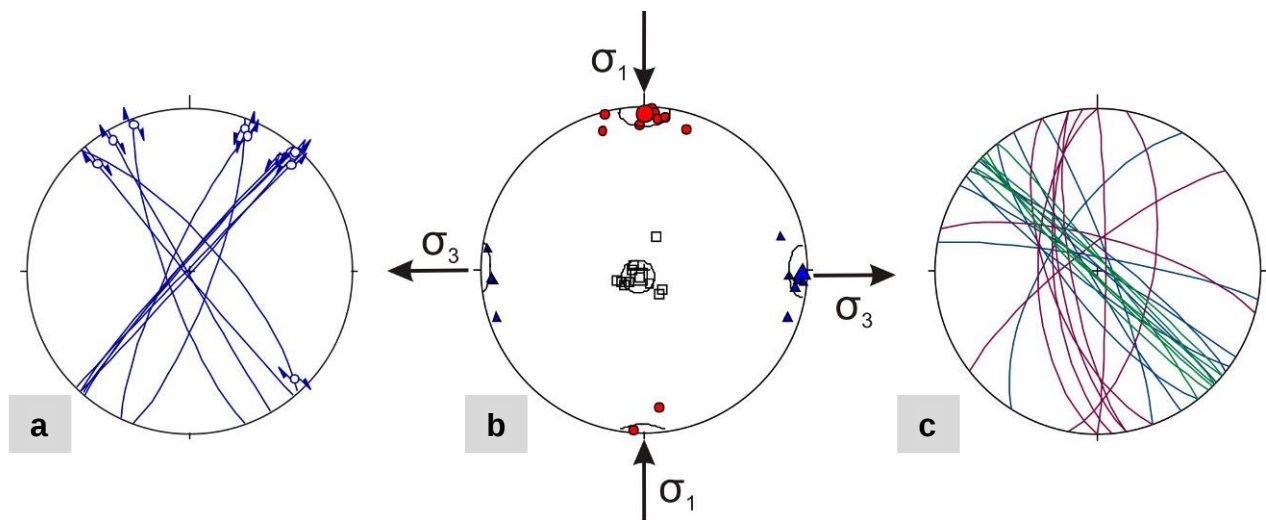


Figure 26: Stereoplots and fault plane analysis of the last deformation phase; A) Slip planes of the sinistral faults (NE-SW) and some conjugated dextral cataclastic veins (NW-SE); b) N-S paleostress field calculated with the PT-axis method (Turner, 1953); c) Alpine joint system, Main set strikes NW-SE.

5.4. SUMMARY OF THE STRUCTURAL DATA

According to the existing structures a clear division in different deformation events can be made. 4 deformation events can be distinguished and are listed, together with their related structures, in the following table.

Def. Event	related Structures	Paleostress/strain field	General Deformation Phase
D I	- Recumbent folds (isoclinal minor folds) - Slaty cleavage S1 - Pressure solution - Quartz Veins	NE-SW shortening	F1 Hercynian Phase
D II	- Open parallel folds - Hinge cracks - Flexural slip structures (Quartz Rods, Movement Strias) - Coarse schistosity S3 - Crenulation cleavage (S1 S3) - Quartz veins	E-W extension	F3 Hercynian Phase
D III	- en-echelon quartz veins - conjugated set of normal faults		Late Hercynian extensional Phase
D IV	- left-lateral strike slip faults with extended core and damage zones - Cataclastic veins in quartzites - Joint system	N-S compression	Cenozoic Alpine Phase

Figure 27: Table of structural data of the outcrop.

5.5. SYNOPTIC 3-DIMENSIONAL SKETCH OF THE OUTCROP

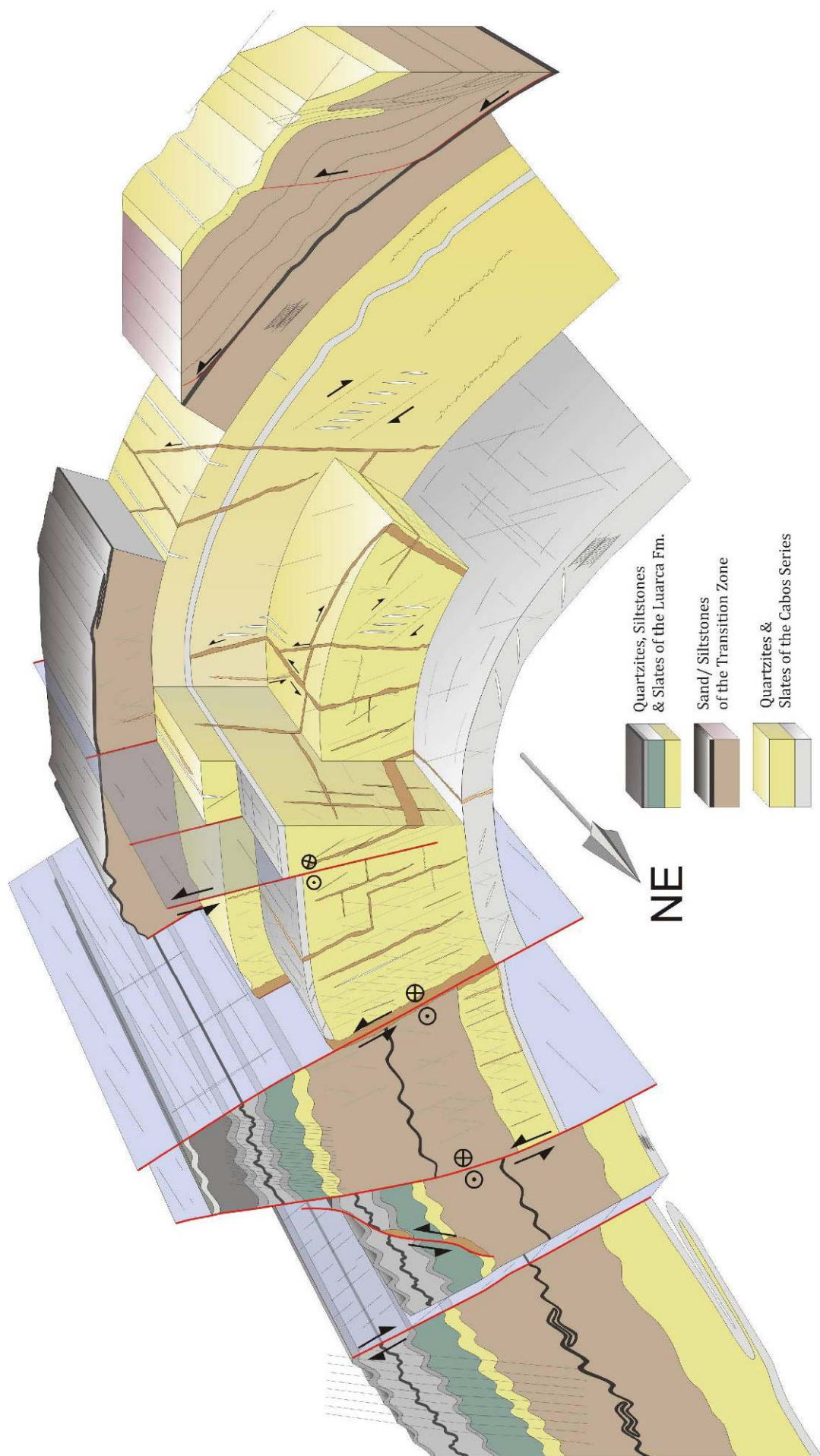


Figure 28: Synoptic 3D sketch

6. MICROSTRUCTURAL INVESTIGATIONS

17 orientated samples have been taken from different positions and different lithologies (see fig. 23 in chapter 5.3). The samples have been prepared and cut to small bricks at the University of Oviedo. The bricks have been sent to the University of Vienna, where 22 thin-sections (30 μm) have been prepared. All sections have been investigated by optical polarization microscope. Further on five ultra thin-sections (20 μm) were produced for Cathodoluminescence (CL) and Scanning Electron Microscope (SEM) studies.

6.1. METHODS

Optical microscope

The used device is a Leica® DM4500 P polarization microscope. Photographs were made with a Leica® DFD295 digital color camera with 3Mpixel, mounted behind the ocular. Pre-processing and automatic scaling of images as well as file management has been done by the Leica® integrated Microscope Assistant software. The microscope has a 10x ocular lense and objective lenses of 1.25x; 2.5x; 5.0x; 10.0x; 20.0x and 63x. Total magnifications of 12.5 to 630 times are achieved. The samples were prepared by the conventional way of gluing the bricks on glass plates with epoxy-resin. The brick is cut by special diamonded rock saws and then is consequently polished to a thickness of 30 μm . For further protection the sections are finally varnished.

Cathodoluminescence optical microscope

For the studies of the Cathodoluminescence a Lumic® HC5-LM Cathodoluminescence microscope was used. Additional devices are a power supply rack to provide the required high voltage for the electron gun and a vacuum pump system to generate a high quality vacuum in the specimen chamber. The acceleration voltage applied on the tungsten filament in the electron gun is 14 KV; the beam current can be varied from about 0.05 to 0.4 mA, depending on the examined material (e.g. Quartz needs much higher excitation energy than calcite to produce visible luminescence). Luminescence images are projected on a silica glass in the optical path of the microscope. With a digital colour camera pictures are taken and processed on a personal computer. Samples with thickness less than 25 μm are most useful for CL studies, so special ultra-thin sections had to be prepared. These sections were coated with graphite to make them conductive and to produce visible electron emittance.

Scanning electron microscope

An Inspect™ S50 Scanning electron microscope with a backscatter electron (BSE) detector was used. The device is supported with a high energy electron beam, operating with up to 30 KV. The specimen chamber has a movable mounting table. Vacuum is generated by done by two turbo vacuum pumps. Depending on the quality of the vacuum a resolution of ~ 4nm can be achieved. Control of the microscope is done by the integrated software.

6.2. SLATES OF THE TRANSITION ZONE

Two samples of slates in the Transition Zone were taken. The first one (P01/09) comes from the eastern limb of the antiforme. There, stratification dips about 70°, which is only slightly different to the dip of the F3-axial planes. Macroscopically identifiable foliation is the S1 primary cleavage due to F1 deformation. It is possibly reactivated by F3 folding and therefore superimposed by S3 axial plane cleavage. The rocks are finely laminated in the range of a few centimetres. Bright quartz-rich layers alternate with darker mica-rich layers. The grain size is generally below 1 mm, excluding the scattered, late formed pyrite minerals. The second sample (P14/09) has been taken from a more central position of the antiforme where layering is quite different to F3 axial planes. In this case cleavage is exclusively due to compaction and F1 deformation. No S3 cleavage has developed in this section. Considerable volume reduction during the compaction of the sediments is evident because of the existence of pressure solution cleavage planes parallel to layering.

Altogether three foliations are existing in the slates. A primary foliation S0' produced by compaction of the sediments and a possible early tectonic vertical compression. The second foliation event S1 can be described as a mostly spaced cleavage, evoked from the F1 folding. Occasionally a rough schistosity S3 shows up, which is due to F3 folding.

Microscopic examinations also define the S1 foliation as a spaced cleavage. The cleavage domains have a wavy shape and occupy approximately 20 % of the rock volume. Constituting minerals of the cleavage domains are muscovite, highly altered feldspar and iron ores. The microlithons have a thickness of about 100 µm and constitute of the original material, which is mostly quartz, flaky muscovite and some feldspar. In all cases the feldspar is strongly altered and converted to aggregates of fine sericite. Only the morphological shape of the aggregates points to the original existence of feldspar. Another fact that proposes feldspar as minor constituent is the result of the powder diffraction analysis of fault gauge material. There, traces of kaolinit can be measured, which is a decomposition product of feldspar. Some authors (e.g. Färber & Jaritz, 1964; Baldwin, 1975) state feldspar-containing sandstones in the lower layers of the Luarca Formation and consequently also in the Transition Zone.

The transition between microlithons and cleavage domains is discrete and the cleavage domains sometimes show an anastomosing shape. The amount of solved minerals is great, which is already indicated by the enrichment of the dark mineral phases in usually quartz-rich parental rock. Recrystallization of the solutions within the system does not occur, but the material is transported out of the system and precipitates in voids. Such spots are, for example, cracks in the tensile areas of bended layers or quartz rods, developed due to flexural slip between the layers.

The formation of the primary cleavage is always parallel to layering and is developed in all layers. In mica-rich layers, the foliation is often made up of the axial plane cleavage S1, which is subparallel to the stratification in most cases (see fig. 31) Further on, very fine-grained, clayey layers often record intense flexural slip deformation. In such cases SC-structures are developed where S is the cleavage produced by the generalized local stress field, and C is S0' respectively S1 from the compressional over-regional stress.

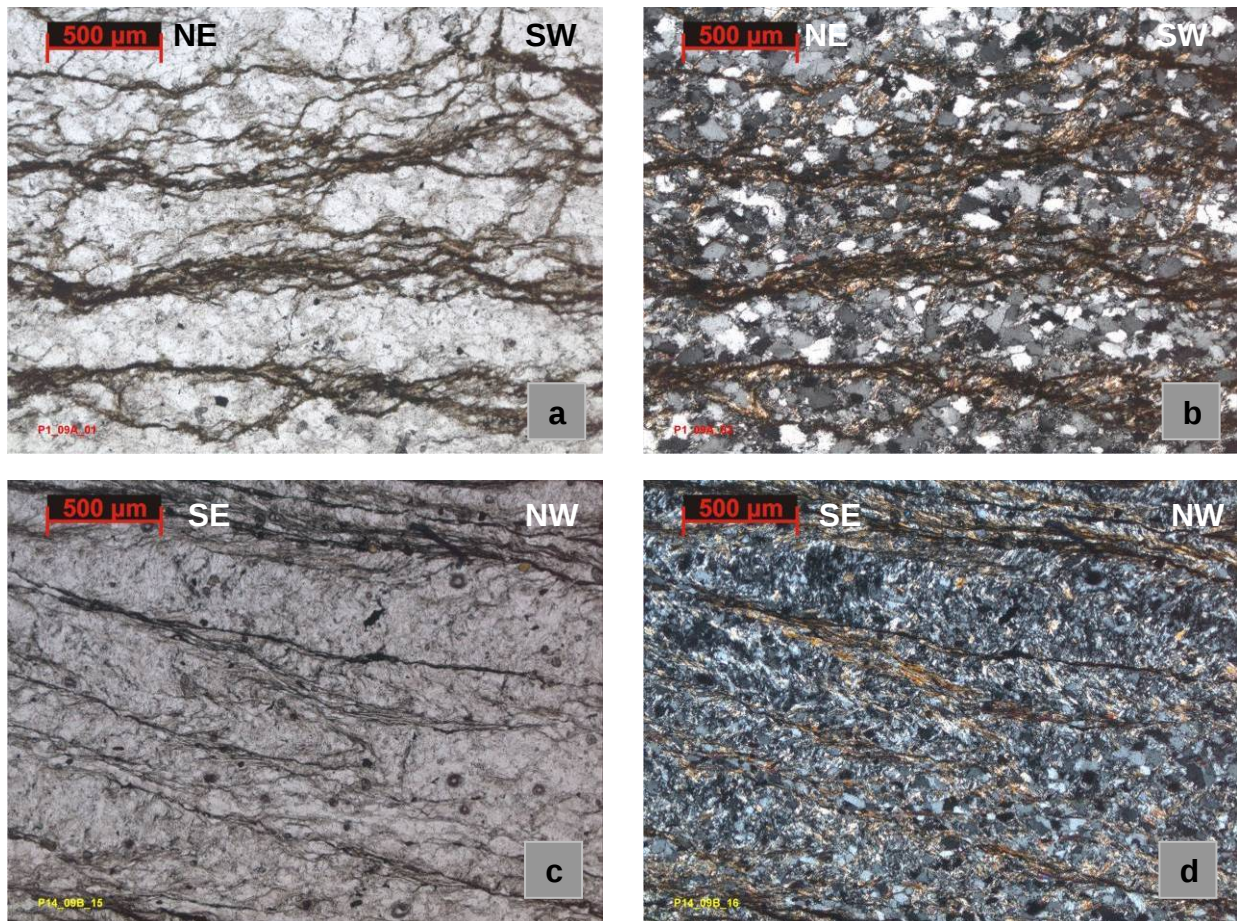


Figure 29: Microscopical aspects of the cleavages in the slates of the Transition Zone: a) Slaty cleavage S1, possibly overprinted by parallel S3 cleavage (P01/09); PL b) Microlithons comprise of quartz and sericite. The cleavage domains are made up of different micas and opaque phases. XPL c) Crenulation cleavage in P14/09; PL and XPL d) shows finer material than picture (b). Microlithons contain more dark minerals, as well as chlorite, which has some relics of an earlier foliation.

Muscovite appears as detrital platy grains, orientated in the cleavage direction. These muscovites are too coarse grained for the growth during the low grade metamorphism. They already existed before. Therefore, they are exclusively orientated in the stress field by mechanical rotation. Muscovite also appears as sericite, its fine grained flaky form. The sericite is finely dispersed in the rock mass and grows preferentially in the interfaces of quartz grains. They show no preferred orientation in the microlithons, and just overgrow the quartz grains. In some cases the sericite appears in clusters, forming a crystal like shape. This is thought to be a secondary formation of sericite from feldspar disaggregation. In sample P14/09 chlorite appears in the microlithons, showing an earlier foliation, where the F1 overprint results in a crenulation cleavage in this section (see Fig. 28 b & c).

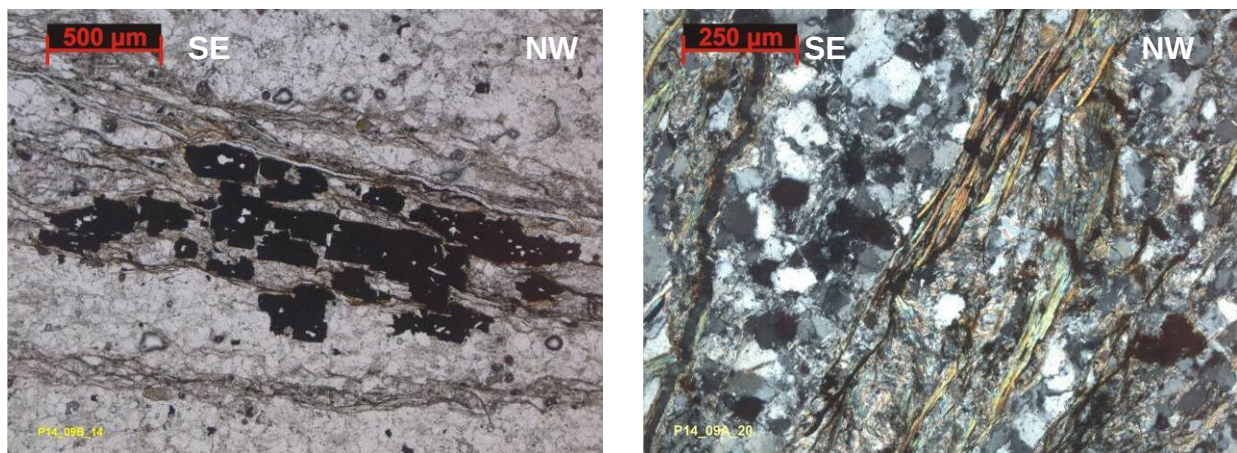


Figure 30: Mineralogical aspects of the slates (P14/09A): a) Pyrit crystals growing in the cleavage domains. PL b) Detrital micas of intense interference colors, orientated in the S1 cleavage planes and fine flaky sericite of washy colors. XPL.

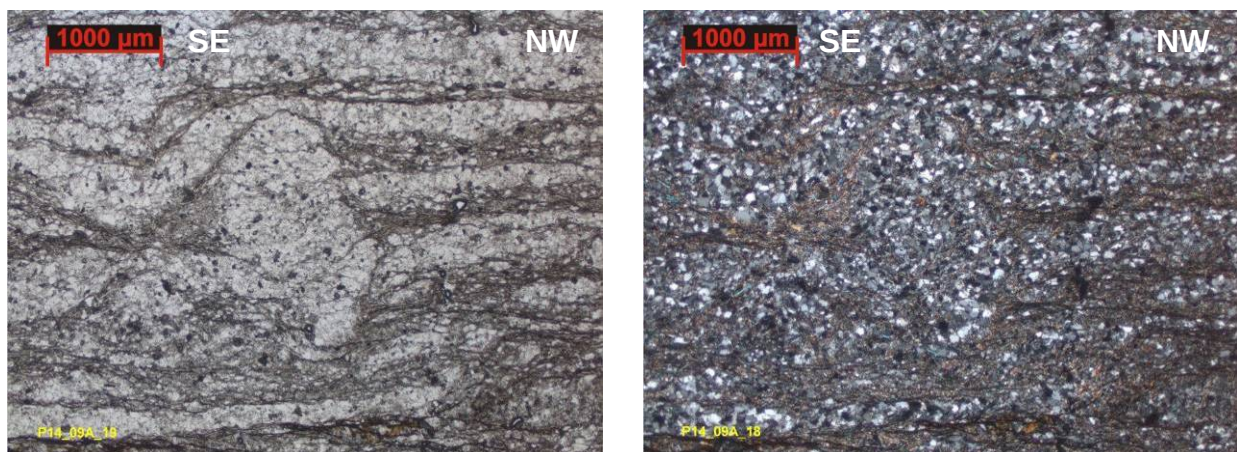


Figure 31: Pressure solution cleavage in P14/09A: S0' deflected by a quartzitic clast. a) PL b) XPL

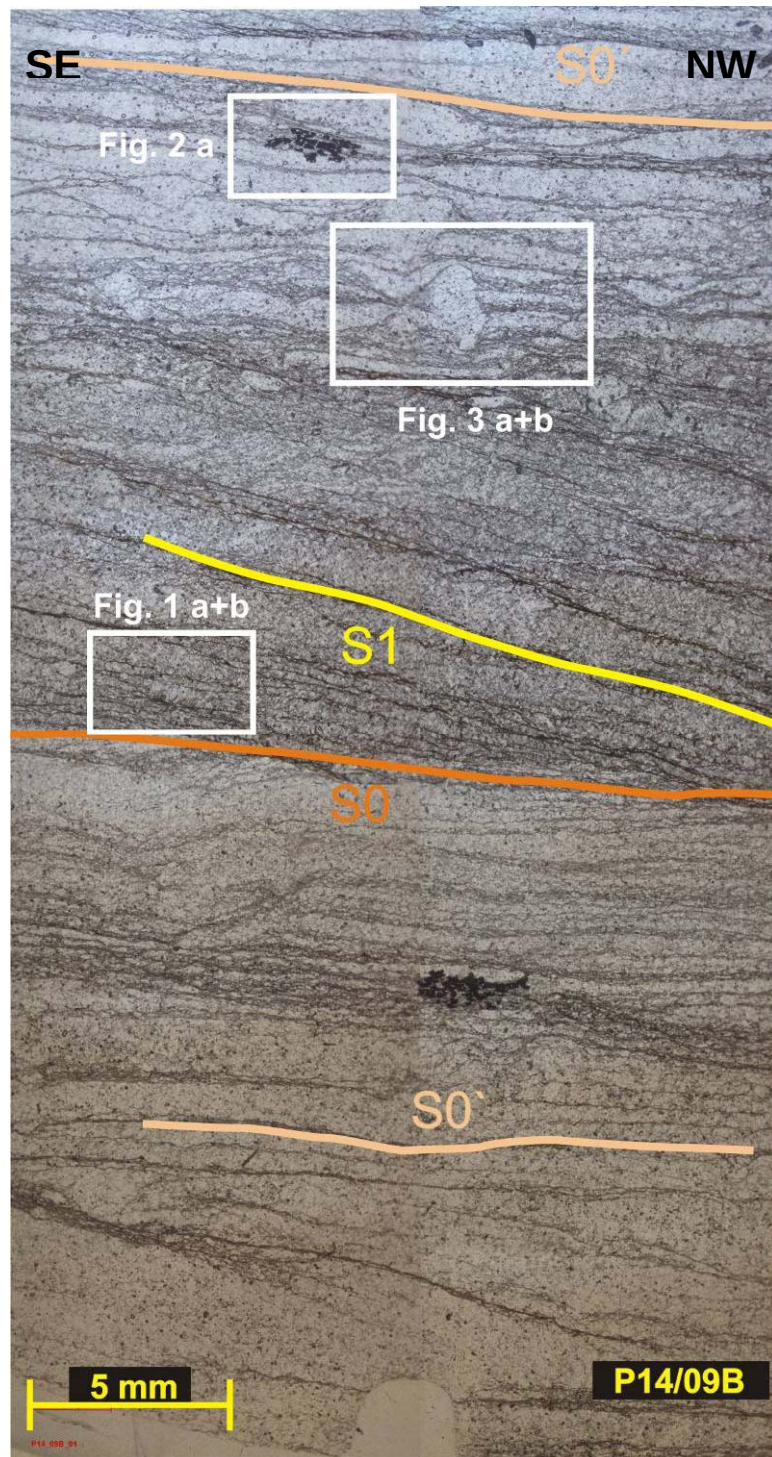


Figure 32: Stitched image of the complete thin section P14/09B. S0' is parallel to S0 (layering), whereas S1 has slightly different orientation. The axial plane cleavage S1 only appears in fine-grained incompetent layers and is not detectable in the quartz-rich competent layers.

6.3. QUARTZITES OF THE CABOS SERIES

Concerning the mineralogical composition, a gradual change from pure quartzite, made up of more than 99 vol% of quartz, to quartzites with approximately 10 vol% of dark minerals can be observed. The pure yellowish quartzites are located in the core of the antiforme and get darker (greyish) approaching the upper lying transition zone.

6.3.1. MINERALOGICAL COMPOSITION

Dark minerals are mainly flaky muscovites, iron ores and heavy minerals. In fine cracks and veins, penetrating the quartzites, also biotite exists. Micas also exist as detrital grains in between grain boundaries of quartz. SEM investigation shows a bimodal composition of these micas. The core comprises of muscovite and the rim of biotite. The iron-bearing phases are, besides biotite, hematite, ilmenite and titanomagnetite. Ilmenite shows alteration reactions to Leucoxene, which is a conglomerate of very fine grained rutile, hematite and other ore phases. Voids with idiomorphic cubic crystal shapes are the relicts of titanomagnetite and ilmenite. BSE images show small spherules of hematite on the interfaces of the voids, and a fine powder of opaque minerals (Leucoxene). migrating along cracks in the host rock. Detrital heavy minerals are apatite, rutile and zircon. Fine grained rutile is a secondary formation.

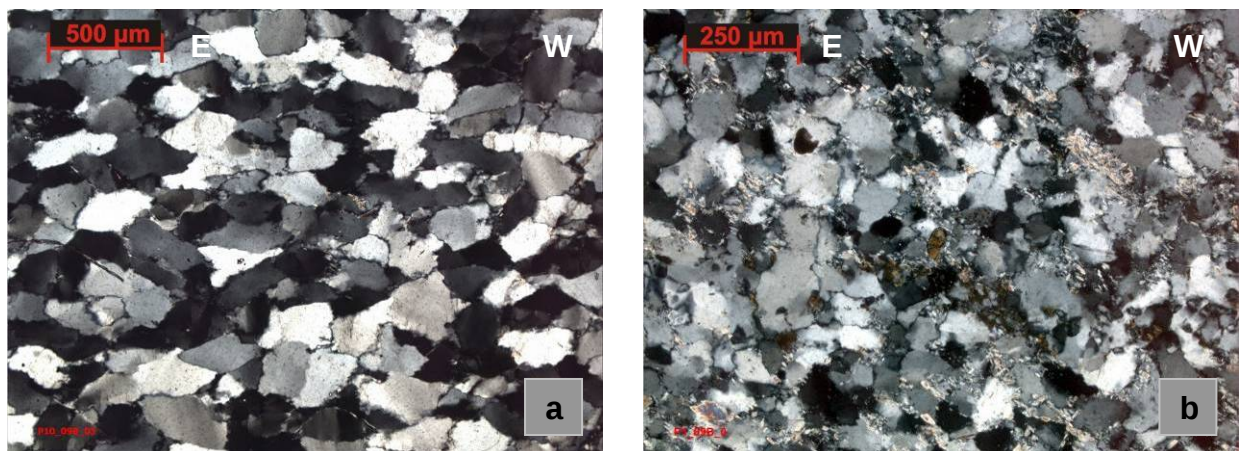


Figure 33: Barrayo Quartzite of the Cabos Series in thin section. A) Pure, clear quartzite with ~99 vol% quartz; XPL. B) Quartzite with considerable amount in micas; XPL.

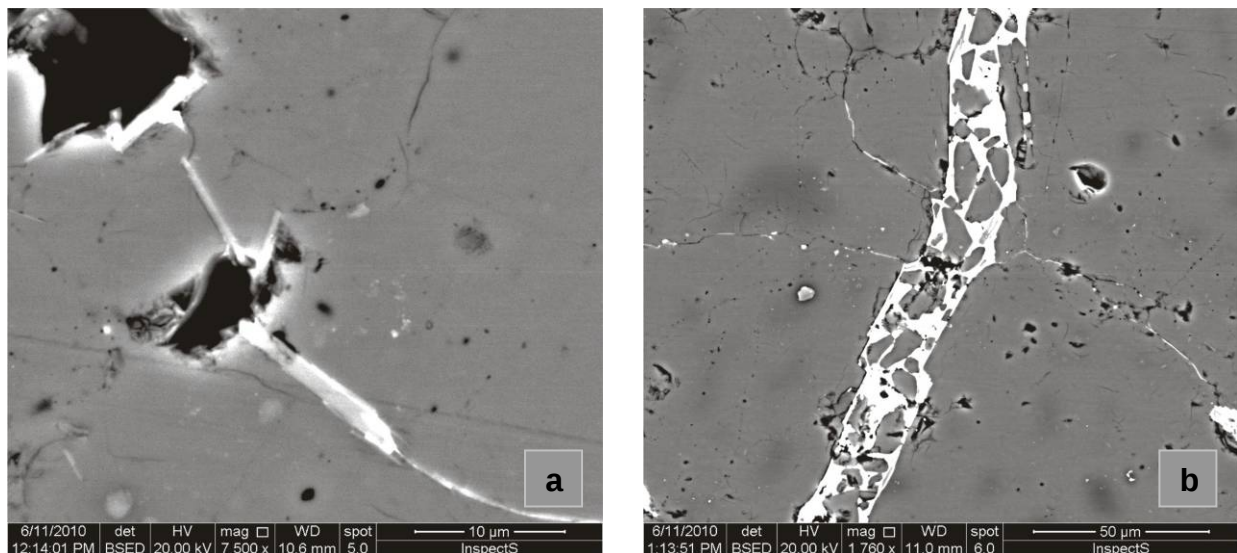


Figure 34: BSE images of micas in the quartzite. A) Interfaces between quartz grains (grey) are made up of voids (black) and micas (white). Brighter colors mean higher content in Fe, so a bimodal composition of the mica in the lower part of the image can be observed. The center is made up of muscovite and the rim of iron-rich biotite. B) Crack in quartzite filled with cataclastic quartz-fragments cemented by biotite.

6.3.2. DEFORMATION

No S1 or S3 cleavage in the quartzites can be observed. However, layering parallel pressure solution surfaces with intervals of several centimetres exist, but they are neither penetrative nor dense enough to use the term cleavage or schistosity. On microscale, the quartz grains show evidence for intracrystalline deformation. In relevant sections, quartz shows preferred orientation of grain boundaries and deformed grains. Quantification for this fabric aspect will be given in an example in the appendix. According to the amount of deformation visible in the components of the quartzites, 3 different types can be distinguished:

1. Low-grade quartzite
2. Medium-grade quartzite
3. Non-metamorphic quartz veins

Low-grade quartzites:

They represent the main constituent of the quartzites. Grain size of the single quartz minerals varies from about 50 µm to 500 µm (depending on their stratigraphical position; see latter paragraph). The detrital grains can be seen in CL-images. The blue colour in CL images proposes quartz grains derived from plutonic rocks and brownish colours mean quartz from metamorphic rocks (Götze et al, 2001). The luminescence colours of quartz often change with increasing time of electron bombardment. Figure 24a shows a blue, plutonic quartz that gradually changes to brown colours, in other words low metamorphic quartz. So, if a quartz grains are of plutonic ori-

gin but suffer low-grade metamorphism, their luminescence colour will change from initially blue to finally dark brown (Richter et al, 2003) as it is the case in these low grade quartzites.

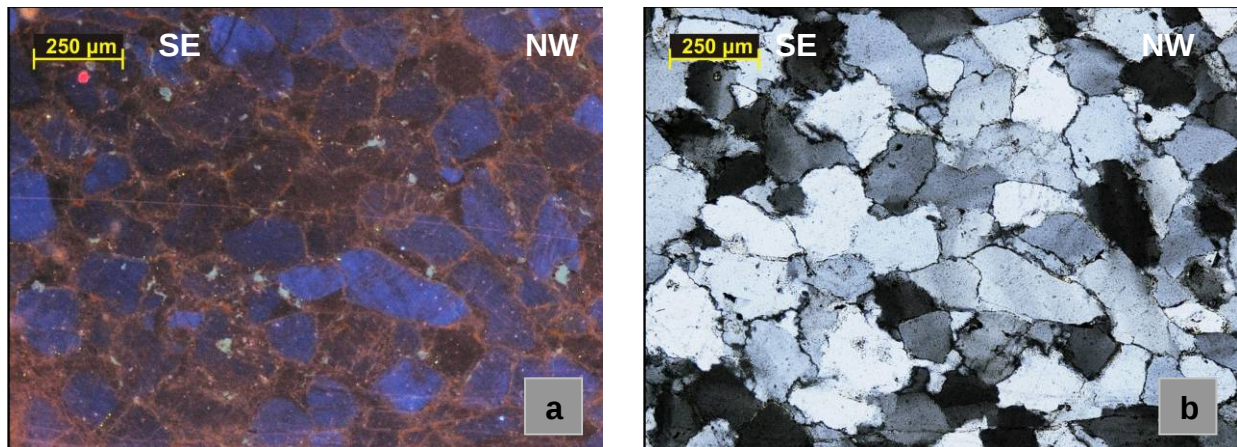


Figure 35: CL (a) and XPL (b) images of the low grade quartzite. The CL picture shows the detrital grains and diagenetic cement. Blue and violet colours indicate quartz from plutonic rocks; brown colours mean quartz from metamorphic rocks. Provenance of the quartz may be some granitic source. Afterwards the rock is affected by low grade regional metamorphism. The CL also reveals inner structures of the grains. The grain in the middle of the image has several internal cracks, which are cemented by diagenetic quartz.

Deformation evidence in the quartz minerals are deformation lamellae, undulose extinction and bulging. Deformation lamellae are very small, linear structures with a high optical relief that penetrate single crystals ($< 1 \mu\text{m}$) and consist of very small fluid inclusions and elongated subgrains (Passchier & Trouw, 1996). They normally appear in low-grade rocks. Undulose extinction is another evidence for intracrystalline plastic deformation. It originates by distribution of dislocations all over the crystal and consequent bending of the crystal lattice. Bulging, or low-temperature grain boundary migration (LTGBM), signifies the penetration of one crystal into the other. This happens because of different dislocation densities. The crystal with lower dislocation density consumes the one with higher dislocation density by migration of the grain boundary. Altogether these mechanisms indicates deformation temperatures less than 300° (Passchier & Trouw, 1996), what also coincides with the suggested lower greenschist facies metamorphism in this area, reported by several authors (Suaréz et al. 1990, Quesada, 1992).

Quartz veins, which have developed in-situ prior to hercynian deformation, are also affected by the identic deformation mechanism like the quartzites.

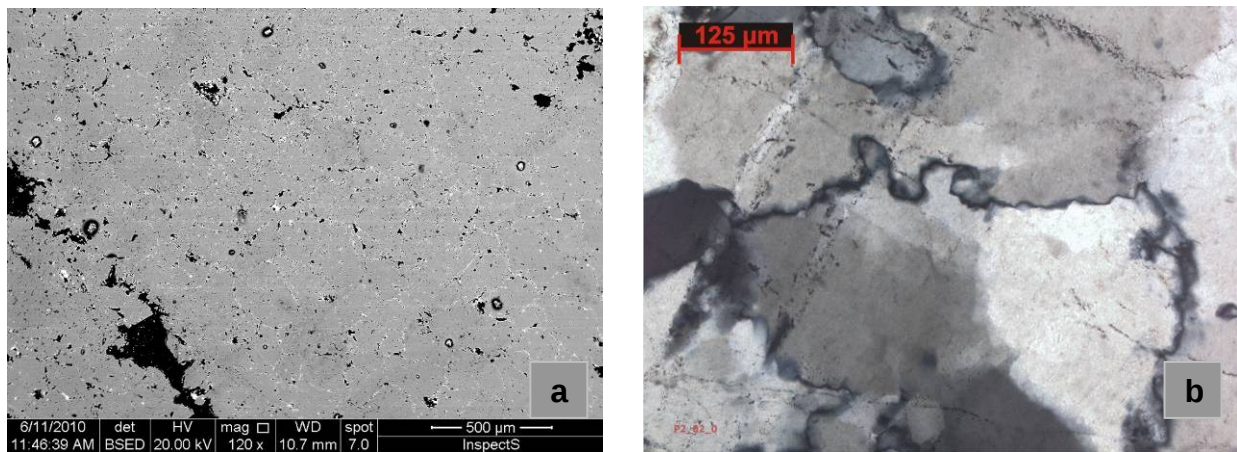


Figure 36: a) BSE Image of the quartzite. No chemical difference can be observed within the quartz. Grain boundaries have slightly brighter colours, meaning higher amount in iron, which is due to the existence of micas on the boundaries. B) Low-temperature grain boundary migration (LTGBM) and sweepy (or patchy) extinction in quartz.

Medium-grade quartzites

As well in the cataclastic zones as in the quartzites itself, components made up of medium grade quartz can be found. These are supposed to be detrital grains of high-grade metamorphic rocks, which were sedimented together with finer material.. Generally a stronger imprint of deformation than in the low grade quartzites is visible, so all deformation evidences are better developed. One additional mechanism of deformation that appears is the subgrain rotation (SR). The evolution of subgrains in a single crystal evokes from a high density of dislocations. Dislocations tend to concentrate on newly developed boundaries, until the lattice of the subgrain is rotated or tilted to an amount, that it is no longer seen as part of the parental grain (Passchier & Trouw, 1996). The accumulation of dislocations is only possible if they are relatively free to move in the lattice (by processes like dislocation climb). This movement gets more likely when temperature rises. Subgrain rotation in quartz happens at temperatures above 400° (Passchier & Trouw, 1996).

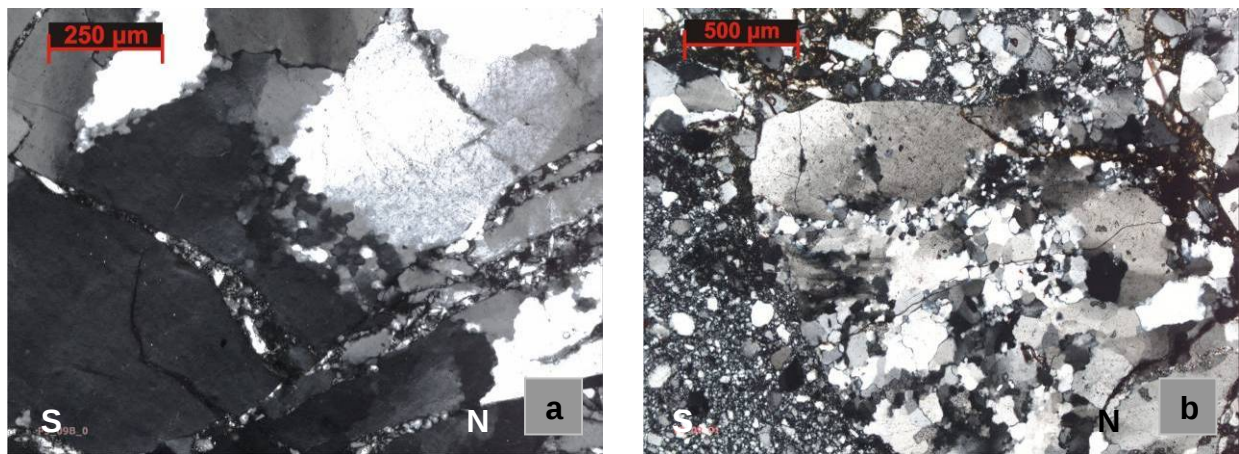


Figure 37: Metamorphic detrital quartite. A) Subgrain rotation recrystallization (SR) in a coarse grained medium-grade quartzite. B) Aggregate of a medium-grade quartzite in a cataclasite.

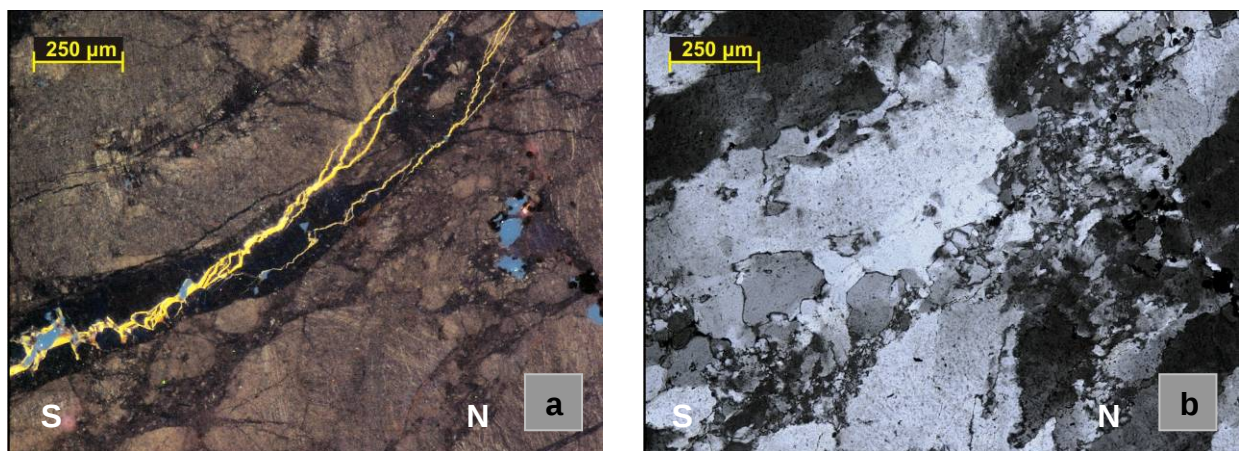


Figure 38: a) CL image of metamorphic quartzite (brown colours). Quartz in the middle (black colors) is authigenic; the yellow crack in the middle (not visible in XPL) is quartz of possible hydrothermal nature, (Götze et al., 2001). In the lower left and upper right, beginning cataclastic deformation can be seen. Light-blue areas are voids. b) XPL of the same section. Detrital grains have some kind of filthy aspect in contrast to diagenetic cement which appears clear.

Undeformed quartz-veins

Sample P16/09 was selected for detailed investigations, because it contains two different kinds of quartz-veins. One of them shows medium-grade metamorphism and second one is none-metamorphic. Besides this, also low-grade quartzite and medium-grade quartzite come into contact. The non-metamorphic veins do not show any evidence for intracrystalline deformation in the optical microscope. They must have developed clearly after the metamorphic events F1 and F3. However, they are affected by the cataclastic alpine deformation.

Through circulating fluids, quartz crystals grow along fracture interfaces in the rock mass. The growth orientation is the same than that of the preexisting grains. The crystal growth happens in several periods. Earlier accretion rims were overgrown and the void was filled progressively. The opening of the vein was in two steps at least. The first generation of quartz growth is authi-

genic, grown during the opening of the crack. Evidence for this comes from CL observations. The luminescence of a mineral is, among other things, dependent of the amount of lattice defects (Götze, 2002). If a mineral grows slowly with all its constituents available at all times, it will crystallize very pure and will incorporate only few lattice defects. Such minerals will not show any luminescence and de facto will be black in CL images. This is also the case for authigenic quartz (Götze et al., 2001). On the other hand, crystals which grow very fast, for example developed from hydrothermal or volcanic media, will not crystallize perfectly and consequently will show this in CL images (Habermann, 2002). Quartz that appears in the center of the undeformed vein in sample P16/09 shows yellow luminescence colours (spectral range ~560-600 nm). Götze et al. (1999) describe this CL emission band with a high content in oxygen vacancies inside the crystal. Often it is found in relation with hydrothermal activities and acidic volcanism. Here it is probably related to low- to medium-tempered fluids that were pumped through the rock mass during seismic slip events. During interseismic phases, fluid pressure rises to a critical values at the interfaces of the seismic faults. During seismic events, newly formed cracks and joints provide a transport media for the fluids and by this way the fluid pressure is reduced. This mechanism is denominated as fault-valve behavior (Sibson, 1977). Not only the formation of quartz veins but also fluidization of cataclasites (see chapter 6.4) may be a consequence of this seismic pumping (Sibson, 1990). The quartz in the center of the vein also shows geometrical zonation. This is attributed to alteration in chemical composition, probably the content in rare earth elements (REE) that also contributes a lot to luminescence phenomena.

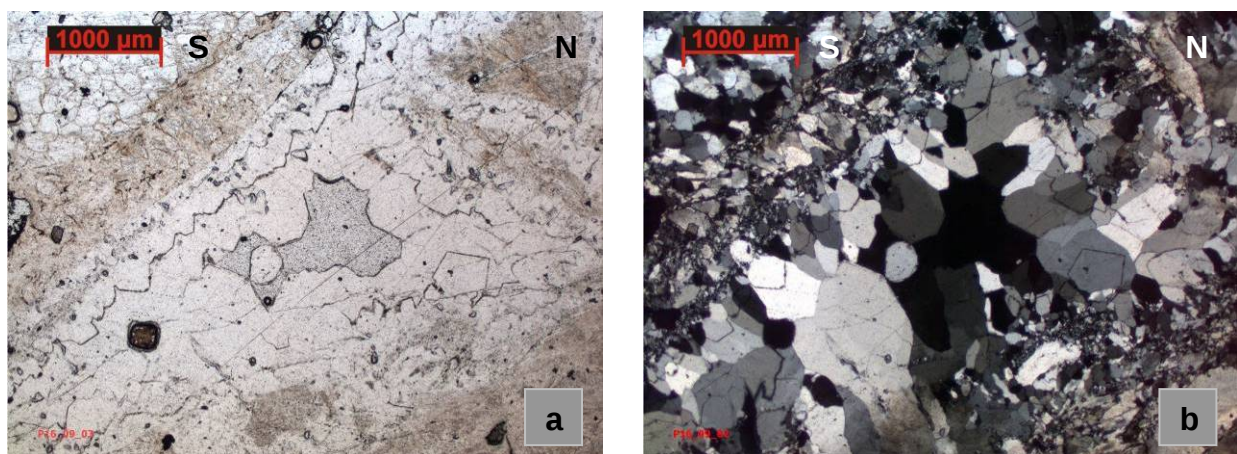


Figure 39: Low-grade quartzite (upper left), medium-grade quartzite (lower right and upper right) and non metamorphic quartz vein (center) united in one image. The PL image in a) shows accretionary rims along former interfaces. The material in the center is hydrothermal quartz. Outside of the rim, the quartz vein is made up of authigenic quartz (evidence from CL images, see fig. 40 a) XPL image (b) shows contact of low-grade and medium- grade quartzite (upper left). Along the vein boundaries, linear cataclastic features appear.

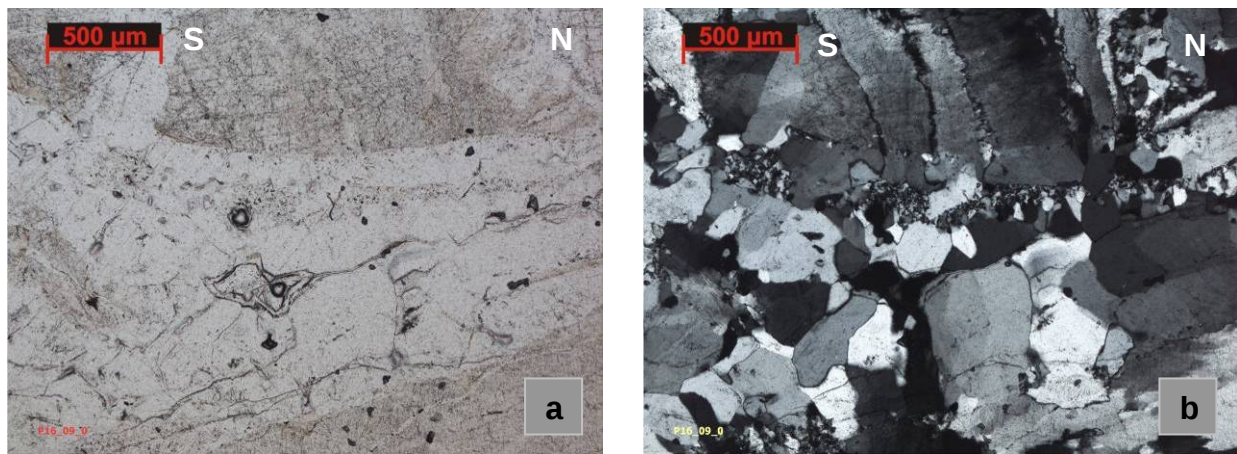


Figure 40: Medium-grade quartz veins (dirty) and non metamorphic veins (clear) in contact. a) PL. b) XPL. New formed quartz uses preexisting crystal growth orientation.

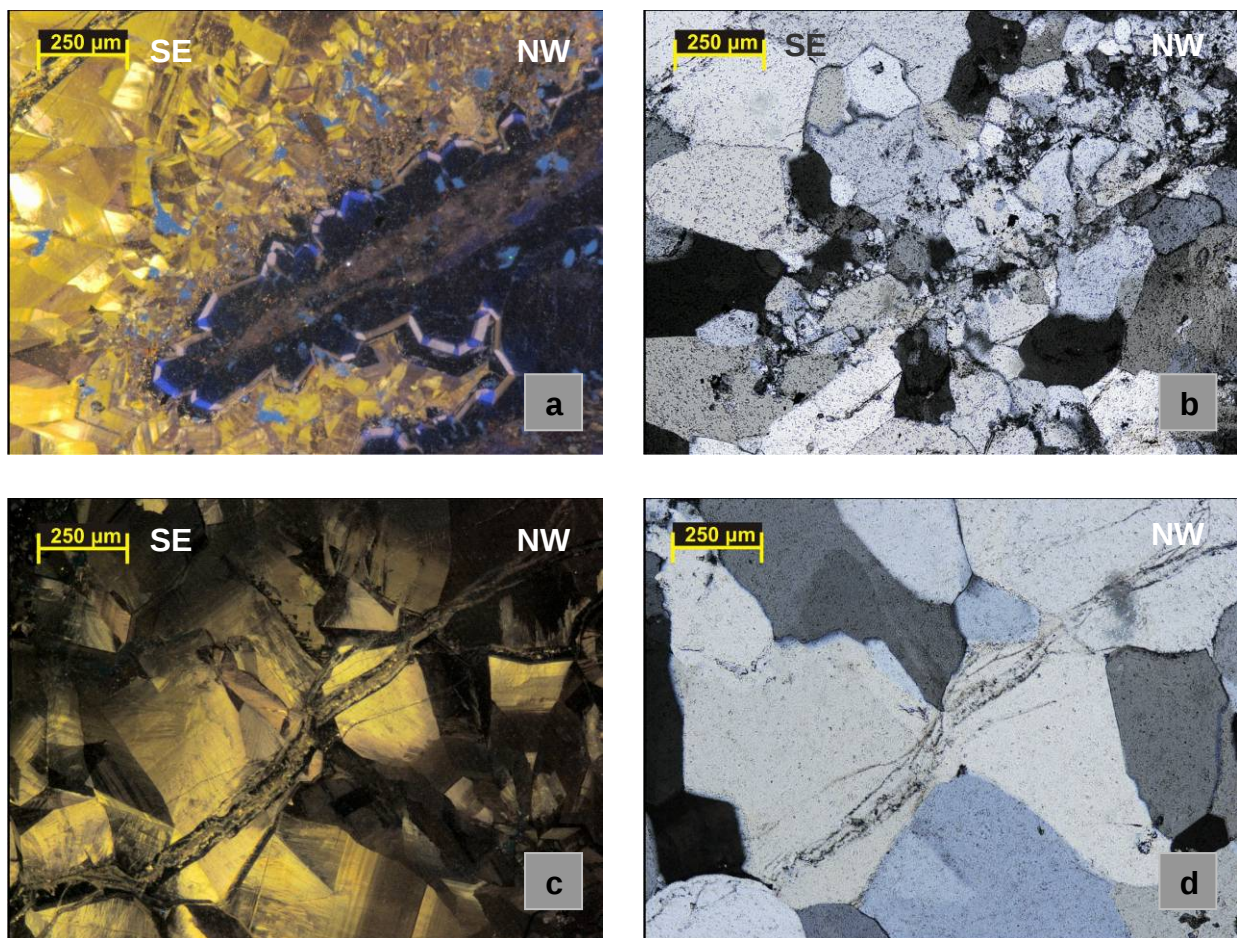


Figure 41: CL images of the non metamorphic quartz veins of sample P16/09 and their XPL counterparts. a) CL image of the fringe of the quartz vein. Brown elongated body in the middle is metamorphic quartzite coated by non luminescent authigenic quartz. In the center of the vein the quartz is highly luminescent, caused by defects in the crystal lattice (see text). b) XPL picture reveals only the different grain sizes of vein and quartzite. c) Center of the vein, showing a fine band of cracks. The quartz crystal exhibits a geometrical zonation, what is attributed to different chemical composition, possibly the content in rare earth elements. d) XPL image of the same section.

6.4. CATACLASTIC ROCKS

The spatial distribution of the cataclastic rocks is described in chapter 5.3. Samples of incohesive cataclastic material were taken from the cemented ridges (of the D- and the Q-structures) and from the core of the faults (Locations see fig. 23 in chapter 5.3). Incohesive cataclasites from the damage zone were not sampled, because it was not possible to extract material from the wall without disaggregation of the rock. According to the amount and type of deformation the rocks have suffered, three distinct end members of cataclasis may be distinguished. The fault gauge of the large faults is made up of very fine comminuted material, denominated as **ultra-cataclasite**. If friction between grains and stress at contact points of grains exceeds their mechanical strength, they will crush and be separated in smaller grains. This process continues until the energy between the grains is not big enough to induce further comminution. Subsequently the grains only frictionally slide besides each other to accommodate further deformation (Sibson, 1997). This state is called cataclastic flow. Further away from the faults the internal deformation energy ceases and clasts may be cracked but the counterparts will not be separated. This type of cataclasite is called **protocataclasite**. Finally cataclastic zones that contain transported fluidized material are denominated as **fluidized cataclasite**. Fluidization is a process, when high pressure fluids infiltrate in the fault zone and mobilize the fault material (Monzawa & Otsuki, 2003). In such cases, the cataclasites behave like a fluid that can migrate through the rock along cracks or joints. Fluidization of fault material is always related to seismic activity (Smith et al., 2008).

As already stated, the upper mentioned types represent end members of possible cataclastic types. In general, the cataclasites represent the transition between these end members.

6.4.1. COMPONENTS OF THE CATACLASITES

The components may be all types of quartzites which have been described in the previous chapter. The range in size is very broad; the lower limit for the recognition of a single component in the optical microscope is about 50 μm but may be much smaller in ultracataclastic areas. Aggregates of quartz minerals are always considered as components of the cataclasite even if they are smaller than 50 μm . Other minerals, like micas or ore phases, are always counted as part of the matrix, because they rarely exceed the 50 μm grain size.

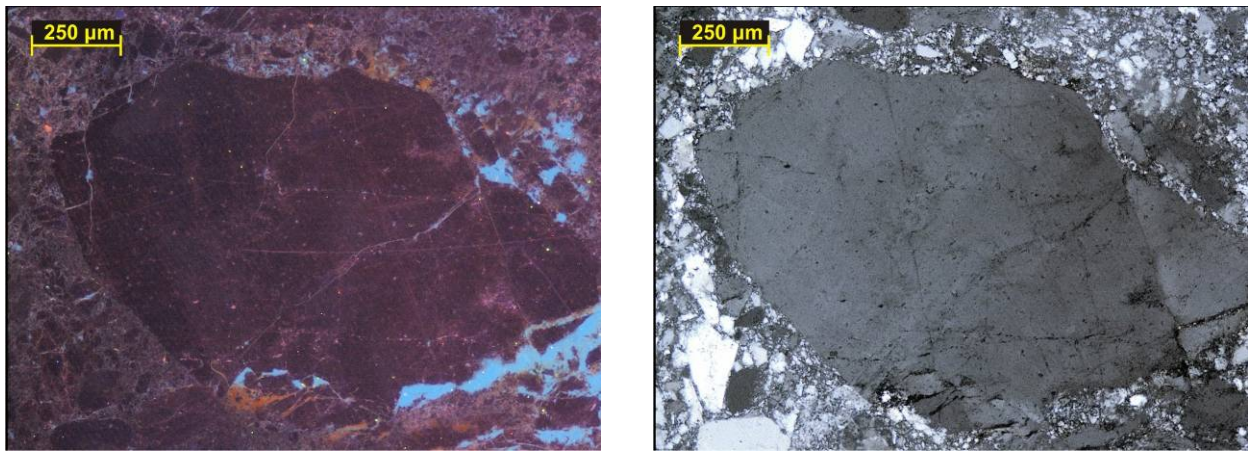


Figure 42: Millimeter-sized component of a medium-grade quartz-vein in a cataclastic zone. A) CL image. B) XPL image. During deformation, small parts are spalling off the edges, consequently diminishing and rounding the component.

6.4.2. MATRIX

The matrix consists approximately 80 % of fine quartz minerals (except the core zones in cataclastic veins). The rest is made up of fine micas, ore phases and heavy minerals which also appear in the quartzites itself. In many cases, the matrix is altered by iron-hydroxide-bearing fluids which leads to alteration of its constituents to a finely disperse distributed mass of goethite and hematite (Limonitization). In plane light the matrix gets a dark brown colour. The dark coloured, altered matrix gives a very good contrast to the incorporating components. This is especially useful for grain size analysis of the cataclasites (see chapter 7).

Fault gauge from the FII fault (sample PFC 01-09) and the ultracataclastic core zone of a cataclastic vein (sample PFC 04-09) were analyzed with X-ray powder diffractometry to reveal their mineralogical composition (see fig. 43a). The fault gauge consists of quartz, muscovite, kaolinite, halite and gypsum. Halite and gypsum may be due to influence of sea water. On the other hand the existence of kaolinite is very interesting. Kaolinite usually develops from feldspars (especially from plagioclase) as a secondary alteration product. In the quartzites, no feldspar is incorporated; consequently the kaolinite has to be transported from the upper lying, feldspar containing slates. The mechanism responsible for material transportation during fault movement is cataclastic flow and/or fluidization. The minerals in the core zone of the cataclastic vein consist of quartz and muscovite and traces of chlorite. One part of the micas must have been transported by fluids or cataclastic flow into this core zone. These zones are therefore supposed to be affected by fluidization.

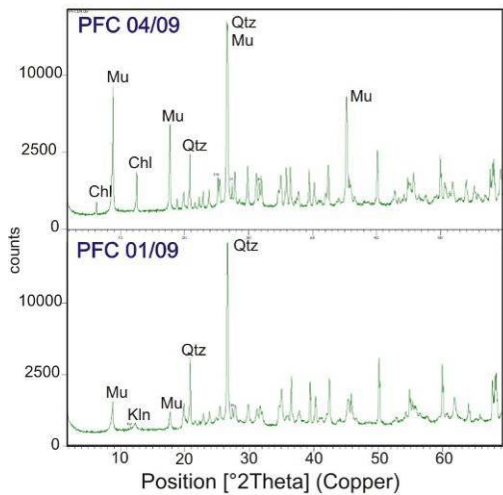


Figure 43: a) X-ray diffractometry results of PFC 01/09 (fault gauge) and PFC 04/09 (core zone). b) Core zone of the cataclastic vein in the outcrop (PFC 04/09). The sample was extracted from the black material in the center.

Additional evidence for fluidization of fault material comes from a flow texture within very fine grained, mica rich matrix. The flaky micas are orientated in the flow direction in the fluidized state. Quartz clasts, with non-spherical shapes, seem to have some sort of shape preferred orientation in the matrix. But this is hard to quantify in thin sections. In chapter 7 possible preferred orientations of quartz components are quantified.

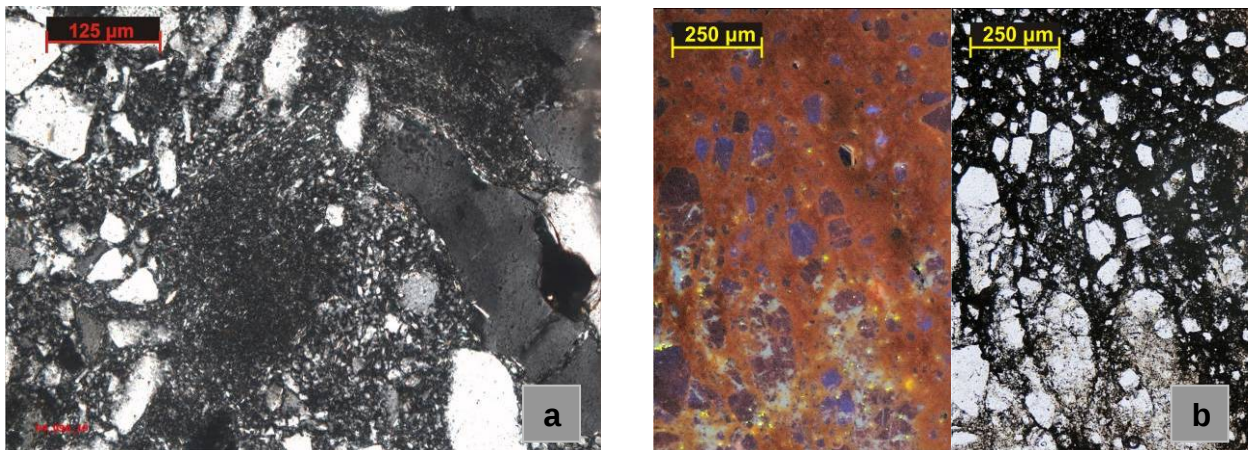


Figure 44: Possible flow regimes of fluidized materials. a) Fine grained, mica rich matrix simulates flow. XPL b) CL (left) and PL image (right) of matrix, altered by Fe-hydroxide fluids. The quartz components show some kind of preferred N-S orientation. Note the sharp contrast of matrix and components.

6.4.3. ULTRACATACLASITE

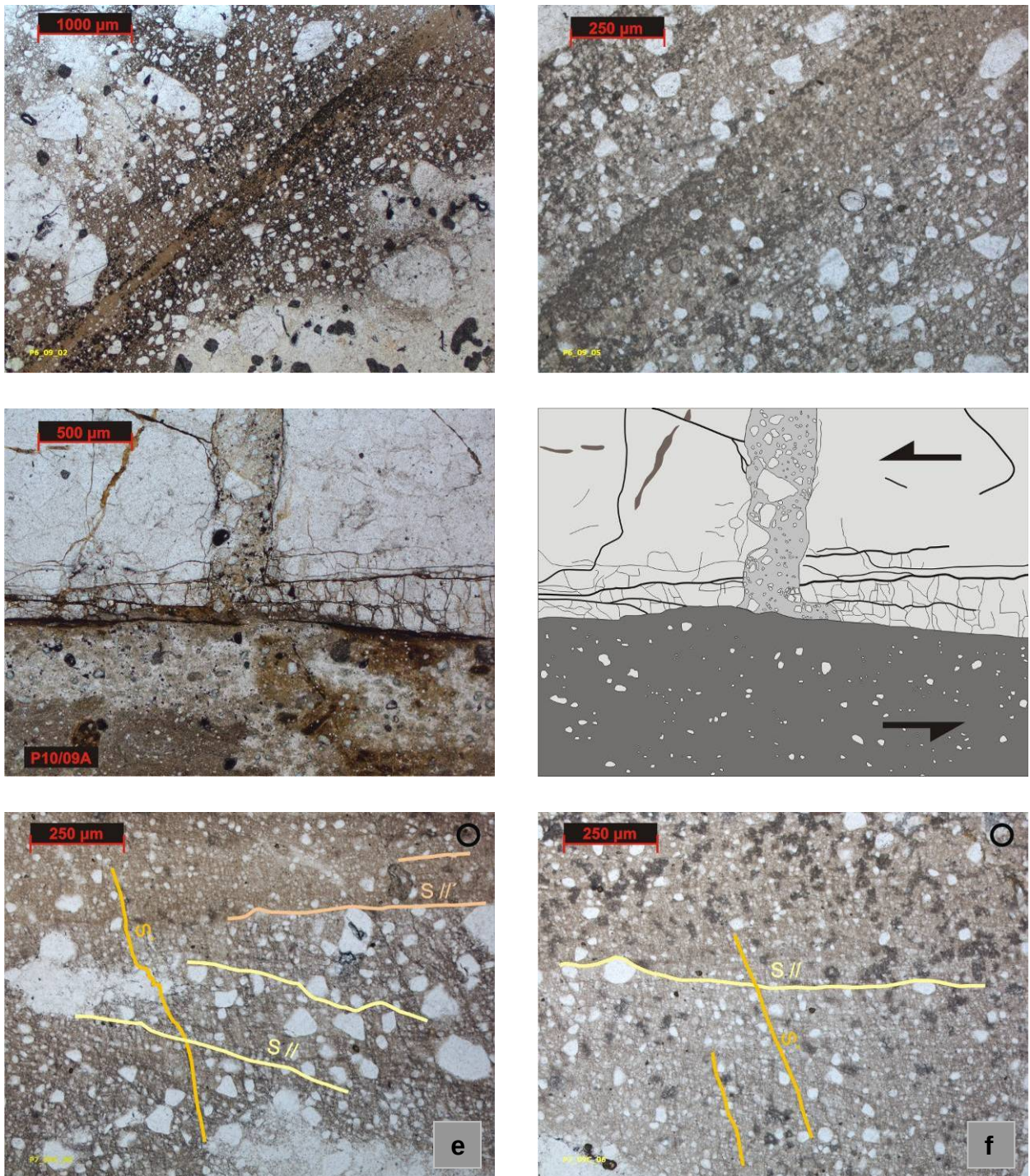


Figure 45: Different aspects of ultracataclastic material. A) core zone. Brown colour is from Fe-bearing fluids, different generations are visible. B) detail of a); Components are subangular to rounded. C) Fault gauge in contact with intact quartzite; In the center a cataclastic injection-dike is visible. Fine material from the fault gauge is injected in the host rock. In the course of this, material is spalled off the wall of the dike; fault gauge material is finer, what means that slip movement continued after the injection. Cracks on the interface between gauge and quartzite mark the damage zone of the ultracataclastic zone. PL. d) interpretation sketch of c). e) + f) Cleavages in the core zones. S // and S //' are parallel to shear zone boundary. Both are frontal sections, so no shear sense is visible. PL.

Ultracataclastic zones are defined to consist of mainly matrix, so they lack components of grain sizes higher than 50 µm. The very fine material is generated as a consequence of strain ac-

commodation in these zones. The fine quartz components in the matrix (they are called components, even if they are beneath the 50 μm grain size) are well rounded and have spherical shapes. A typical ultracataclasite, for example, is the fault gauge, but also inside the cataclastic veins there are areas with very fine cataclasites. In several samples different generations of ultracataclastic material can be distinguished. The different generations are visible by different colours.

One sample (P10/09A) was taken directly from fault FII. There the fault gauge and wall rock are directly in contact (fig. 44 c + d). The section shows a central ultracataclastic zone and a cataclastic dike that intrudes into the quartzite. The material within the dike is of coarser grain sizes and parts of the wall rock are spalled off and incorporated. This material was injected from the fault gauge in an opened crack and in the course of this event also parts of the quartzites were broken off. This is good evidence for fluidization of the ultracataclastic material.

If an ultracataclasite is developed, in the center of cataclastic bands secondary foliation occurs (see fig. 44 e + f). In shear zones, when cataclastic material is generated, a certain normal stress component exists. One foliation, visible in the core zone, is due to pressure solution induced by this normal stress and is parallel to the shear zone boundaries. Another foliation exists which is oblique to the shear zone. The nature of this cleavage also suggests pressure solution mechanism but its origin is not clear. There has to be some shortening regime perpendicular to the shear zone that produces pressure solution and consequently a visible foliation.

6.4.4. PROTOCATACLASITE

Clasts, that are cracked but not completely disaggregated and therefore remain in its usual position are considered protocataclasites. In some cases the cracks are healed by cement. The mechanisms that are responsible for cracking the clasts are referred to mechanical forces between the grains or infiltrating over-pressured fluid. Protocataclasites may be the predecessors of any other cataclastic material, but not necessarily. Another mechanism to produce smaller counterparts of a larger grain is spalling. Then, only small parts on the edges break off and are further comminuted in the shear zone.

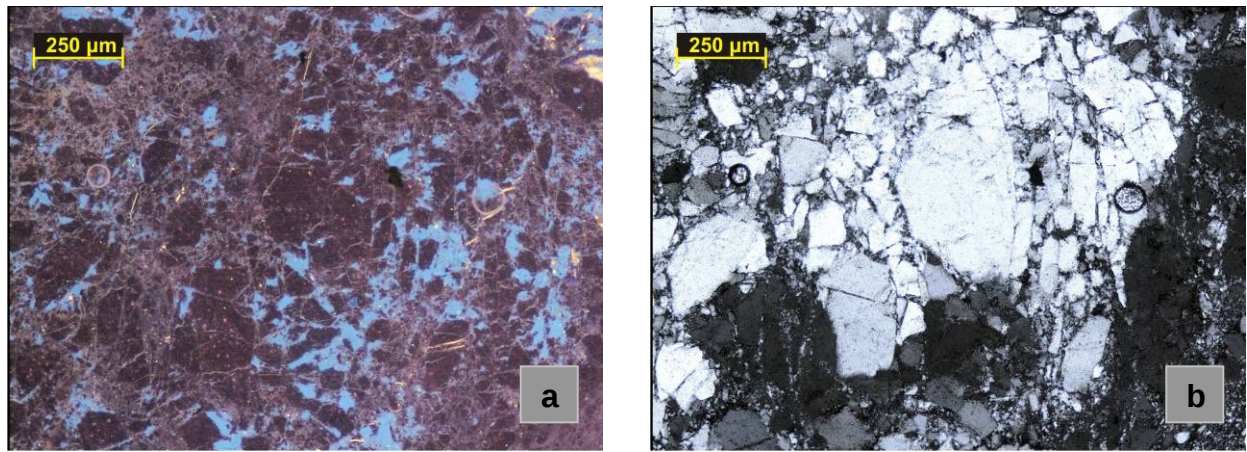


Figure 46: Protocataclasite. A) CL image, blue areas are pore space. B) XPL image.

6.4.5. FLUIDIZED CATACLASITE

One aspect of fluidized materials already was shown in figure 44 c + d. Fact is, that in the core and damage zone of the main strike slip faults cataclastic veins exist, which have completely different orientations than the faults. Precisely they cannot be kinematically related to the fault system. In addition to it, they do not show movement indicators. So the cataclastic material in these veins is not produced in-situ but transported from other locations, which are the fault gauge zones. Several authors have worked on fluidized material and proposed indicators to be able to identify fluidization exactly. Monzawa and Otsuki (2003) state that the comminution of grains in fluidized cataclasites ends with a certain grain size and this would be visible in GSA. Other authors (Smith et al., 2008) establish the preferred orientation of grains as a valuable evidence for fluidization. Grain shape analysis is the best tool to prove this. The following chapter will present whether these statements are correct or not in respect to the available samples.

7. GRAIN SIZE ANALYSIS (GSA) OF FAULT ROCKS

Grain Size Analysis (GSA) is a very useful and modern tool to describe and classify fault rock material. The idea is to measure different parameters of single components of fault rocks and then find the best fitting mathematical model that describes their spatial distribution within the rock mass. First steps to achieve particle size distributions (P.S.D.s) of fragmented geological materials were made with incohesive rocks, by sieving and consequent fractionation by grain size (e.g. Krumbein & Tisdell, 1940). The relation mass vs. grain size was used to describe distributions. Best mathematic descriptors are the Mass-size Power Law Relation and the Weibull Relation (Blenkinsop, 1991). Nowadays GSA is made by computer assisted image analysis of 2D-sections of fault rocks. Advanced image processing software allows automatic detection of grain boundaries and measures different parameter for each grain. The results are then statistically evaluated in spreadsheet programs.

7.1. FRACTAL RELATION D

Investigations have shown that the set of grains in a natural fragmented geological material has a fractal dimension (i.e. a non-integer dimension)(Blenkinsop, 1991). To describe natural fractals one benefits from the scale-independent character of such sets. The dimension is determined by applying $\log(N)$ versus $\log(\epsilon)$ in the so-called Log-Log Plot. In the case of fault material N means the number of grains smaller than the size fraction ϵ . The fractal dimension is then derived by the fractal relation $N(\epsilon) \sim \epsilon^{-D}$. In the range of small ϵ -values the plot gets linear with a mean slope of D . If the linearity is given over several orders of magnitude of ϵ -values the fractal relation is said to be natural. The mathematical relation was improved by several authors, for

example Otsuki (1998) introduced the modified power function
$$P = \frac{N}{N_t} = \left[1 + \left(\frac{D}{D_c} \right)^\alpha \right]^{-\beta/\alpha},$$

where P is the fraction of N grains larger than diameter D of the total amount of grains, N_t . D_c is the minimum diameter of grains where the curve cuts off (resp. loses its linearity). α is a measure for the sharpness of the cutoff and β is the fractal dimension (Montsawa & Otsuki, 2003).

Analysis of grain size distributions from 2D-sections gives evidently different results than 3D measurements would give. Two phenomenons are responsible for this error. Firstly larger particles have a higher probability to be intersected in a thin section and therefore are overrepresented (The sampling effect), secondly a 2D section will show a smaller mean particle size than the real 3D value would be (Truncation effect)(Blenkinsop, 1991). Exner (1972) introduced the term 'tomato-salad problem' for this phenomenon. Mathematical solutions for this problem exist

only for artificial sets of spheroids but not for natural cracked material. However, a major advantage of the fractal description is that the fractal dimension of a 3D distribution is higher by 1 than the 2D distribution (Sammis et al., 1987). So the D_{3D} value is simply derived by adding 1 to the D_{2D} value. Values expressed in this work are always 3D values, a common practice in literature and therefore the values can be directly compared.

7.2. PROCEDURE

Analysis of cataclastic material was made only in suitable sections. Suitable means, areas where components and matrix can be clearly distinguished from each other. This is mostly the case, when the matrix has suffered alteration and iron phases color the matrix dark brown. A sharp contrast is crucial for the automatic detection of the single grains in digital image analysis. Both, digital images from the optical microscope and BSE-images were used. The magnifications have been varied in two samples to study any possible variation within a single set, following the procedure suggested by Monzawa & Otsuki (2003). Working with several magnifications will show effects that are related to the scale of view. The studied samples are:

Sample/Picture	Type	Structure	Magnification	OI	BSE	Manually
P5	Cataclasite	D	12.5/50/100/200	x		
P11	Coarse cataclasite	-	25/50/100/200	x		
P8 A / 04	Protocataclasite	R	25x	x		x
P10 B/ 01	Cataclasite	FG	50x	x		x
P17 A/ 004	Cataclasite / F	Q	600x		x	
P17 A /005	Cataclasite / F	Q	450x		x	
P17 A/ 006	Core Zone/ Ultra-cataclasite	Q	800x		x	x

Some of the images have been digitalized manually, in order to have compare results with the automatically digitalized images and to increase the quality obtained with automatically digitalized pictures. The analysis software used is ImageJ®, a Windows® based freeware application. Grey-scale images are processed with various implemented tools to achieve a good contrast and then transferred to binary images, where each pixel only has 1 or 0 values. Binary images are furthermore edited by removing or adding pixels (erode and dilate process) until the image subjectively reflects the real microstructure.

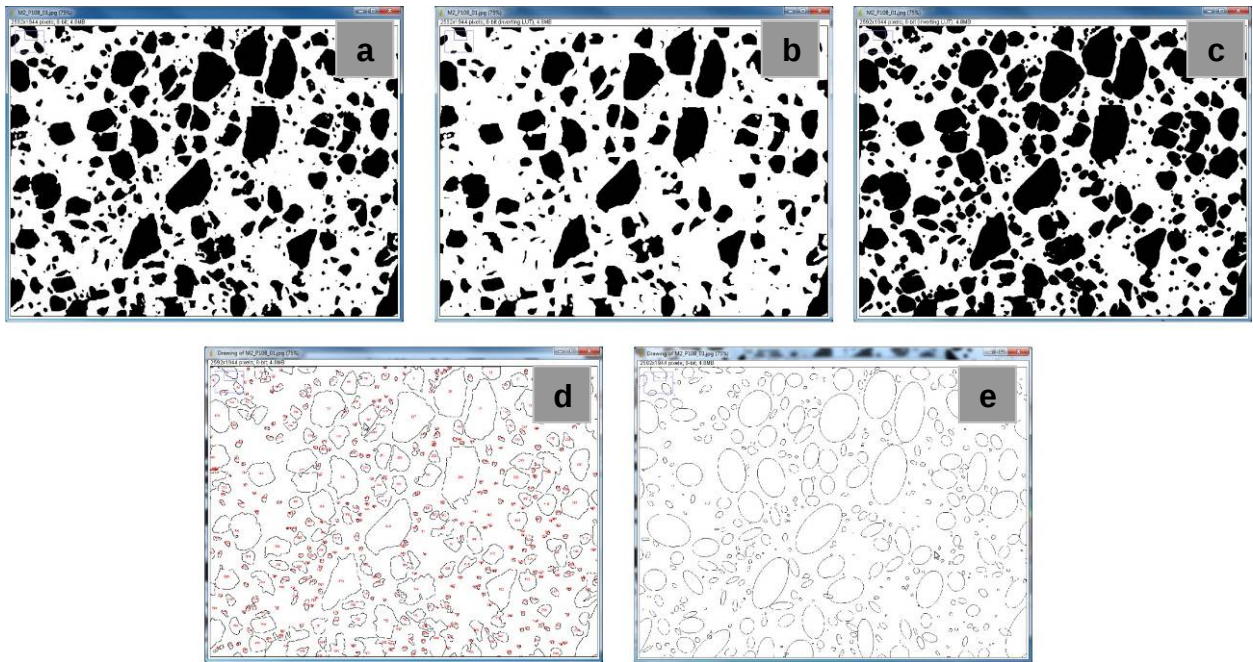
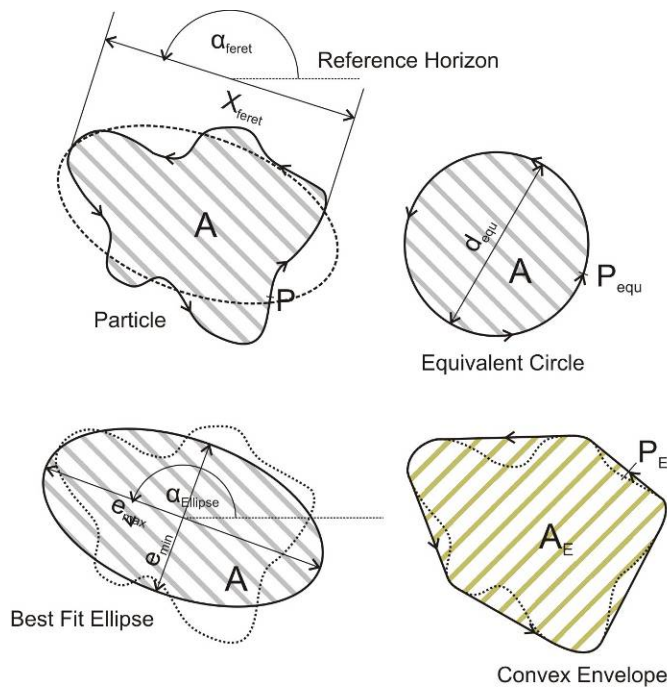


Figure 47: Procedure in GSA with ImageJ®: From top left to lower right: **a)** binary image (automatically digitalized); **b)** after 4x erosion process; **c)** after 4x dilation process; **d)** analyzed outlines; **e)** analyzed best fit ellipses.

After the parameters are set in the software (see next paragraph) the particles are analyzed and the results are exported into a spreadsheet program. There the Log-Log Plot is made and the fractal dimension is calculated. Further parameters quantified in the spreadsheet: Aspect ratio vs. log (N); circularity vs. number of particles and the orientation of the ellipse angle (to reveal a possible shape preferred orientation).

7.3. PARAMETERS

In GSA different parameters can be directly measured in the images and others are mathematically derived from them. Two groups of parameters are distinguished. Size describing and shape describing parameters. An overview of parameters, partly used in this work is given in figure 47. Several new shape parameters and software applications have been introduced recently. For example Heilbronner & Keulen (2006) worked with the Paris Factor and matrix densities to map whole sections and determine protocataclastic and ultracataclastic areas. This work deals with D-values, aspect ratios, circularity and ellipse orientation.



Size Descriptors - Measured Parameters:

Area A

Perimeter P

Feret Diameter x_{feret}

Area of convex Envelope A_E

Perimeter of convex Envelope P_E

Major Axis of Best Fit Ellipse e_{max}

Minor Axis of Best Fit Ellipse e_{min}

derived calculations:

Equivalent Diameter $d_{equ} = 2 \cdot \sqrt{A/\pi}$

Equivalent Perimeter $P_{equ} = d_{equ} \cdot \pi$

Mean Size (Blenkinsop, 1991): $(x_{feret} - feret_{90})/2$

Shape Descriptors:

Aspect Ratio $AR = e_{max}/e_{min}$ (Ellipticity)

Circularity $= 4 \cdot \pi \cdot (A/P^2)$

Roundness $= 4 \cdot A / (\pi \cdot e_{max}^2)$ or $1/AR$

Solidity $= A/A_E$

Paris Factor $= 2 \cdot ((P - P_E)/P_E) \cdot 100$ [%]

$\Delta A = ((A_E - A)/A) \cdot 100$ [%]

(Heilbronner & Keulen, 2006)

Figure 48: Parameters used in GSA

7.4. RESULTS OF THE SINGLE SAMPLES

7.4.1. P5

For sample P5 and P11 an area within the thin section was selected and consequently photographed with different magnifications in the optical microscope. For every single magnification the GSA was accomplished. The grain size interval, taken into account for D-value calculation, had to overlap with the respective neighbouring magnification to a certain amount. Finally the single D-values were normalized and combined in a global D-value. The reason for this procedure is to examine the influence of scale on the calculated parameters. The results will show the reasonableness of the method and, in case, will reveal systematic errors.

Sample P5 was taken from one of the cemented cataclastic crests, namely from a R-structure. Microscopical observations show two distinct generations of cataclastic material. One generation of brown colour and coarse components and the other of very fine, white material that seems to have been injected. The images are from the fine, possibly fluidized material. The components are angular to subangular, but not rounded.

The results show a decrease of D with increasing magnification. This means, small grain sizes are underrepresented in higher magnifications (or in the smaller grain size intervals). An overall weighted D-value was calculated within the fractal range of 10 μm and 1000 μm : **DP5 = 2.848**

The matrix content is 68.57%/70.57%/76.82%/83.07% for 200x/100x/50x/12.5x magnifications. The **mean matrix content is 74.75 %**.

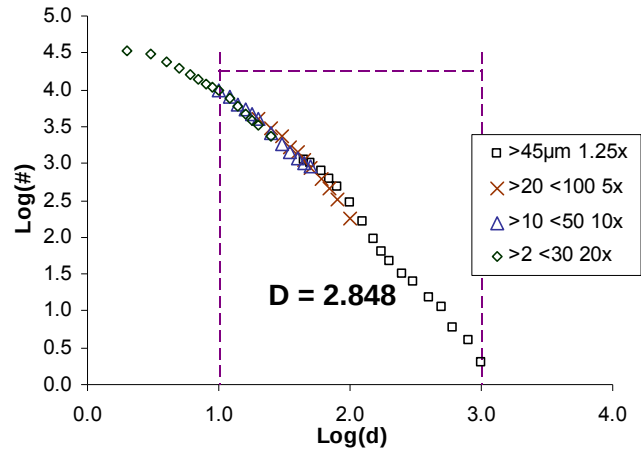
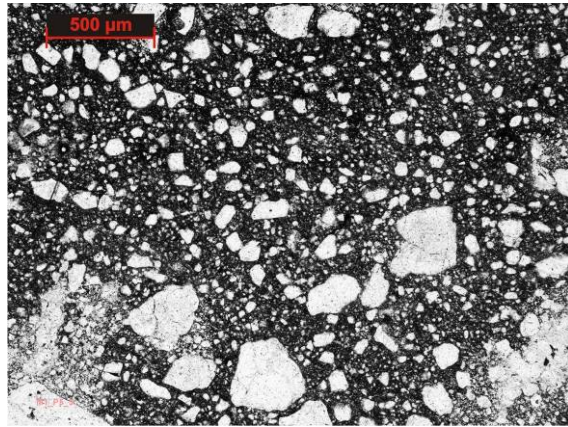


Figure 49: Cataclastic material from sample P5 and Log-Log Plot. D-value was derived by normalization of the single D-values of each magnification. The numbers in the legends are the grain size interval boundaries and the microscope objective magnification. Dashed lines tag the fractal range (2 decades in this case).

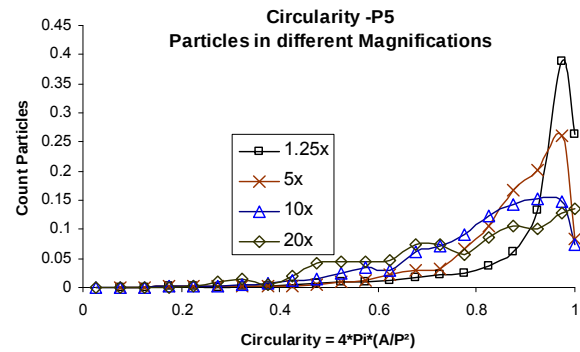
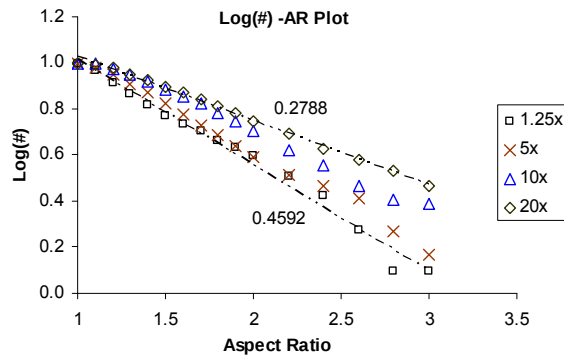


Figure 50: Aspect Ratio vs. log (number of grains); with decreasing grain size the grains get more equal in shape. Circularity is higher in higher grain sizes (explanation see text).

Besides the D-value, the aspect ratio (or also called ellipticity) and the circularity (approximation of the particle to a circle) were calculated. The semi-logarithmic plot of aspect ratio vs. grain numbers also reveals linear behavior. Additionally can be stated that with increasing magnifications (and consequently with decreasing grain size) the shape of the grains get more equal. This is a natural phenomenon, because smaller grains are supposed to be affected by more shear deformation (i.e. cataclastic flow) and therefore more equal in shape. The plot circularity vs. count of particles within an interval shows rounder particles in lower magnifications. This is somewhat strange and is related to systematic errors of the digitalization, because sometimes grains that are in contact at one point are treated as one grain only. Its perimeter is then quite too large in relation to its area, so circularity (which is too high for these grains).

7.4.2. P11

Sample P11 was taken non-orientated from the damage zone. In this area, sampling in the wall is not possible, due to the fragile state of the rocks. A lot of cataclastic veins appear in this area

with orientations parallel to the main faults. In the thin section a wide range of grain sizes appears. None of the components are rounded and they are incorporated in a very fine-grained white matrix. Slip indicators are missing, though in some veins in the outcrop slickensides were observed. The cataclastic material is supposed to have developed from little mechanical comminution of the host rock and to the main part from fluidization.

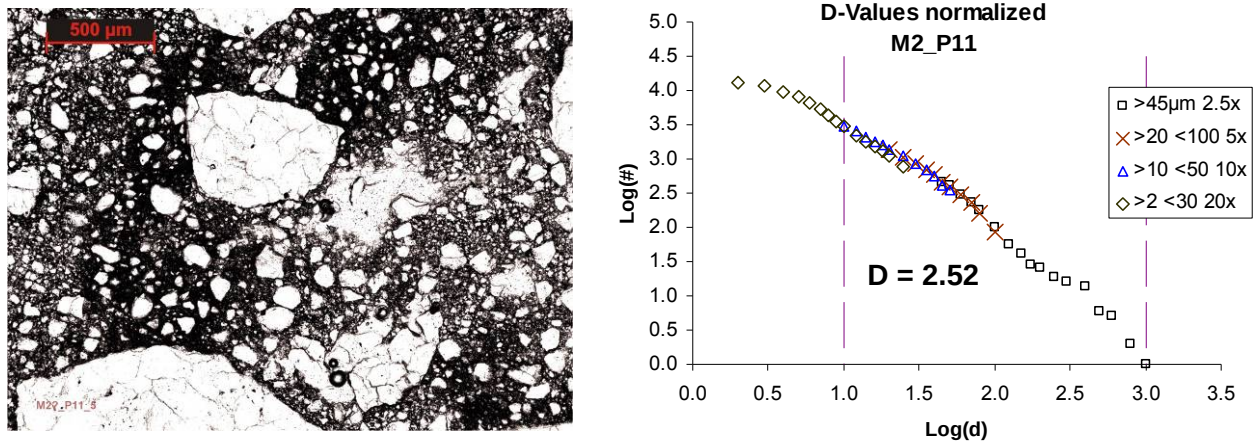


Figure 51: PL-Photograph (50x) and log-log plot of the sample P11. Same fractal range than in Sample P5.

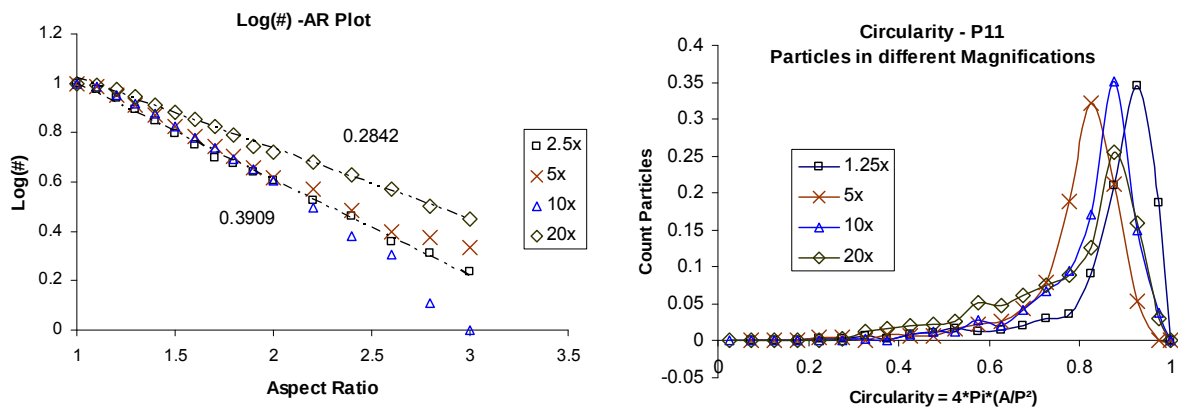


Figure 52 Aspect ratio and circularity plot for sample P11

An overall weighted D-value was calculated within the fractal range of 10 μm and 1000 μm : **DP5 = 2.52**

The matrix content is 57.47%/56.88%/61.78/60.21% for 200x/100x/50x/25x magnifications. The **mean matrix content is 59.01 %**.

The aspect ratio plot also shows variation with different magnifications like in sample P5. However, the circularity is constant in this analyzed sample. The picture quality is somewhat better than in the latter analysis, and the error made by automatic digitalizing is much smaller in this case. The shift in the aspect ratio plots for the different magnifications is smaller in this sample

than in sample P5 but the plot generally has the same aspect. Circularity plots are equal for all magnifications, so there is no longer a systematic error in the measurements.

7.4.3. P8

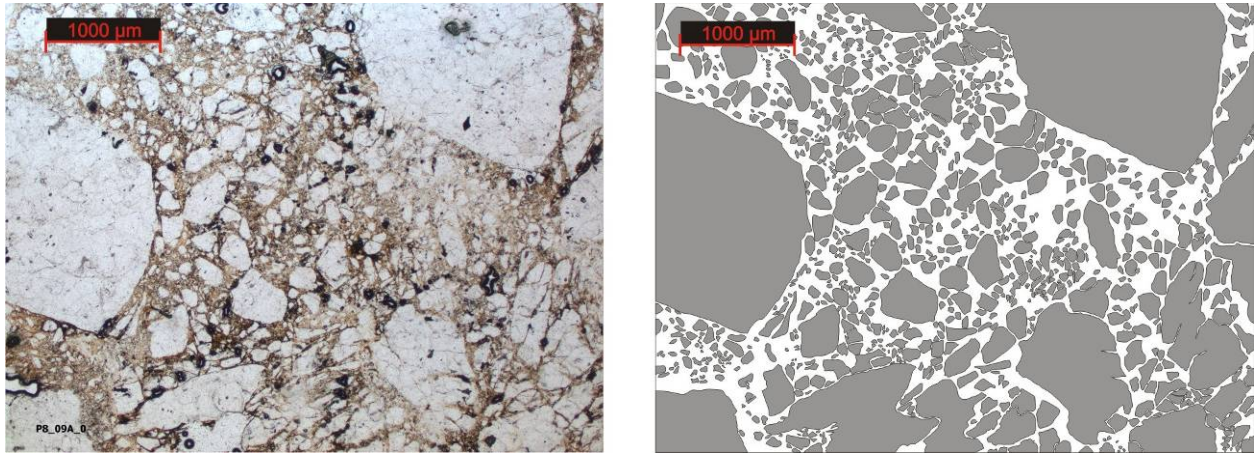


Figure 53: Protocataclasite in Section P8; a) PL image; b) digitalized mask.

A protocataclastic section in sample P8 was digitalized manually from an optical microscope image. The smallest grain sizes considered for digitalization are about 5 μm in its longest axis. The matrix content is 32.78 %. The log-log plot has no good linear correlation. There is a large cut-off in the small grain size area and the curve is non-linear. For this case, the D-value is calculated with the modified power function

$$P = \frac{N}{Nt} = \left[1 + \left(\frac{D}{D_c} \right)^\alpha \right]^{-\beta/\alpha}$$

(Otsuki, 1998). The cut off diameter is set to 180 μm, α is 1.8 and a β-value of 1.53 is derived. The 3D D-value then is **D = 2.53**.

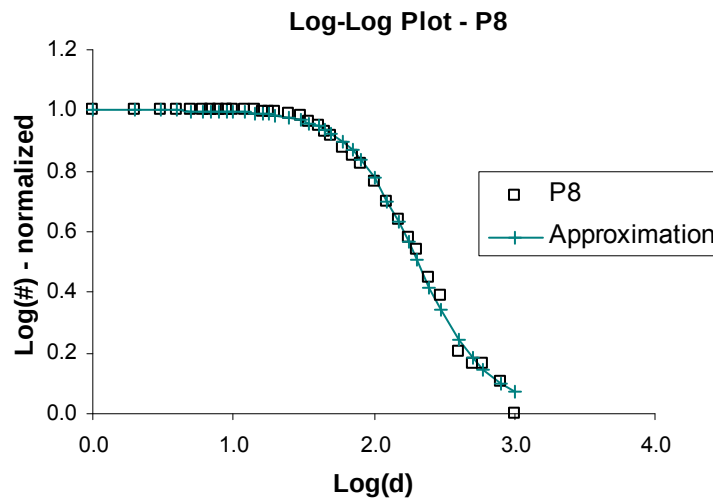


Figure 54: Log-Log Plot of P8 and the modified power function

7.4.4. P10

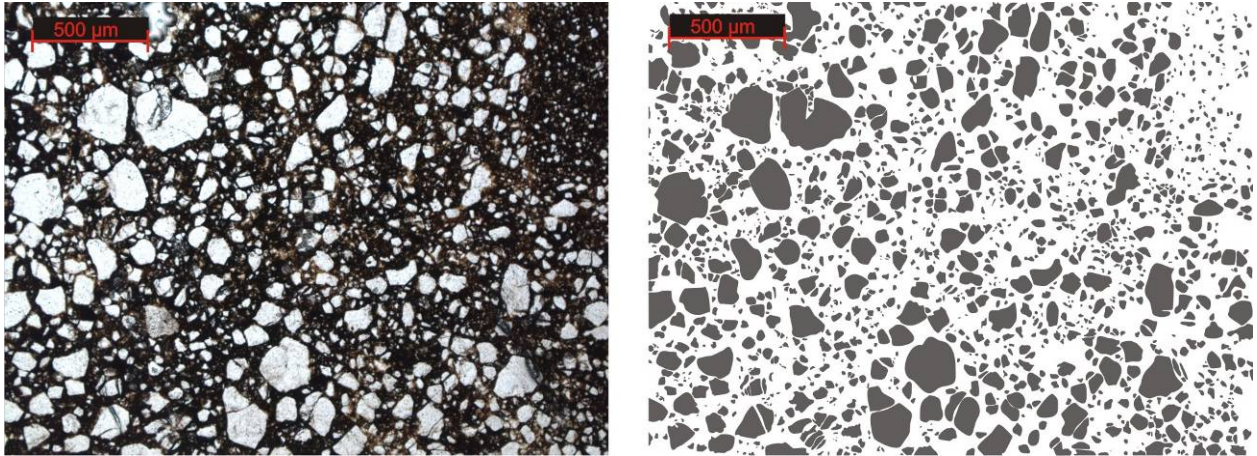


Figure 55: Images of sample P10; Left: photo from optical microscope; Right: manually digitalized.

Sample P10 was automatically and manually digitalized, to see if significant differences occur. The sample is supposed to be a mix of ultracataclastic and protocataclastic material. The matrix content is **60.8 %** for the manually digitalized image and **62.4 %** for the automatic digitalization. Again the log-log plot does not fit a linear distribution over a wide range, so the modified power function is used to calculate the D-value. D_c is set to 75 μm and α is 1.7. Then the D-value (3D) is **2.65**. The grain size distribution of the two images does not differ strongly. So it is convenient to use only automatically digitalized images if possible (i.e. if a good contrast between grains and matrix is given).

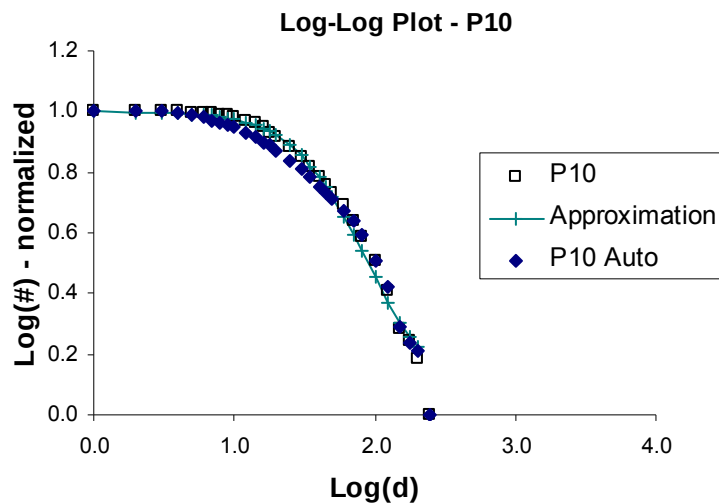


Figure 56 Log-Log Plot for sample P10; GSD for manually and automatically digitalized images are similar.

7.4.5. P17

Three different BSE-images of sample P17 have been analyzed. One of them was manually digitalized. The images are from core zones of a cataclastic vein and comprise of very fine ultracataclastic material. Mean matrix content is **62.5 %**. The power function delivers D-values of about 2.8 but the function does not fit very well the distribution. Here the linear regression gives best values, and the D-value is calculated to **2.92**. The mean grain size is considerable lower than in the other samples, and the D-value is the highest.

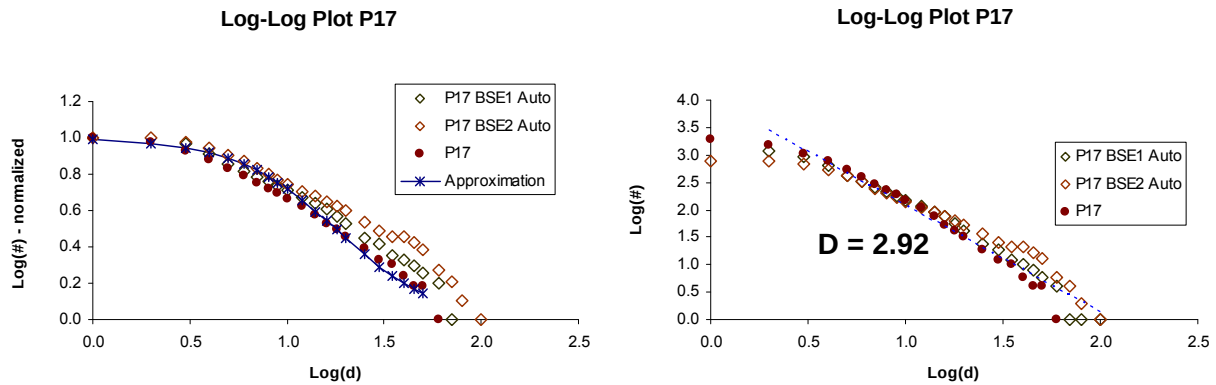


Figure 57: GSD of sample P17; left: approximation with power function; right: linear regression that fits the distribution best.

For this sample, shape preferred orientation measurements of the grains have been carried out. Only grains with an aspect ratio higher than 1.5 were considered. The results show a certain preferred orientation, but statistical calculations show that they are not significant. Therefore, whether fluidization plays a role in the formation of the cataclasites of sample P17 or not can not be proved by shape measurements. The suitability of shape fabric measurements with GSA is demonstrated in an example in the appendix.

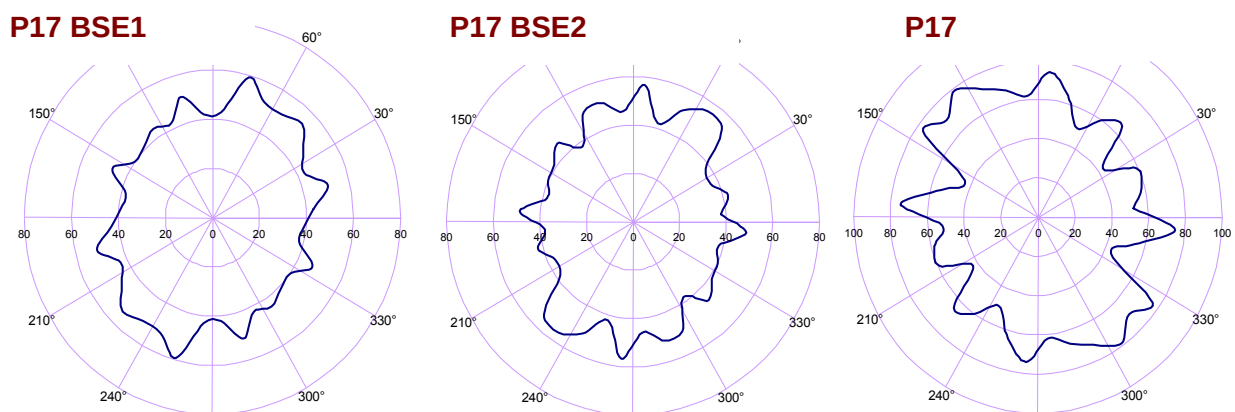


Figure 58: Polar plots for grain orientation of sample P17.

7.5. SUMMARY OF THE RESULTS

Sample	Type	D-value (3D)	Approximation function	Matrix content [%]
P5	Cataclasite	2.85	Linear	74.75
P11	Cataclasite	2.52	Linear	59.01
P8 A	Protocataclasite	2.53	Power	32.78
P10 B	Cataclasite	2.65	Power	61.5
P17 A	Fluidized Cataclasite	2.92	Linear	62.5

P5 and P17 have the lowest mean grain sizes and the highest D-values. The negative correlation of mean grain size and fractal dimension D for natural fault materials was already stated by Blenkinsop (1991). Cataclastic samples, which only have developed by mechanical comminution (P10 B) have D values of about 2.6. Similar values for fault gauge material are also given by other authors (Sammis et al, 1987 or Sammis & Biegel, 1989). In this case, the high D-values in the samples P5 and P17 may be related to further cracking of larger grains by the aid of fluids. The particles transported in fluidized material during a seismic pumping event have high kinetic energy (Monzawa & Otsuki, 2003) and are able to fragmentate larger grains in the case of a collision, thus increasing the fractal dimension of the grain size distribution. Therefore the samples P5 and P17 are regarded to be related to fluidization. Protocataclastic material has slightly lower D-values, starting at 2.5 (P8 A), which is also stated in other publications (Heilbronner & Keulen, 2006). In the end it is clearly stated, that fluids were responsible to a certain state in the development of some cataclastic zones.

8. CONCLUSION

Following conclusion can be made from investigations in the course of this thesis:

- The antiformal bulge, the so-called Portizuelo Antiforme, is a F3-related dome-like structure made up of quartzites of the Cabos Series, sand- and siltstones of the Transition Zone and slates of the Luarca Formation.
- The different rheologies of the units result in a quite distinct evolution of deformation structures. Hercynian deformation is best recorded in the soft layers of the Luarca Slates. There, intense secondary folding and cleavage occurs. The hard quartzitic rocks of the Cabos Series are characterized by brittle deformation.
- Two big faults crop out at the beach. They are left-lateral NE-SW striking transform faults with a certain thrust component. The total offset of the rocks along the fault is about 130 meters. The evolution of the faults is related to alpine deformation.
- Cataclastic veins are abundant in the damage zone and are related to the fault system. They crop out as hardened crests in the abrasional platform. Concerning the spatial distribution, three different sets can be distinguished. They are bound to preexisting planes, like joints, stratigraphic boundaries and quartz veins. Microscopic examinations show the beginning cataclasis at such weak spots.
- Some cataclasites have an intrusive character, and crosscut the rockmass like volcanic dikes. They are supposed to have generated during seismic events and comprise of fluidized fault material. Fluid pressure rises during interseismic phases until a critical value. During the subsequent seismic slip, cracks are opened and the overpressured fluids, consisting of a liquid phase and cataclastic material, infiltrate the rock. The liquid phase cements the incohesive cataclastic rocks.
- Grain Size Analysis (GSA) shows different fractal dimensions for protocataclastic, cataclastic and partly fluidized material. Evidences for fluidization are higher D-values, because larger grains are further crushed by colliding particles. D-values of fault gauge material are in the range of 2.6. D-values for suspected fluidized cataclasites are around 2.9.

REFERENCES

- Andrade, S.A, Santos, J.F., Oliveira, J.T., Munha, J. & Gonçalves, F. (1991). Magmatismo orogénico na transversal Odivelas-Santa Suzana. XI Reunión Geología Oeste Peninsular, Guía de excursión, 47-54.
- Antón, L., Muñoz-Martín, A. & de Vicente G. (2010). Alpine paleostress reconstruction and active faulting in western Iberia - *Cent Eur J Geosci*, **2(2)**: 152-164.
- Aramburu, C., Truyols, J., Arbizu, M., Ménéndez-Bedia, I., Zamarreño, I., García-Ramos, J.C., Suárez de Centi, C. & Valenzuela, C. (1992). El Paleozoico Inferior de la Zona Cantábrica. In: Gutierrez, J.C. et al. (Eds), *Paleozoico Inferior de Ibero-América*, 397-421.
- Aramburu, C., 1995. El Precámbrico y el Paleozoico Inferior. In: Aramburu, C. and Bastida, F. (Eds) *Geología de Asturias*, Ediciones Trea, 35-50.
- Arce, L.M. (1997). *Guía De Los Espacios Naturales De Asturias*, Ediciones Trea S. L., Gijón, 1997.
- Arenas, R., Gil Ibarguchi, J.I., Gonzales Lodeiro, F., Klein, E., Martínez Catalán, J.R., Ortega Girones, E., de Pablo Macia, J.G. and Peinado, M. (1986). Tectonostratigraphic units in the complexes with mafic and related rocks of the NW of the Iberian Massif - *Hercynia*, **2**: 87-110.
- Baldwin, C.T. (1975). The stratigraphy of the Cabos Series in the section between Cadavedo and Luarca, Province of Oviedo, N.W. Spain - *Brev. Geol. Ast.*, **19**: 1-16.
- Barrois, C. (1882). Recherches sur les terrains anciens des Asturies et de Galice - *Mem Soc Géol Nord*, **4**: 292-303.
- Bastida, F. & Pulgar, J.A. (1978). La Estructura del Manto de Mondonedo entre Burela y Tapia de Casariego (Costa Cantábrica, NW de España) - *Trabajos de Geología*, **10**: 75-159.
- Bastida, F. (1980). Las estructuras de la primera fase de deformación herciniana en la zona Asturoccidental-leonesa (Costa Cantábrica, N.W. de España). Thesis doctoral, Univ Oviedo.
- Bastida, F. (1981). La esquistosidad primaria: una síntesis sobre sus características y desarrollo - *Trabajos de Geología*, **11**: 35-54.
- Bastida, F. (1982). La esquistosidad primaria y las microestructuras de las cuarcitas en la Zona Asturoccidental-leonesa - *Trabajos de Geología*, **12**: 159-185.

Bastida, F., Martínez Catalán, J.R. & Pulgar, J.A. (1986). Structural, metamorphic and magmatic history of the Mondoñedo nappe (Hercynian belt, NW Spain) - Journal of structural geology, **8**: 415-430.

Bastida, F. and Aller, J. (1995). Rasgos geológicos generales. In: Aramburu, C. and Bastida, F. (Eds) Geología de Asturias, Ediciones Trea, 27-34.

Blenkinsop, T.G. (1991). Cataclasis and Processes of Particle Size Reduction. – Pure And Applied Geophysics, **136**: 59-86, Basel.

Coney, P., Jones, D.L. & Monger, J.W.H. (1980). Cordilleran Suspect Terranes - Nature, **288**: 329-333.

Crimes, T.P. (1987). Trace fossils and correlation of Late Precambrian and Early Cambrian strata - Geol Mag, **124**: 97-189.

Drot, J., Matte, P. (1967). Sobre la presencia de capas del Devoniano en el límite de Galicia y León (NW de España) - Notas Com Inst Geol Min Esp Madrid, **93**: 87-92.

Exner, H.E. (1972). Analysis of Grain- and Particle-size Distributions in Metallic Minerals – Int. Met. Rev., **159**: 25-42.

Färber, A. & Jaritz, W. (1964). Die Geologie des westasturischen Küstengebietes zwischen San Esteban de Pravia und Ribadeo (NW Spanien) – Geologische Jahrbuch, **81**: 679-738.

Farias, P. & Marquínez J. (1992). El Relieve. In: Aramburu, C. and Bastida, F. (Eds) Geología de Asturias, Ediciones Trea, 163-172.

Fernández Pereiro, J. (1992). El Concejo De Valdés. In: Geografía De Asturias, T. III, Editorial Prensa Asturiana S.A., Oviedo, 1992.

García-Mondéja, J. (1996). Plate reconstruction of the Bay of Biscay - Geology, **24**: 635-638.

Götze, J., Plötze, M., Fuchs, H. & Habermann, D. (1999). Defect structure and luminescence behavior of agate - results of electron paramagnetic resonance (EPR) and cathodoluminescence (CL) studies - Mineral Mag, **63**: 149-163.

Götze, J., Plötze, M. & Habermann, D. (2001). Origin, spectral characteristics and practical applications of the cathodoluminescence (CL) of quartz – a review - Mineralogy and Petrology, **71**: 225-250.

- Götze, J. (2002). Potential of cathodoluminescence (CL) microscopy and spectroscopy for the analysis of minerals and materials - *Anal Bioanal Chem*, **374**: 703-708.
- Habermann, D. (2002). Quantitative cathodoluminescence (CL) spectroscopy of minerals: possibilities and limitations - *Mineral. Petrol.* **76**: 247-259.
- Heilbronner, R. & Keulen, N. (2006). Grain size and grain shape analysis of fault rocks. – *Tectonophysics*, **427**: 199-216, New York.
- Julivert, M. (1971). Decollement tectonics in the hercynian Cordillera of Northwest Spain – *Amer Jour Sci*, **270**: 1-29, New Haven.
- Julivert, M., Fontbote, J.M., Ribeiro, A. & Conde, L. (1972). Mapa tectónico de la Península Ibérica y Baleares – *Inst Geol Min Esp*, Madrid.
- Julivert, M. and Truyols, J. (1983). El Ordovicio en el Macizo Ibérico. In: Rios, J.M., Comba, J.A. (Eds), *Geologia de España*: 192-246.
- Kossmat, F. (1927). Gliederung des varistischen Gebirgsbaus - *Abh. Sächs. Geol. L.-A.*, **1**: 1-39.
- Krumbein, W.C. & Tisdell, F.W. (1940). Size Distributions of Source Rocks of Sediments - *American Journal of Science*, **238**: 296-205, New Haven.
- Lotze, F. (1945). Zur Gliederung der Varisziden der Iberischen Meseta – *Geotekt Forsch*, **6**: 78-92, Berlin.
- Lotze, F. (1957). Zum Alter nordwestspanischer Quartzit-Sandstein-Folgen. *Neues Jahrb Geol Paläontol*, **10**: 464-471, Stuttgart.
- Marcos, A. (1973). Las Series del Paleozoico Inferior y la estructura herciniana del occidente de Asturias (NW de España) - *Trabajos de Geología*, **6**: 113 pp, Oviedo.
- Martinez Catalan, J.R., Perez Estaún, A., Bastida, F., Pulgar, J.A. and Marcos, A. (1990). Structure of the West-Asturian-Leonese Zone. In: Gutierrez, J.C. et al. (Eds), *Paleozoico Inferior de Ibero-América*. 453-461.
- Matte, P. (1968). La estructura de la virgation hercynienne de Galice(Espagne) - *Rev. Géol Alpine*, **44**: 1-128.
- Matte, P. (1991). Accretionary history and crustal evolution of the Variscan belt in Western Europe - *Tectonophysics*, **196**: 309-337.

- Monzawa, N. & Otsuki, K. (2003). Comminution and fluidization of granular fault materials: implications for fault slip behavior. – *Tectonophysics*, **367**: 127–143, New York.
- Otsuki, K. (1998). An empirical evolution law of fractal size frequency of fault population and its similarity law – *Geophysical Research Letters*, **25**: 671-674, Washington.
- Panozzo, R. & Hürlimann, H. (1983). A simple method for the quantitative discrimination of convex and convex-concave lines - *Microscopia Acta*, **87**: 169-176.
- Passchier, C. W. & Trouw, R.A.J. (1996). *Microtectonics*, Springer Berlin, 289pp.
- Pérez Estaun, A. (1973). Datos sobre la sucesión estratigráfica del Precámbrico y la estructura del extremo sur del antiforme de Narcea (NW de España) - *Breviora Geol Asturica*, **17**: 5-16.
- Pérez Estaun, A. and Martínez, F.J. (1978). El Precámbrico del Antiforme del Narcea en el sector de Tineos-Cangas de Narcea (NW de España) - *Trabajos de Geología*, **10**: 367-387.
- Pérez Estaún, A., Marcos, A., Martínez Catalán, J.R., Bastida, F and Pulgar, J.A. (1992). Estratigrafía de la Zona Asturoccidental-leonesa. In: Gutierrez, J.C. et al. (Eds), *Paleozoico Inferior de Ibero-América*, 453-461.
- Peucat, J.J., Bernard-Griffith, J., Gil Ibarguchi, I., Dallmeyer, R.D., Menot, R.P., Cornichet, J. & Iglesias, M. (1990). Geochemical and geochronological cross section of the deep Variscan crust: the Cabo Ortegal high-pressure nappe (NW Spain) - *Tectonophysics*, **177**: 263-292.
- Quesada, C. (1989). Late-Proterozoic terranes within the basement of the Paleozoic Iberian Terrane (extended abstract). *Int. Conf. Tectonothermal Expression of Terrane Accretion within the Appalachian Orogen. IGCP Project 233. Abstr. vol.*, 89-92.
- Quesada, C. (1990). Precambrian Terranes in the Iberian Variscan Foltbelt. In: Strachan, R.A., and Taylor, G.K. (Eds), *Avalonian and Cadomian Geology of the North Atlantic*, 109-133.
- Quesada, C. (1991). Geological constraints on the Paleozoic tectonic evolution of tectonostratigraphic terranes in the Iberian Massif - *Tectonophysics*, **185**: 225-245.
- Quesada, C. (1992). Evolución Tectónica del Macizo Ibérico (Una historia de crecimineto por acreencia sucesiva de terrenos durante el Proterozoico superior y el Paleozoico). In: Gutierrez, J.C. et al. (Eds), *Paleozoico Inferior de Ibero-América*. 173-190.
- Ramsay, J.G., Casey, M., Kligfield, R. (1983). Role of shear in development of the Helvetic fold-thrust belt of Switzerland – *Geology*, **11**: 439-442.

Ramsay, J.G. and Huber M.I. (1987). The techniques of modern structural geology. Volume 2: Folds and fractures. Academic Press. London. 391pp.

Ribeiro, M.L. (1987). Petrogenesis of early Paleozoic peralkaline rhyolites from the Macedo de Cavaleiros region (NE Portugal) - *Geologische Rundschau*, **6**: 147-168.

Ribeiro, A., Munha, J. & Marques, F. (1989). Accreted terranes in northwestern Iberia: Tectono-thermal evolution of the uppermost allochthonous thrust complex in Morais and Bragança massifs (Tras-os-Montes, NE Portugal). In: *Tectonostratigraphic Expression of Terrane Accretion in the Circum-Atlantic Paleozoic orogens*. PICG 233 Conf. (Athens, Ga). p.87.

Richter, D. K., Götte, T., Götze, J. & Neuser, R. D. (2003). Progress in application of cathodoluminescence (CL) in sedimentary petrology - *Mineralogy and Petrology*, **79**: 127–166

Sammis, C.G, King G. & Biegel, R. (1987). The kinematics of Gouge Deformation – *Pure Appl. Geophys.*, **125**: 77-82, Basel.

Sammis, C.G. & Biegel, R.L. (1989). Fractals, Fault-gouge and Friction – *Pure Appl. Geophys.*, **131**: 254-271, Basel.

Schäfer, H.J., Gebauer, D., Nägler, T.F. & von Quadt, A., 1988. U-Pb zircon and Sm-Nd studies of various rock-types of the Ossa-Morena Zone (Southwest Spain). In: *Cinturones orogénicos. II Congreso Geológico de España, symposia vol.* 51-57.

Schulz, G. (1858). Descripción geológica de la provincia de Oviedo, Impr. y Libr. de José Gonzales, 138 pp.

Sduzy, K. (1961). Das Kambrium Spaniens. Teil II. Trilobiten -*Akad Wiss Lit Abh Math-Naturwiss*, **7-8**: 217-408.

Sibson, R.H. (1977). Fault rocks and fault mechanism - *Journal of the Geological Society*, **133**, 191-213.

Sibson, R. H. (1990). Conditions for fault-valve behavior - *Geological Society, London, Special Publications*, **54**, 15 - 28.

Smith, S.A.F., Collettini, C. & Holdsworth, R.E. (2008). Recognizing the seismic cycle along ancient faults: CO₂-induced fluidization of breccias in the footwall of a sealing low-angle normal fault - *Journal of Structural Geology*, **30**: 1034-1046, New York.

Stünitz, H., Keulen, N., Hirose, T. & Heilbronner, R. (2010). Grain size distribution and microstructures of experimentally sheared granitoid gouge at coseismic slip rates – Criteria to distinguish seismic and aseismic faults? – *Journal of Structural Geology*, **32**: 59-69, New York.

Suárez, O., Ruiz, F., Galán, J. & Vargas, I. (1978). Edades Rb/Sr de granitoides del Occidente de Asturias (NW de España) - *Trabajos de Geología*, **10**: 437-442.

Suárez, O., Corretgé, L.G. & Martínez, F.J. (1990). Distribution and Characteristics of the Hercynian Metamorphism. In: Dallmeyer and Martínez (Eds) *Pre Mesozoic Geology of Iberia*, Springer Berlin. 129-133.

Turner, F. (1953). Nature and dynamic interpretation of deformation lamellae in calcite of three marbles - *American Journal of Science*, **251**: 276-298.

APPENDIX

I. SHAPE PREFERRED ORIENTATION OF QUARTZITE

For one section of the quartzites from the Cabos Series a shape analysis was made, concerning the preferred orientation of the grains. This is done the same way like with the cataclastic components of sample P17, by measuring the orientation of the equal-area best fit ellipse of each grain. The orientations are subdivided in intervals and weighted with their frequency. The results are illustrated in a polar plot diagram. Only particles with aspect ratios larger than 1.5 have been considered.



Figure 59: Quartzite used for shape preferred orientations measurement. Left: XPL image; Right: digitalized outlines of the grains.

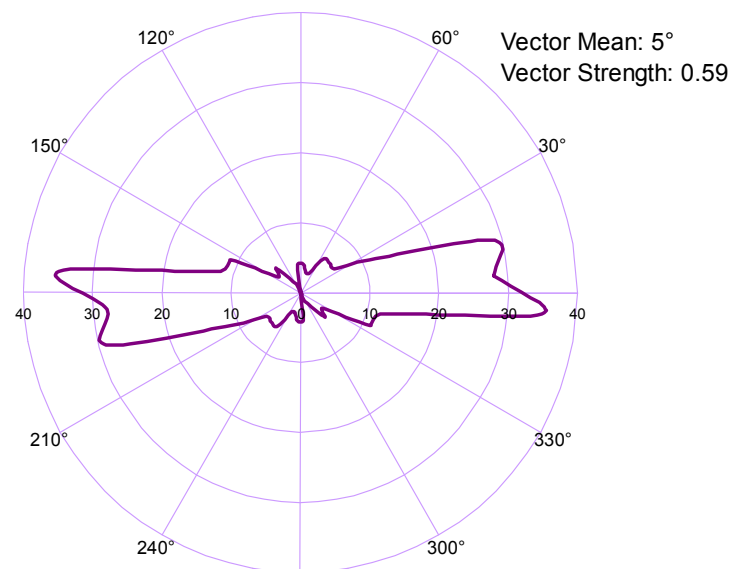


Figure 60: Polar plot of weighted orientations.

The results show a statistically relevant shape fabric, approximately E-W in the image. So the metamorphic events (in this case F1) caused a preferential orientation of the already existing grains. In this case GSA was made to show the suitability of the method to identify any shape fabric of particles in a rock.

II. SPREADSHEET FOR GSA CALCULATIONS

For further use, 4 MicrosoftTM ExcelTM spreadsheets are attached to this thesis in a CD-Rom. In this spreadsheets the GSA calculations have been made. The polar plots have been made with the shareware add-in Polar Plotter v.1.5 ©2004 by Andy Pope, which can be downloaded from: http://www.andypope.info/charts/polarplotter_addin.zip. For the use of the spreadsheet, just copy data from the image analysis software in the data sheets. Make sure you insert the same calculated parameters than those from this thesis. Additionally the processed images used for GSA are attached.

Europass Lebenslauf

Angaben zur Person

Nachname(n) / Vorname(n)	Laner Richard
Adresse	Goldberg 18, 4761 Enzenkirchen (Österreich)
Mobil	068120264202
E-Mail	richi.laner@gmail.com
Staatsangehörigkeit	Österreichisch
Geburtsdatum	8 Aug 79
Geschlecht	Männlich



Gewünschte Beschäftigung / Gewünschtes Berufsfeld

Ingenieurgeologie / Tunnelbau

Berufserfahrung

Zeitraum	01/07/1998 - 01/06/1999
Beruf oder Funktion	Hilfsmonteur
Wichtigste Tätigkeiten und Zuständigkeiten	Montage von Stahlrahmenhallen
Name und Adresse des Arbeitgebers	WOLF System GmbH Osterhofen (Deutschland)
Tätigkeitsbereich oder Branche	Baugewerbe / Bau
Zeitraum	01/12/2000 - 28/02/2001
Beruf oder Funktion	Snowboardlehrer
Name und Adresse des Arbeitgebers	Skischule Schernthaner Kleinarl (Österreich)
Tätigkeitsbereich oder Branche	Tourismus
Zeitraum	01/12/2001 - 28/02/2002
Beruf oder Funktion	Snowboardlehrer
Name und Adresse des Arbeitgebers	Skischule Schernthaner Kleinarl (Österreich)
Tätigkeitsbereich oder Branche	Tourismus
Zeitraum	01/07/2002 - 31/08/2002
Beruf oder Funktion	Bautechniker
Wichtigste Tätigkeiten und Zuständigkeiten	Praktikum in der Bauleitung
Name und Adresse des Arbeitgebers	Universale Bau GmbH Wien (Österreich)
Tätigkeitsbereich oder Branche	Baugewerbe / Bau
Zeitraum	01/10/2003 - 01/08/2004
Beruf oder Funktion	Statiker

Wichtigste Tätigkeiten und Zuständigkeiten	Erstellen von Bauplänen und Statik im Hochbau
Name und Adresse des Arbeitgebers	Statikbüro Gritsch GmbH Zieglergasse, 1070 Wien (Österreich)
Tätigkeitsbereich oder Branche	Baugewerbe / Bau
Zeitraum	01/08/2004 - 30/09/2004
Beruf oder Funktion	Bautechniker
Wichtigste Tätigkeiten und Zuständigkeiten	Praktikum in ZT-Büro für Siedlungswasserbau. Vermessungs- und Planungsarbeiten.
Name und Adresse des Arbeitgebers	DLP Attnang-Puchheim (Österreich)
Tätigkeitsbereich oder Branche	Baugewerbe / Bau
Zeitraum	01/11/2004 - 01/07/2006
Beruf oder Funktion	Hilfselektriker
Wichtigste Tätigkeiten und Zuständigkeiten	Instandhaltungsarbeiten, Veranstaltungstechnik, Revitalisierung der HT-Pläne Geringfügige Beschäftigung
Name und Adresse des Arbeitgebers	Wiener Konzerthaus GmbH Wien (Österreich)
Tätigkeitsbereich oder Branche	Veranstaltungswesen
Zeitraum	01/09/2006 - 01/08/2007
Beruf oder Funktion	Technische Zeichner
Wichtigste Tätigkeiten und Zuständigkeiten	Erstellen von Bauplänen im Tiefbau Freiberuflich
Name und Adresse des Arbeitgebers	Nievelt Ingenieur GmbH Stockerau (Österreich)
Tätigkeitsbereich oder Branche	Erbringung Von Freiberuflichen, Wissenschaftlichen Und Technischen Dienstleistungen
Zeitraum	01/08/2007 - 28/02/2008
Beruf oder Funktion	Bautechniker
Wichtigste Tätigkeiten und Zuständigkeiten	Grundbaustatik, Planung von Baugrubensicherungen
Name und Adresse des Arbeitgebers	Dipl. Ing. Kurt Ströhle ZT-GmbH Wien (Österreich)
Tätigkeitsbereich oder Branche	Baugewerbe / Bau
Zeitraum	01/03/2008 - 31/12/2008
Beruf oder Funktion	Zivildienst
Wichtigste Tätigkeiten und Zuständigkeiten	Betreuung und Pflege von mehrfach behinderten Erwachsenen
Name und Adresse des Arbeitgebers	Wohnverbund 18 Antonigasse, 1080 Wien (Österreich)
Tätigkeitsbereich oder Branche	Gesundheits- Und Sozialwesen

Schul- und Berufsbildung

Zeitraum	01/09/1989 - 30/06/1997
Bezeichnung der erworbenen Qualifikation	Reifeprüfung
Name und Art der Bildungs- oder Ausbildungseinrichtung	BRG Schärding (AHS) Schärding (Österreich)

Zeitraum 01/10/1997 - 01/07/2004
 Bezeichnung der erworbenen Qualifikation Kein Abschluss
 Hauptfächer/berufliche Fähigkeiten Bauingenieurwesen
 Name und Art der Bildungs- oder Ausbildungseinrichtung Technische Universität (Hochschule) Wien (Österreich)

Zeitraum 01/10/2004 - 01/08/2009
 Bezeichnung der erworbenen Qualifikation Bakk. rer. nat.
 Hauptfächer/berufliche Fähigkeiten Geologie
 Name und Art der Bildungs- oder Ausbildungseinrichtung Universität Wien (Hochschule) Wien (Österreich)

Zeitraum 01/08/2009 - 15/07/2010
 Bezeichnung der erworbenen Qualifikation Msc. rer. nat.
 Hauptfächer/berufliche Fähigkeiten Strukturgeologie
 Name und Art der Bildungs- oder Ausbildungseinrichtung Universität Wien (Hochschule) Wien (Österreich)

Persönliche Fähigkeiten und Kompetenzen

Muttersprache(n) **Deutsch**

Sonstige Sprache(n)

Selbstbeurteilung
 Europäische Kompetenzstufe (*)

Englisch

Spanisch / Kastilisch

Französisch

Verstehen				Sprechen				Schreiben	
Hören		Lesen		An Gesprächen teilnehmen		Zusammenhängendes Sprechen			
C1	Kompetente Sprachverwendung	C1	Kompetente Sprachverwendung	C1	Kompetente Sprachverwendung	C1	Kompetente Sprachverwendung	C1	Kompetente Sprachverwendung
B1	Selbstständige Sprachverwendung	B2	Selbstständige Sprachverwendung	B1	Selbstständige Sprachverwendung	B1	Selbstständige Sprachverwendung	B2	Selbstständige Sprachverwendung
A1	Elementare Sprachverwendung	A2	Elementare Sprachverwendung	A1	Elementare Sprachverwendung	A1	Elementare Sprachverwendung	A2	Elementare Sprachverwendung

(*) [Referenzniveau des gemeinsamen europäischen Referenzrahmens für Sprachen](http://europass.cedefop.europa.eu)

Soziale Fähigkeiten und Kompetenzen - Gute Fähigkeit zu multikultureller Zusammenarbeit durch längeren Studienaufenthalt im Ausland
 - Hohes zwischenmenschliches Verantwortungsbewusstsein durch Arbeit mit Menschen mit Behinderung
 - Belastbar im sozialen Umfeld

Organisatorische Fähigkeiten und Kompetenzen - Führungsfähigkeit in Kleingruppen (Projektarbeiten im Studium)
 - Projektleiter in technischem Büro
 - Hohes Maß an Selbstorganisation und Selbstbildung

Technische Fähigkeiten und Kompetenzen - Gut im Umgang mit GPS-Geräten und gute Erfahrung bei Geländekartierungsarbeiten
 - Sprengbefugter

IKT-Kenntnisse und Kompetenzen - souverän im Umgang mit MS-Office
 - sehr gute Kenntnisse in Auto-CAD
 - sehr gute Kenntnisse in Corel-DRAW und Corel-PHOTOGRAPH
 - gute Kenntnisse in Arc-GIS

	- Basiskenntnisse in Go-CAD (3D-Modellierung)
Künstlerische Fähigkeiten und Kompetenzen	- Grafische Gestaltung und Design
Sonstige Fähigkeiten und Kompetenzen	- Alpine Erfahrung (Kartierungsarbeiten im Hochgebirge, Alpine Touren sowie Ausbildung im Zuge der Snowboardlehrausbildung)
Führerschein(e)	C, A, B
Zusätzliche Angaben	Auslandsaufenthalt von 01/02/2009 bis 01/04/2010 in Oviedo, Spanien (Studium und Feldarbeit im Zuge der Master-arbeit)